

ENCYCLOPEDIA *of* COASTAL SCIENCE

2ND EDITION

Edited by
Charles W. Finkl and
Christopher Makowski

 Springer

ENCYCLOPEDIA *of* EARTH SCIENCES SERIES

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ENCYCLOPEDIA OF COASTAL SCIENCE

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Professor Charles W. Finkl has edited and/or contributed to more than eight volumes in the Encyclopedia of Earth Sciences Series. He has been the Executive Director of the Coastal Education and Research Foundation and Editor-in-Chief of the international *Journal of Coastal Research* for the past 35 years. He is also the Series Editor of the Coastal Research Library (Springer). In addition to these duties, he is Distinguished University Professor Emeritus at Florida Atlantic University (FAU) (Boca Raton, Florida). He is a graduate of Oregon State University (Corvallis) and the University of Western Australia (Perth). Work experience includes the International Nickel Company of Australia (Perth), Coastal Planning & Engineering (Boca Raton, Florida), and Technos Geophysical Consulting (Miami, Florida). He has published numerous peer-reviewed technical research papers and edited or coedited and contributed to many books. Dr. Finkl is a Certified Professional Geological Scientist (Arvada, Colorado), a Certified Professional Soil Scientist (Madison, Wisconsin), a Certified Wetland Scientist (Lawrence, Kansas), and a Chartered Marine Scientist (London). Academically, he served as a Demonstrator at the University of Western Australia, Courtesy Professor at Florida International University (Miami), Program Professor and Director of the Institute of Coastal and Marine Studies at Nova Southeastern University (Port Everglades, Florida), and Full Professor at FAU. During his career, he acquired field experience in Australia; the Bahamas; Puerto Rico; Jamaica; Brazil; Papua New Guinea and other SW Pacific islands; southern Africa; Western Europe; and the Pacific Northwest, Midwest, and Southeast USA. Dr. Finkl is a member of several professional societies including the Geological Society of America; Soil Science Society of America; Institute of Marine Engineering, Science and Technology; and the Society of Wetland Specialists. He is a recipient of the International Beach Advocacy Award (Florida Shore and Beach Preservation Association), Certificate of George V. Chilingar Medal of Honor (Russian Academy of Natural Sciences), and Lifetime Commitment to Coastal Science Award (International Coastal Symposium).

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Professor Rhodes W. Fairbridge (deceased) has edited more than 24 Encyclopedias in the Earth Sciences Series. During his career he has worked as a petroleum geologist in the Middle East, been a WWII intelligence officer in the SW Pacific, and led expeditions to the Sahara, Arctic Canada, Arctic Scandinavia, Brazil, and New Guinea. He was Emeritus Professor of Geology at Columbia University and was affiliated with the Goddard Institute for Space Studies.

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Second Edition

edited by

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Dedication

To Rhodes W. Fairbridge, colleague, mentor, and friend who inspired my professional interests in the coast. His persistent nudging, teasing, and invitations to everything coastal drew me away from my first professional interest, soil science, to a new and exciting world of coastal science. His inspirations led me to coastal research, publication, and teaching in the coastal realm, and this interest has passed through me to Chris Makowski, my coeditor.

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Preface to the Second Edition

This two-volume set is an update or upgrade to the original *Encyclopedia of Coastal Science* that was edited by Maurice Schwartz (Western Washington University, Bellingham, Washington, USA). The first edition, a mammoth volume of 1211 pages, was published in 2005 by Springer. At that time, Rhodes W. Fairbridge (Columbia University, New York City, New York, USA) was the Series Editor of the Encyclopedia of Earth Sciences Series. Rhodes was the founding editor of the Series with the first volume (*The Encyclopedia of Oceanography*) appearing in 1966 under the imprimatur of the Van Nostrand Reinhold Company (New York). Interestingly, Maurice Schwartz received his doctoral degree under the tutelage of Dr. Fairbridge while investigating geomorphological, sedimentological, and geospatial relationships of sandy beaches resulting from sea-level rise within the purview of field observations, wave tank experiments, and conceptual frameworks initially worked out and formalized by Per Bruun. Upon concluding his studies, Maurice codified some basic principles in his dissertation into what he called the “Bruun Rule,” which today still evokes appreciation, speculation, and sometimes even contention.

Notwithstanding the term itself, it is perhaps emblematic of broader issues, conceptual interpretations, and divergent points of view in the coastal sciences as the discipline matures into a stand-alone multidisciplinary endeavor that touches aspects of many pure and applied sciences, not the least of which includes the human attributes of coasts that touch in some way the lives of much of the world’s population. Today, about 45% of the world’s population lives within 150 km of the coast and as population density and economic activity in the coastal zone increases, so do pressures on coastal ecosystems. Coastal areas comprise about 20% of the Earth’s surface and by the year 2025, it is estimated that coastal populations will account for 75% of the total world population (World Ocean Network). As such, coastal-marine science deals with an important part of the Earth’s surface and a significant portion of the global human population in a region of

critical importance to the environment and human well-being.

The total length of global coastlines is somewhat of an enigma due to many complicating factors such as tortuosity of the shore itself. That is to say, the rougher coastlines are (in plan view), the more their fractal nature increases, and the more difficult it becomes to determine shoreline length because it varies by scale and resolution. That being said, it is a general consensus that the global shoreline length is in the range or on the order of 1.16 million km (World Factbook) to 1.63 million km (World Resources Institute). Complicating the issue still further, this work does not deal with shorelines or “coastlines” *per se* but with “coasts” in the broadest sense of the term that includes some indeterminate distances alongshore, offshore, and onshore.

The *Encyclopedia of Coastal Science* thus deals with all aspects of “processes, features, and areas” along the coast in what is commonly referred to as the “coastal zone.” This multifaceted region, which functions as the interface between land and water, is studied within the purview of land-, marine-, and estuarine-based sciences and engineering, where concepts, principles, and practices are often seen to interdigitate, overlap, and interlock in various manners to produce a cohesive perspective that commonly falls under the aegis of coastal science. Not to be forgotten is the human side of all these approaches because the coastal zone attracts all manner of endeavor to the extent that modern demographics show dangerously high concentrations of industrial, commercial, military, transportive, and touristic uses. The human component of the coastal zone, which is naturally dynamic, is to various degrees interwoven into aspects of most studies in this region and may focus, for example, on aspects of shoreline retreat (coastal erosion) and remediation efforts, both natural and engineering, or fear or concern over global climate change and attenuated processes such as sea-level rise and the consequences for low-lying coastal areas, whether real or imagined within specified time frames. These high human population densities and associated activities already

threaten about one-third of all coastal regions with environmental degradation from infrastructure development and pollution. The scope of this coastal encyclopedia is thus wide ranging, broadly based, and yet detailed in approach.

This second edition has been thoroughly updated and modified where required. Because a number of topics dealt with subject matters that were rather static (190 topics), such as the coastal ecology, geomorphology, geology, and climatology of particular geographic regions, they were reprinted as per the first edition unless changes (corrections) were required. Many other topics were updated (94 topics) because there were important advances in the subject matter as the knowledge base increased. And since 2005, there have been many scientific and technical developments that warranted inclusion of new topics (43 topics), adding significantly to the corpus of this encyclopedic work.

Preparation of the second edition required a thorough review and evaluation of topics included in the first edition that involved correspondence with previous authors. Their comments and recommendations were invaluable to our efforts as we relied heavily on their expertise in our adjudication. This task was complicated by the fact that some first-edition authors had passed on, others were retired, and some were not able to participate as desired. In those instances the editors themselves reviewed the existing documents, often with the assistance of outside reviewers to ensure adequate coverage. When all of that was said and done, it became clear that many new topics needed to be included in the second edition and again we appreciated the assistance of peer review from associates. We thank the Springer staff for their diligent attention to the details of production because without their help this task would have been monumental on our own. Kudos to them for managing the computerized databases and for keeping us on a timely schedule.

As the current Series Editor of the Encyclopedia of Earth Sciences Series, I am honored to pick up where

Maurice Schwartz left off and offer this second edition of the *Encyclopedia of Coastal Science*. It will be a timely reference in a field of study that is rapidly changing in response to increases in human utilization of the coastal zone and environmental impacts that are related to global climate change and potential sea-level rise. Having completed my doctoral studies at the University of Western Australia (Perth), I came to know Rhodes as this was his *alma mater* and we collaborated on many local research items. When I returned to the United States, I had the opportunity to apprentice under Rhodes Fairbridge by working on the encyclopedia series with him and in that understudy, or apprenticeship, I was able to learn from his knowledge and wisdom. It thus seems fitting that this new edition of the *Encyclopedia of Coastal Science* comes under the co-editorship of myself and Chris Makowski, who has now assumed the apprentice role under me. I am passing on the skills that I once learned long ago, and both Chris and I are proud to make the Encyclopedia of Earth Sciences Series even stronger with this latest edition to the coastal research community.

Last but not least, we wish to again thank the Springer staff for their diligence and assistance with the revision of this work. They worked long hours and tirelessly assisting with correspondence, communication, and data management to ensure a smooth workflow and provide as collegial an environment as possible for a complicated task that involved numerous authors and support staff. The work for all of us was not onerous, but it was at times tedious, time consuming, and somewhat exhausting. The end result justified all efforts to produce the best revision possible. We thus thank, in particular, the following Springer staff for their welcomed support: Petra van Steenbergen, Sylvia Blago, and Johanna Klute.

Charlie Finkl
Chris Makowski

Preface to the First Edition

Map measurements of the world's coastline length have yielded a figure of 500,000 km. However, when all of the very real and intricate coastal crenulations are considered, the actual length is probably closer to 1,000,000 km. Added to this is the fact that 40% of the 6,000,000,000 people presently inhabiting the earth live within 100 km of a coastline. From these observations, it can be seen that coasts are a very major geomorphic and social feature on the face of the planet. And for this reason, scholars in a multitude of disciplines have long been studying the many facets of the zone where the land meets the sea.

In this collected volume, authorities in many fields expound on certain aspects of their expertise, not so much in a dictionary of terms sense as in a series of essays that may be broken down into such categories as: atmosphere and oceanography, ecology, engineering and technology, geomorphology, and human activities related to the coasts (see Appendix 6: Topic Categories). The reader may not completely agree with some of their views; in fact, some of the authors do not agree entirely with each other. Perusing through professional journals in these fields would show the same variety of opinions on a given topic. For that is the nature of science, holding forth on a subject as interpreted by long and careful study of the evidence. What is then to be found here, between the covers of this volume, are 306 entries that contain a wealth of information on different aspects of the world's coast, which we all hold so dear. If there are any questions of omissions or judgment, the fault then lies entirely with me, the editor.

In a similar vein, one would expect the terminology of a science to have universal acceptance; but, sadly, that is not the case. For example, the ubiquitous term "shoreline" can be employed in the historical geomorphic sense of the line formed by the edge of the water against the land as it rises and falls through tidal cycles or atmospheric changes; or as

it has been defined by the US Coast and Geodetic Survey for mapping purposes as the high-tide line, high-water line, or wet-dry boundary. In order for the term to mean the same thing wherever it appears in this volume the geomorphic meaning has been adopted, or clarified where it has deviated from that. For further clarification of this dichotomy, the reader is referred to the entry titled Coasts, Coastlines, Shores, and Shorelines.

Though there have been many trials and tribulations during the four to five years that it has taken to bring this volume to publication, there have also been moments of humor that lightened the load along the way. While explanations for late contributions ran rampant, none was more acceptable than that from the contributor down-under who, while working on a major topic, brought forth two "bubs" (a girl and a boy) to add to an already large family. For sheer inventiveness to a contributor who did not want to repeat a previously published survey, that progressed around a continent in a clockwise fashion, there was your editor's suggestion that he simply proceed in a counter-clockwise direction. In the end it came out only halfway there. Then too, probably the best single line in any entry contained here is the quote to the effect that a certain coastal feature is "... rather like pornography—difficult to define, but you know it when you see it!" That could only be topped by correspondence from another down-under contributor who used wildly colorful expressions that cannot even be repeated here.

Of course, I am most appreciative to all of the very many people who have been involved in this project. However, two individuals stand out most significantly. The first and foremost is the editor-in-chief of this earth-science encyclopedia series, Rhodes Fairbridge, who has been my teacher, mentor, and friend for the past 40 years. The second is Peter Binfield, my editor at Kluwer Academic Publishers, who has guided me through this project with expertise, patience, and humor. To both of these

gentlemen, I offer my most profound gratitude. Thanks are also due to Russ Burmester, Vicki Critchlow, Gene Hoerauf, Larry Palmer, Kevin Short, and Chris Sutton, at Western Washington University, for considerable technical support.

Sadly, media specialist Kevin Short, who worked his computer magic on many graphics in the volume including

the cover photo, passed away suddenly in January of 2004 at the age of 44. Colleagues and friends alike will miss Kevin and remember him for his kindness, humor, and creativity.

Maurice Schwartz

Accretion and Erosion Waves on Beaches

Douglas L. Inman and Scott A. Jenkins

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An accretion/erosion wave is a local irregularity in beach form that moves along the shore in the direction of net littoral drift. The initial irregularity may be caused by a wide variety of events such as the bulge from an ephemeral stream delta, the material from the collapse of a sea cliff, erosion or accretion associated with convergence and divergence of wave energy over an offshore bar, erosion downdrift of a structure such as a groin, sudden loss of sand by slumping at the head of a submarine canyon, or rapid accretion due to beach nourishment as when dredge spoil is placed on a beach. Given the wide variety of causes leading to local beach irregularities, accretion/erosion waves are common transport modes along beaches.

The wave-like form of an accretion/erosion wave is related to the change in sediment transport rate along the beach (divergence of the drift). Specifically, an irregularity in beach topography along an otherwise straight beach produces wave refraction and diffraction that locally modifies the littoral drift system. Wave convergence at an accretionary bulge reduces the littoral drift passing the bulge, causing downcoast erosion. Consequently, the accretionary bulge moves downdrift with an erosional depression preceding it. An initial erosional depression in beach form, as in the lee of a groin, moves downdrift as a traveling sand deficit because the transport potential of the downdrift side of the depression is always greater.

Accretion and erosion waves are best observed from the air or by comparison of beach profiles with time and distance along the beach. The associated change may be several hundred meters in beach width, but more typically is about 10–20 m over a distance of about 1–2 km and may be masked locally by cusps and other small-scale beach features.

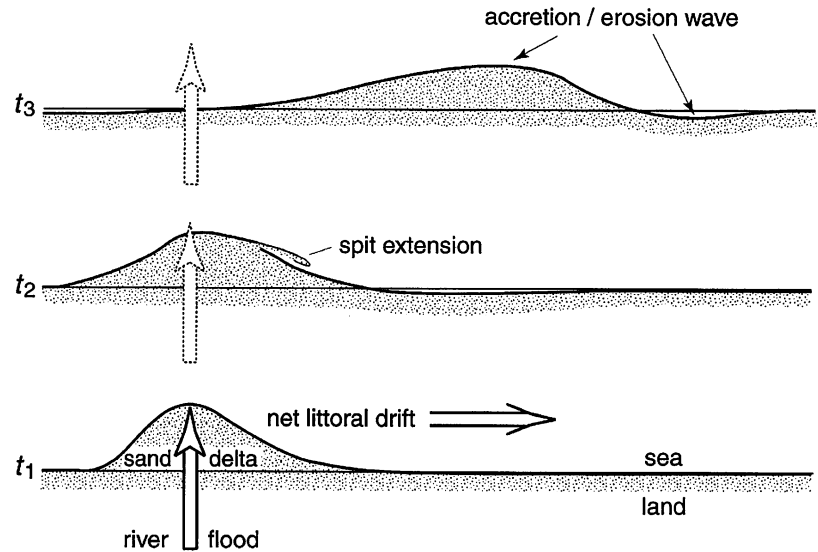
Background

The concept of an accretion/erosion wave was developed to account for the downdrift movement of a sand delta deposited across the beach by an ephemeral stream (Inman and Bagnold 1963). It was observed that the downdrift movement occurred as an accretionary bulge preceded by an erosional depression (Fig. 1). The net littoral drift perturbs deltaic accretion through a series of spit extensions (t_2). Over time, the cumulative spit extensions will progressively displace the accretionary bulge in the downdrift direction, while local wave refraction and refraction-induced divergence of the littoral drift cause erosion downdrift of the bulge (t_3). The areas of accretion and erosion migrate downdrift together in a phase-locked arrangement referred to as an accretion/erosion wave. The movement of the accretion/erosion form was quantified by surveys of the flood delta of the San Lorenzo River in central California (Hicks and Inman 1987), and further evaluated from the sudden release of sand at the San Onofre power plant in southern California (Inman 1987).

The propagation rate of the accretion/erosion wave form is slow initially because the on-offshore dimensions and volume of sand to be moved are at a maximum, and a significant fraction of that volume remains outside the region of rapid transport by waves (Fig. 1; t_1). Once the material enters the surf zone, the longshore transport rates are much higher and the entire accretion/erosion form moves faster, spreads out along the beach, and decreases in cross-shore amplitude. Measurements near the sand release at San Onofre, CA, showed that the form of the accretional wave initially traveled with a speed of about 0.6–1.1 km/year in the 1.8 km near the

Douglas L. Inman: deceased.

Accretion and Erosion Waves on Beaches, Fig. 1 Formation of accretion/erosion wave downdrift from an episodically formed sand delta at time, where $t < t_2 \ll t_3$. (Modified from Inman and Bagnold 1963)



Accretion and Erosion Waves on Beaches, Table 1 Propagation speeds of accretion/erosion waves

Location	Type/cause	References	Net downdrift transport rate	Speed of accretion/erosion wave (km year ⁻¹)	
			10 ³ m ³ year ⁻¹	Nearfield ^a	Farfield
Santa Cruz, CA	Accretion from San Lorenzo River Delta	Hicks and Inman (1987)	268	0.5–1.5	
Santa Cruz, CA	Erosion and bypass accretion from Santa Cruz Harbor	Hicks and Inman (1987)	268		2.2–2.8
Santa Barbara, CA	Erosion from harbor and bypass accretion	Inman (1987)	214		2.5–2.8
San Onofre, CA	Accretion from sand release	Inman (1987)	200	0.6–1.1	
Oceanside, CA	Erosion from harbor	Inman and Jenkins (1985)	200		2.2–4.0
Assateague Island, MD	Erosion from jetties at Ocean City Inlet	Leatherman et al. (1987)	153	~0.3	
Outer Banks, NC	Migration of Oregon Inlet	Inman and Dolan (1989)	590	0.023	
Nile Delta, Egypt	Accretion wave from onshore migration of sand blanket	Inman et al. (1992)	1000		0.5–1.0

^aNearfield is within 1–2 lengths of the perturbing feature such as a sand delta

release point (nearfield) and much faster farther from the release (farfield). The delta from the Santa Cruz River floods of 1982/83 was large (800,000 m³) and extended offshore to depths of over 10 m, and that material moved about 0.5–1.5 km/year during the first year (Hicks and Inman 1987). Subsequently, the downdrift erosion and accretion waves from the delta moved with speeds of 2.2–2.8 km/year (Table 1).

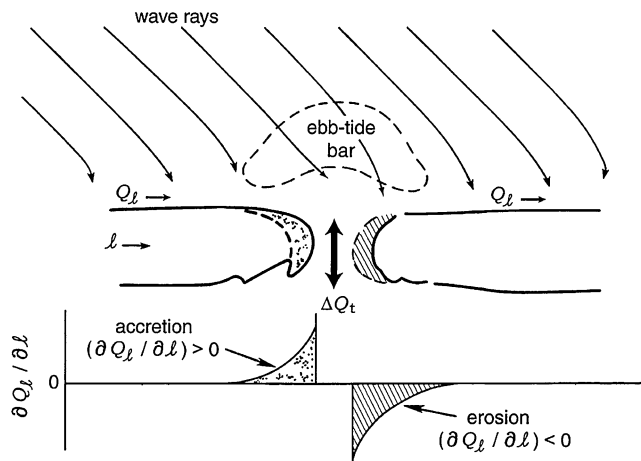
It has been observed that any structure that interrupts the littoral drift of sand along a beach results in an erosional chain reaction traveling downdrift from the structure (Inman and Brush 1973). The propagation rates of the downdrift erosion wave was evaluated from beach surveys following the construction of the harbor jetties at Santa Barbara, CA (Inman 1987), and the enlargement of the harbor at Oceanside,

CA (Inman and Jenkins 1985). Once in the farfield of the structure, the erosion wave, followed by the accretion wave moved downdrift at 2.5–2.8 km/year at Santa Barbara and 2.2–4.0 km/year at Oceanside (Table 1).

Accretion/erosion waves also occur along beaches and barriers downdrift of tidal inlets (e.g., Inman and Dolan 1989). The erosion wave from the jetties at Ocean City, MD, is a well-known example. The inlet between Fenwick and Assateague barrier islands was stabilized by jetties in 1935. The jetties trapped the littoral drift and caused an erosion wave to travel downdrift along Assateague Island, resulting in a landward recession of the entire barrier island of 460 m in 20 years (Shepard and Wanless 1971).

The Nile Delta experiences accretion/erosion waves driven by the currents of the east Mediterranean gyre that sweep

across the shallow shelf with speeds up to 1 m/s. Divergence of the current downdrift of the Rosetta and Burullus promontories entrains blankets of sand that episodically impinge on



Accretion and Erosion Waves on Beaches, Fig. 2 Schematic diagram of the divergence of drift ($\partial Q_l / \partial l$) at a migrating tidal inlet with net tidal flux of sediment, Q_t . (Modified from Inman and Dolan 1989)

the beach. These sand blankets cause shoreline irregularities with average amplitudes of 100 m and wavelengths of about 8 km that travel along the shore at rates of 0.5–1 km/year as accretion/erosion waves (Inman et al. 1992) (see entry on *Littoral Cells*).

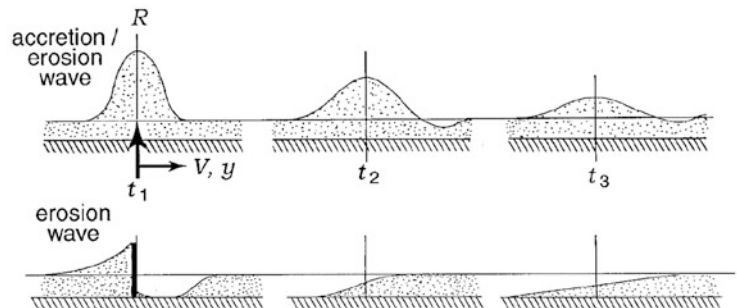
A related example of a traveling accretion/erosion feature occurs when the littoral drift impinges on an inlet causing it to migrate downdrift (Fig. 2). The migration proceeds as an accretion of the updrift bank in response to positive fluxes of sediment delivered by the net littoral drift Q_ℓ , while the downdrift bank of the inlet erodes due to a negative divergence of drift across the inlet, $\partial Q_\ell / \partial \ell < 0$. The negative divergence of the drift across the inlet is caused by wave refraction over the ebb-tide bar and by a loss of a portion of the drift to flood-tide entrainment at the inlet, ΔQ_i . Also the offshore tidal bar, maintained by the ebb-tide flow, moves downdrift with the inlet migration. Although the migration rates of the up- and downdrift banks of the inlet and the tidal bar are phase-locked, they are out of phase with the local net sediment changes in the shorezone bordering the inlet. The inlet banks and channel form an accretion/erosion sequence

Accretion and Erosion Waves on Beaches, Fig. 3

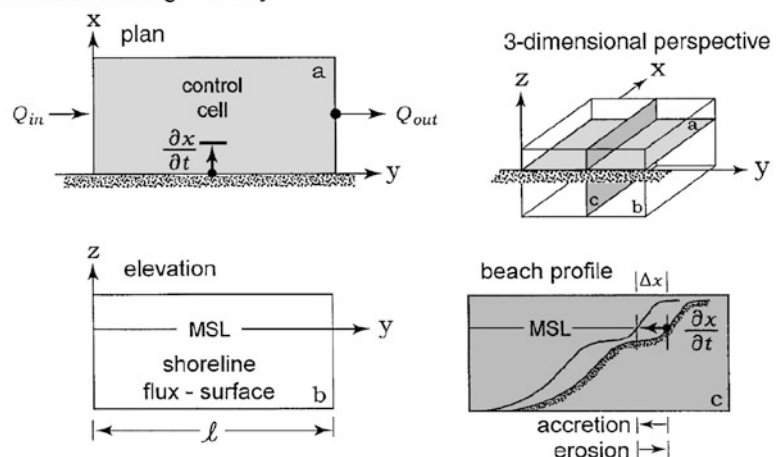
on Beaches, Fig. 3 The balance of sediment for a propagating accretion/erosion wave. (Modified from Inman and Dolan 1989)

a. Sediment balance

$$\underbrace{\frac{\partial Q_{\ell}}{\partial t}}_{\text{local time rate of change}} = \underbrace{\frac{\partial}{\partial y} \left(\epsilon \frac{\partial Q_{\ell}}{\partial y} \right)}_{\text{diffusion}} - \underbrace{v \frac{\partial Q_{\ell}}{\partial y}}_{\text{advection}} + \underbrace{R(t)}_{\text{source/sink}}$$



b. Control cell geometry



that travels along the beach and surf zone while the ebb-tide bar forms an accretion wave that moves along the shore in deeper water. Their relative on/offshore positions depend on the inlet tidal velocities that are functions of the size of the inlet and the volume of tidal flow through it (Inman and Dolan 1989; Jenkins and Inman 1999).

Accretion/erosion waves associated with river deltas and migrating inlets are common site-specific cases that induce net changes in the littoral budget of sediment. However, it appears that accretion/erosion waves in some form are common along all beaches subject to longshore transport of sediment. This is because coastline curvature and bathymetric variability (e.g., shelf geometry and offshore bars) introduce local variability in the longshore transport rate.

Mechanics of Migration

An accretion/erosion wave is a wave- and current-generated movement of the shoreline in response to changing sources and sinks in the local balance of sediment flux along a beach. The downdrift propagation of the wave form is driven by advective and diffusive fluxes of sediment mass (Fig. 3a). For convenience, these processes are usually expressed in terms of the longshore flux of sediment volume Q_ℓ into and out of a control cell (Q_{in} , Q_{out} ; Fig. 3b). By convention, fluxes of sediment into the cell are positive and fluxes out are negative. The net change of the volume fluxes between the updrift and downdrift boundaries of the control cell ($Q_{in} - Q_{out}$ = divergence of drift) will result in a net rate of change in the position of the shoreline $\partial x/\partial t$. Shifts in shoreline position will in turn cause the beach profile within the control cell to adjust to new equilibrium positions (Fig. 3b). The new profile positions alter local wave refraction causing adjustments in the flux of sediment leaving the control cell (Q_{out}). The variation in Q_{out} will alternately accrete and erode the beach downdrift of the control cell. As a consequence, propagation of the accretion/erosion wave involves a chain reaction in the local sediment flux balances. The reaction is set off by a disturbance on the updrift side of the control cell that yields a shoreline response on the downdrift side.

At a tidal inlet, these dynamics are impacted by additional fluxes of sediment into or out of a control cell centered at the inlet. When the tidal transport of sediment is ebb-dominated ($\Delta Q_t > 0$), the sediment flux into the control cell builds the ebb-tide bar and increases the rate of sediment that passes over the bar to the downdrift side of the inlet (Fig. 2). This stabilizes the inlet position by decreasing deposition on the updrift side and erosion on the downdrift side. Flood-dominated tidal transport ($\Delta Q_t < 0$) has the opposite effect and will cause the inlet to migrate faster (Jenkins and Inman 1999).

Cross-References

- Beach Features
- Beach Processes
- Coasts, Coastlines, Shores, and Shorelines
- Energy and Sediment Budgets of the Global Coastal Zone
- Littoral Cells
- Longshore Sediment Transport
- Scour and Burial of Objects in Shallow Water
- Sediment Budget

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Africa, Coastal Ecology

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Introduction

The continent of Africa straddles the equator and extends to about 35° latitude both north and south. It is, therefore, dominated by warm water regions and its coastline includes environments such as coral reefs and mangroves, except at the northwest and southwestern extremes, where temperate environments occur. The African coastline, excluding Madagascar (4,000 km), totals 35,000 km. A review of the literature

dealing with sandy beaches in Africa (Bally 1986) found the earliest paper to have been published in the 1880s and, of more than 1,000 papers published over 100 years, 19% concerned north Africa, 16% west Africa, 15% east Africa and Madagascar, and 55% southern Africa, particularly South Africa. Early papers were primarily taxonomic but later publications were mainly ecological. English was the language of 59% of the papers, followed by French (29%), German (7%), and Italian (3%). This sketch for literature on sandy beaches is probably representative of all papers dealing with coastal ecology in Africa. If so, it can be concluded that coverage of coastal ecology in Africa is patchy and, whereas the coastal ecology of South Africa has been extensively researched, most other regions are less well known.

The Physical Environment

Precipitation along the African coast is mostly low, with the exception of Madagascar, central east Africa, and west Africa, where rainfall is high (Fig. 1). It is in central to northwest Africa that the largest rivers enter the sea, whereas greatest evaporation occurs in northwest Africa and in the Red Sea. Lowest salinity in coastal waters is therefore in the Gulf of Guinea and highest values are in the Red Sea. The coasts of Africa experience semi-diurnal tides, with the exception of the east coast of Madagascar, the horn and the Mediterranean coast of Egypt, where tides are mixed. Spring tide range mostly approximates 2 m, except in the Mozambique Channel where it is larger and can exceed 4 m, and along the Red Sea and Mediterranean coasts where tides are small (Davies 1972). The entire east coast experiences temperatures permanently above 20 °C, as does the central west coast. Upwelling in the southwest and northwest lowers temperatures regionally. Major currents are the Somali and Mozambique/Agulhas in the east and the Benguella, Guinea, and Canary currents in the west. Wind and wave energy is greatest in the south and lowest in the equatorial regions, the Mediterranean and the Red Sea.

Regional Descriptions

West Africa

The northwest coast, from Morocco to Senegal, is sandy and relatively unindented, but from Dakar southeast to Monrovia it becomes very indented and there are numerous off-shore islands. Moving further south, the coast becomes low lying and sandy and more deltaic in nature, with large lagoons further east. The eastern area is dominated by the delta of the Niger River. Further south, the Congo River has a major influence and freshwater can extend far offshore. High

precipitation and numerous rivers in central west Africa result in warm, low salinity water, known as Guinean waters, circulating in the Gulf of Guinea.

The most notable feature of this coast is the extent of mangrove forests, estimated at 28,000 km² or 15% of the world's mangroves. Coastal lagoons range from tidal swamps and seasonal marshlands, associated with the river deltas and estuaries, to extensive coastal lagoons, which are typical of the Guinea coast. Sea grass beds are not well developed and there are no true coral reefs due to the cool waters of the Benguela and the Canary currents. Sandy beaches occur throughout the coast, particularly along the coast of Mauritania and northern Senegal. There are permanent areas of upwelling off Senegal, Zaire, and Namibia, coupled to the Canary and Benguela currents.

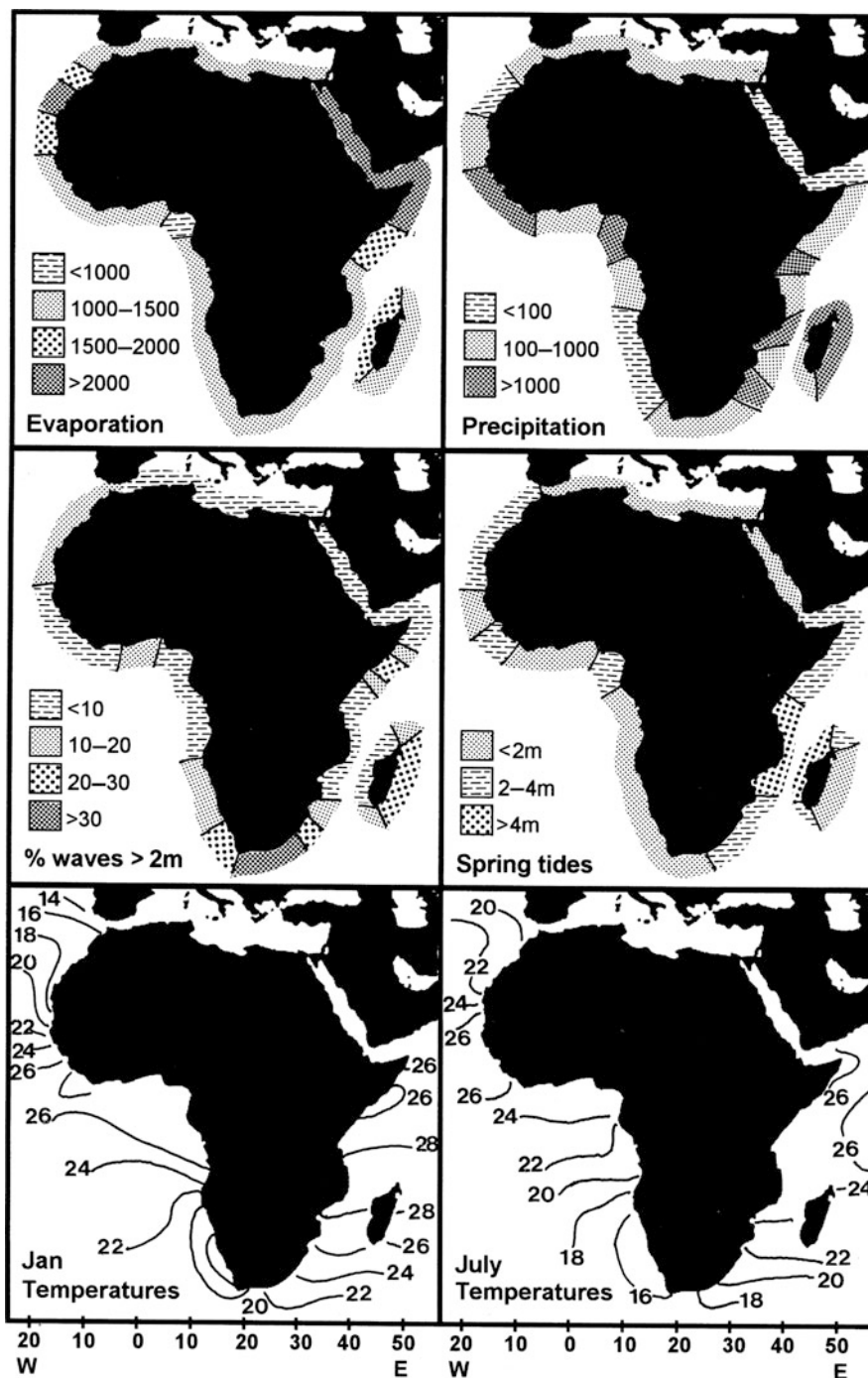
The marine resources of the west African region are important in the local and regional economies. Although seaweeds are not diverse, invertebrates (lobsters and shrimps) are exploited and there is a rich ichthyofauna with about 250 species and high levels of endemism. Five species of marine turtles nest along the coast and on the islands and there are three species of crocodiles. Millions of migratory birds, especially waders, visit the coast seasonally. The northwest African manatee occurs in suitable habitats from Senegal to Angola.

The coastal human population in west Africa exceeds 50 million and the rate of industrialization and urban population growth is accelerating along the coastal zone. Oil production has increased markedly in Nigeria, Angola, and Gabon. Tourism is an important earner of foreign exchange in the economies of several countries including Gambia and Guinea Bissau. Increased fishing effort and the introduction of more efficient technologies has led to over exploitation of fisheries resources.

East Africa

Oceanic current patterns and monsoon seasons have a major influence on the biogeography and biodiversity of east Africa. The main oceanic currents are the Somali, Madagascar, and Mozambique/Agulhas currents. The continental shelf is mostly narrow, varying from a few kilometers off Pemba to nearly 145 km in the Bight of Sofala, Mozambique. The shelves and banks are areas of intensive biological activity and productivity; in general, the narrower the shelf, the less productive the sea area. The western Indian Ocean is fairly poor in fisheries compared with other regions. The mainland coast is relatively unindented due to the absence of large rivers and coastal waters are moved by coast parallel currents. Much of the Mozambican coastline consists of low coastal plains forming long stretches of sand beaches and dunes interspersed with muddy rivers. North of the Zambezi estuary, many small coral islands fringe the shore. To the south, the islands of the Bazaruto group and Inhaca are mainly sand.

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Fig. 1 Physical features of
 African coasts



The total area of mangrove coverage in east Africa is about 10,000 km² or 5% of the world mangrove. Diversity of mangrove communities in east Africa is higher than in the west, with 10 species compared with only 7. Sea grass beds are found in all countries where there are low energy environments. The east African coast features fringing and patch reefs along the coastline from Somalia to Mozambique. Sandy beaches are

well developed throughout. Small-scale local upwelling occurs seasonally, particularly in the waters off Somalia.

The coast of east Africa supports an enormous diversity of life. More than 350 marine algal species and more than 135 species of coral are known and there are more than 900 fish species, mostly associated with coral reefs. The five marine turtles are abundant. The region has a varied

assemblage of seabirds, including frigate birds, tropic birds, boobies, shearwaters, terns, noddies, and gulls. The dugong is present, but its distribution and migration along the mainland coasts is not known. At least 15 species of cetaceans occur along this coast.

The coastal human population in east Africa is about 80 million. Shipping/port development and tourism are the fastest growing industries in the east African coastal zone. These industries are becoming incompatible with conservation in areas where pollution threatens to destroy the scenic beauty of beaches and coral reefs.

Red Sea

The Red Sea, over 2,000 km long and up to 360 km wide, is an important repository of marine biodiversity on a global scale. It features a range of coastal habitats including coral reefs, mangroves, salt marshes/sabkhas, rocky and sandy beaches, dune systems, and sea grass beds. These habitats accommodate more than 500 species of seaweeds, 1,000 fish species, 200 species of stony coral, 130 species of soft corals, more than 200 species of echinoderms, and 200 species of birds. There is a considerable variety of coral reef types with great structural complexity. The diversity and the number of endemic species of corals are extremely high.

The coastal human population is about five million. Urbanization, oil and other industries are developing rapidly in the Red Sea and tourism plays a major role in the economy of several countries. Pressures from recreation and tourism are high in the northern part and in the Gulf of Aqaba. Much of the oil produced in the Middle East is transported through the Red Sea and land-based activities (e.g., power and desalination plants, sewage treatment plants, industrial facilities, solid, wastes, and others) are adversely affecting the coastal and marine environment.

North Africa

The north African part of the Mediterranean is warm and arid. While it exhibits a low level of biological productivity, the Mediterranean Sea, as well as the surrounding land, is characterized by a moderate degree of biological diversity. Among the ecosystems that occupy coastal marine areas, the rocky intertidal, estuaries, and sea grass meadows (*Posidonia* beds, especially in Libya and Tunisia) are of significant ecological value. The Nile delta is a major feature of this area. Endangered species include the Mediterranean monk seal, marine turtles, and marine birds. It has been estimated that 500 species from the Red Sea have entered the Mediterranean through Suez, since damming the Nile has reduced the freshwater barrier (Por 1968). The coastal population is about 40 million and human impacts are considerable. Tourism is of great importance in several areas.

Southern Africa

This region, including South Africa and Namibia, is subtropical to temperate, with cool upwelling waters in the west. It mostly consists of open high-energy coast receiving swell from the Southern Ocean. There are numerous large zetaform bays on the exposed south coast and few sheltered waters. The west is arid and the east is moist. Sandy beaches and rocky shores dominate and there are many small estuaries. Large volumes of sand transport characterize the south, with extensive, coupled dune and beach systems. Rocky shores are notable for their diversity of large limpets. Productive kelp beds occur on the west coast, and salt marshes are typical of estuaries and lagoons in the south and west. Biodiversity is highest in the east and decreases towards the west coast. Fur seals and penguins occur around the islands and two species of turtles breed on the east coast of South Africa.

There are a number of invertebrate fisheries: in addition to abalone in the southwest and subsistence fishing on the east coast, lobsters are exploited in the south and west, and penaeid prawns on the east coast. Demersal and seine fisheries are well developed, especially on the west coast. The coastal population is about 15 million and there are five major harbors. A wide range of activities impinges on the coast: besides recreation, tourism and industry, these include dune mining for heavy minerals, beach mining for diamonds, and damming of rivers.

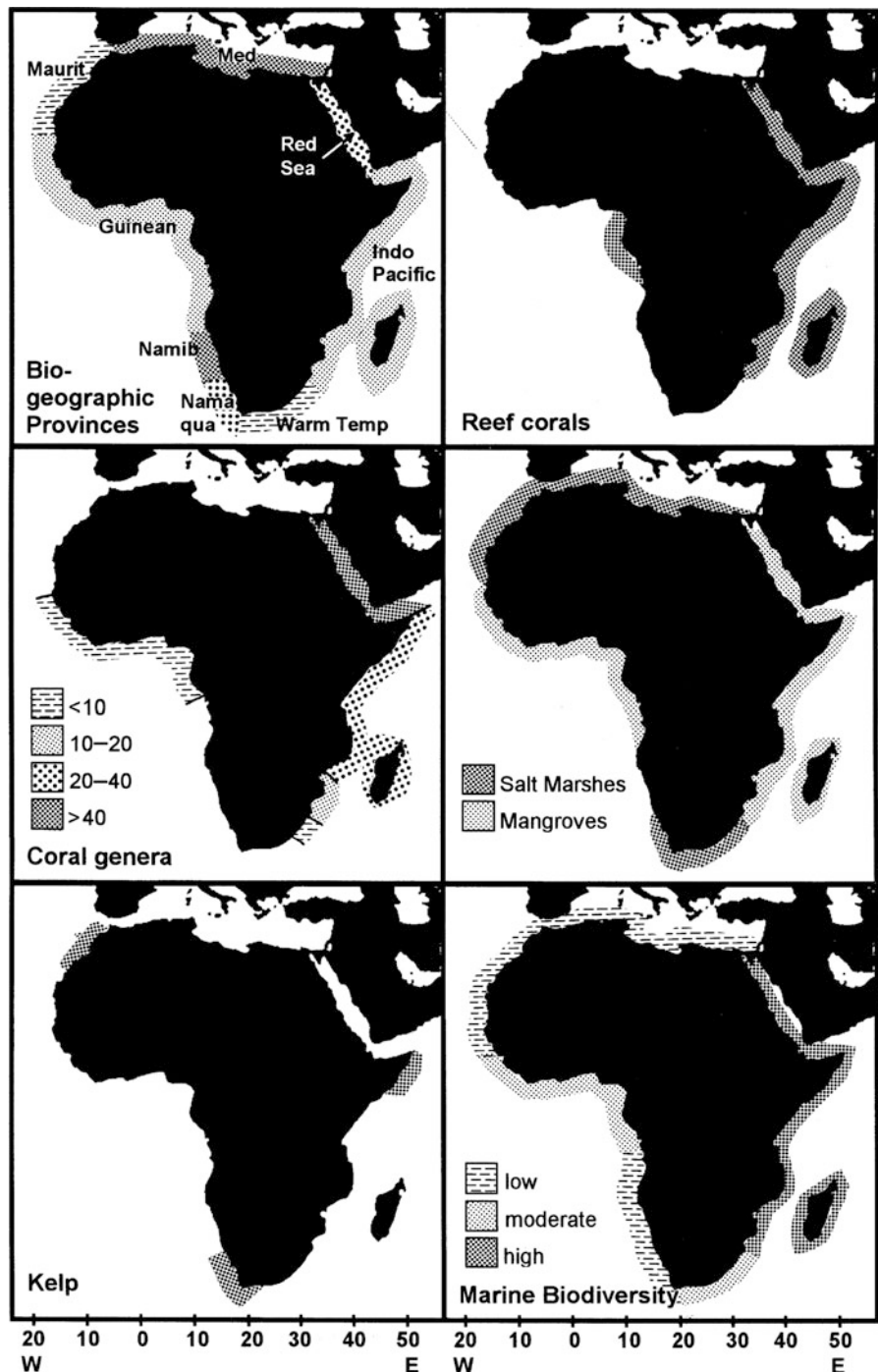
Biogeography

The “shallow water” or coastal marine biogeography of Africa is dominated by warm water regions (Fig. 2), the Indo-Pacific component on the east coast and the Atlantic component on the west coast. The Red Sea and Mediterranean coasts have distinctive, warm water faunas, which may be considered tropical and subtropical, respectively. Their marine boundaries are at Gibraltar and at the entrance to the Red Sea. Southern Africa harbors three provinces in addition to the Indo-Pacific component to the east: a warm temperate region on the extreme south coast and two temperate regions, the Namaqua and Namib provinces on the southwest. The eastern boundary between the southern warm temperate region and the Indo-Pacific is a broad transition region spanning the Eastern Cape and Natal coasts of South Africa. The junction between the Namaqua and Namib provinces, both influenced by upwelling, lies in southern Namibia near Luderitz and the boundary between the Namib and the Guinean is in southern Angola. The boundary between the tropical Guinean province and the north African.

Mauritanian province lies around 15°N. Biodiversity is greatest in the Indo-Pacific province and in the Red Sea

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Fig. 2 Biogeographic features of African coasts



on the east coast and lowest in the temperate provinces in the southwest and northwest (Namaqua, Namib, and Mauritanian) and the Mediterranean.

Environments

The warm water provinces are characterized by corals, mangroves and, in some cases, sea grass beds, whereas the

temperate provinces harbor kelps, where there is upwelling, and salt marshes. Sandy beaches and associated dunes, rocky shores and estuaries occur throughout.

Coral Reefs

True coral reefs occur around Madagascar and along the African east coast from the Red Sea to Mozambique with some reef forming (hermatypic) corals extending further south into South Africa. On the west coast, the influence of

upwelling restricts corals to a more limited region around the equator and islands off the northwest, but these are not true coral reefs. Those on the east coast are much richer in genera, being a part of the Indo-Pacific region (Fig. 2). Coral reefs are mostly of the fringing type and occur predominantly in two provinces, the Red Sea and East Africa/Madagascar (Sheppard et al. 1992). Reef development in between, around the horn and Gulf of Aden, is limited by upwelling. In the northern Red Sea reefs are well developed and drop into deep, clear water, whereas in the south they occur in shallower, more turbid water and are less well developed. Besides fringing reefs, barrier reefs, patch reefs, and even atolls can occur. Coral diversity in the Red Sea is high and there is a clear north/south zonation into 13 communities. Coral reefs also display a depth zonation with maximum coral diversity at 5–30 m depths, depending on exposure and water clarity. Coral cover is mostly less than 50% on slopes but in sheltered areas *Porites* can attain 80% cover. Coral reefs support a diverse associated fauna of cryptic invertebrates and fishes, which play a variety of roles in these complex communities. The Red Sea harbors more than 1,000 species of fishes, many associated with coral reefs. Greatest fish abundance occurs near the reef top, whereas greatest species richness tends to occur at depths of 10–15 m.

Sandy Beaches

Sandy beaches, backed by dunes, are the dominant coastal form, comprising as much as 70% of the coastline, and have been well studied in South Africa. Beach form is controlled by the interaction between sediment particle size and wave and tide energy. Conditions of high wave energy, large tide ranges, and fine sand give rise to wide flat beaches referred to as dissipative; whereas, narrow steep beaches, called reflective beaches, develop under conditions of low wave and tide energy and coarse sand. African beaches range from high-energy dissipative systems on the southwest tip of the continent, through intermediate types to microtidal reflective beaches in the Mediterranean and Red Sea. Macrotidal beaches and sand flats occur in the Mozambique Channel. High-energy dissipative systems have been shown to be richest in terms of productivity, with large populations of filter feeders (Brown and McLachlan 1990). Such beaches, together with macrotidal flats, support diverse benthic faunas. Reflective systems, which are typical of tropical regions with modest tide ranges, generally support lower diversity and lack surf zones. Temperate high-energy beaches with extensive surf zones often develop blooms of surf zone diatoms, which fuel rich food chains. Besides meiofauna and microfauna, beaches support a macrofauna of scavenger/predators, filter feeders and deposit feeders. Among the macrofauna, donacid clams, gastropods of the genus *Bullia*, ghost crabs (*Ocypode*), cirolanid isopods, mysid shrimps, and polychaete worms are typical. The species richness per beach ranges

from less than 10 species on reflective beaches to more than 30 on tropical dissipative beaches and tidal flats. Zonation on sandy beaches is less marked than on rocky shores and throughout most of Africa consists of three zones: ghost crabs at the top of the shore, cirolanid isopods on the mid-shore, and a variety of species lower down. In the temperate areas, sandhoppers (talitrid amphipods) and/or oniscid isopods occur at the top of the shore. Beach fauna is, therefore, controlled primarily by physical factors and beach type, with biogeography and climate playing a lesser role.

Sea Grass Beds

In the Mediterranean, the Red Sea and throughout the western Indian Ocean, sea grass beds are common in intertidal areas, coastal lagoons, shallow sandy bottoms with good light penetration and sandy areas adjacent to shallow reefs. At least 10 species are common and include members of the genera *Possidonia*, *Thalassia*, *Halodule*, *Syringodium*, *Halophila*, *Cymodocea*, and *Thalassodendron*. Sea grasses are not true grasses but are closer to pondweeds. They are aquatic plants with leaves above the surface and rhizomes and stems buried in the sand. Sea grass beds create habitats distinguished by high primary productivity, enhanced by encrusting algae epiphytes. However, there is limited direct herbivory, probably due to distasteful compounds and indigestibility of seagrasses. Instead, seagrasses are consumed via detritus food chains after processing by microorganisms. Sea grasses stabilize the sediment, create habitat, and serve as nursery areas and support a high diversity of associated species, especially molluscs, polychaets, crustaceans, and fishes. Many commercial fishes utilize them as nursery areas. Ten species occur in the Red Sea, *Halodule* and *Halophila*, being the commonest.

Dunes

Dune forms depend on rainfall (vegetation growth) and sediment transport (supply X wind) and are closely coupled to beach type. Simple vegetated foredune ridges and hummocks are the most widespread type. Such dunes typically occur behind low energy and reflective beaches and are characteristic of the moist tropics. Transgressive dune sheets occur in windy areas with large volumes of sand transport, for example, the south coast of South Africa, and parabolic dunes occur in places with predominantly unidirectional winds. Coastal dunes are rapidly colonized by plants, *Ipomoea* and *Scaevola* being typical foredune pioneers. These may be followed by scrub and thicket until climax forest is reached. In the arid southwest and north of Africa coastal dunes are highly mobile and support limited vegetation. Climax dune vegetation may be scrub in arid areas, or even pioneers in hyperarid situations. In the moist tropics, forest may extend right down to the beach. Coupled to landward succession in dune vegetation is a change in animal communities: bird

diversity increases as vegetation structure becomes more vertical; insects and small mammals also respond to this gradient. In general, therefore, coastal dune ecology is a function of dune forms and vegetation succession, which in turn are controlled by sand supply and climate.

Rocky Shores

Rocky shores form about 30% of the coastline of Africa. Intertidal rocky-shore organisms are exposed to a wide range of physical conditions. Unlike the situation on sandy beaches, organisms on rocky shores cannot burrow to escape adverse conditions: they must simply be tough enough to tolerate the fluctuations. Tides establish a gradient of physical stress, with the high-shore being sun-baked and desiccated, and the low-shore more mild. This gradient leads to a distinctive vertical zonation of organisms, with the top of the shore characterized by a small number of species, low biomass and productivity, and progressive increases in all these variables as one moves down the shore. Physical stress tends to limit the extent to which species can advance up the shore, but biological interactions between species often set limits to how far they extend down the shore. For example, competition from mussels ousts barnacles; grazers, such as limpets and chitons, control the growth of algae; predators, such as whelks, crabs, fish, and birds, limit the zonation of their prey. Moving horizontally along rocky shores, different gradients come into play. The first of these is wave action, which operates on a scale of meters to kilometers. Mobile predators and grazers tend to be inhibited by waves, whereas algae benefit from a reduction in grazing and from an increase in the turnover of nutrients. Filter feeders, such as mussels, are enhanced because waves import the organic particles that comprise their food. The overall effect is that biomass and productivity are highest on wave-beaten shores, such as those occurring on the south coast of Africa (Branch and Griffiths 1988).

At larger scales of 100s to 1,000s of kilometers, there are gradients of productivity. High nutrient levels associated with upwelling accelerate algal growth, supporting high biomass of grazers and predators but a low diversity of species. On a similarly large scale, there are climatic gradients from temperate to tropical conditions. High temperatures in the tropics lead to greater physical stresses, so that rocky shores there tend to be sparsely occupied, with low biomass and productivity, but high levels of species diversity. Tough coralline turfs become a dominant element among the seaweeds. Grazers and sedentary predators are often confined to shelters during low tide to avoid potentially lethal conditions. On the other hand, mobile predators, such as crabs and fish, are more abundant in the tropics than on temperate shores. In sum, vertical changes in community structure on rocky shores are dictated by gradients of physical stress overlain by biological interactions. Moving horizontally along rocky shores, community patterns and processes controlling them change

radically in response to wave action, productivity, and climate.

Kelp Beds

Kelps are large and complex in structure, with a root-like holdfast, a long stipe and frond-like blades. Kelps are fast growing plants and are restricted to coastal areas where sunlight is readily available and nutrient levels are high. In Africa, they are confined to zones of upwelling, where cool and nutrient-rich water is brought to the surface by winds that drive surface waters offshore. Upwelling is concentrated on the west coast of southern Africa, off west Africa and in the region of the Somalian horn of Africa, where the world's only tropical upwelling areas can be found. Kelp forests thus occur in the southwest in the Namaqua and Namib provinces (*Ecklonia*, *Laminaria*, *Macrocystis*) and in the northwest in the Mauritanian province (*Laminaria*) and off Somalia. Kelp plants are of direct economic significance because they produce alginic acid, widely used in food and other products.

Kelps form dense underwater forests, which break the force of wave action. They are themselves, however, powerfully influenced by waves, which tangle and tear out swathes during storms. Whole kelp plants wash ashore on sandy beaches where they contribute substantially to energy flow. Sandy shores "supplemented" in this way have high levels of biomass and productivity, and their life is concentrated at the top of the shore where the kelp deposits. Surprisingly little kelp is directly eaten by herbivores; most is abraded from the growing plants and contributes to a pool of organic matter that fuels particle-feeders and filter feeders, such as sea cucumbers, mussels, ascidians, and sponges that dominate much of the floor of kelp beds. Large fragments of kelp also break off and support urchins and abalone. Intertidal grazers also benefit: the highest biomasses of grazers ever recorded in the world occur on rocky shores on the west coast of South Africa. Their existence depends on the vital "subsidy" they receive in the form of drift from adjacent subtidal kelp beds. Clearly, kelp beds play significant ecological roles that extend well beyond their confines. They are commercially valuable in their own right, but also sustain other species of commercial importance, such as abalone.

Estuaries

Estuaries, lagoons, and river mouths are highly variable around the coasts of Africa, depending on climate. Other than for the Zambesi in the east, all the major rivers (Kunene, Orange, Congo, Niger, Volta) drain to the west and the Nile empties into the Mediterranean. The Nile (20,000 km²), Zambesi (700 km²), Volta (9,000 km²), Senegal (8,000 km²), Oueme (1,000 km²), and Niger (36,000 km²) have extensive delta systems. Large coastal lagoons occur in west and in southeastern Africa and many wetlands are of considerable conservation value. In the arid

areas of southern and southwest Africa, the Red Sea and the Mediterranean coast, most estuaries are ephemeral and close during the dry season. Many estuaries have been impacted by damming of rivers, in some cases the reduction in freshwater supply eliminating normal salinity gradients and/or reducing floodplains (e.g., Niger). Mangroves, salt marshes, and phytoplankton contribute to primary production and support benthic fauna of high abundance but low diversity. Estuaries are important nursery areas for penaeid prawns and a variety of fishes.

The regulation and impoundment of freshwater in river systems is probably the single most important threat to natural functioning of estuarine systems in many parts of Africa. Most of these estuaries are classified as temporary open/closed. Floods, in particular, are important in maintaining functional links between the estuaries and the sea, but because of impoundment schemes, changes in the frequency and intensity of flood events is becoming evident. Many estuaries close more frequently and for longer periods since the removal of accumulated sediment in marsh channels by floods is not as effective as before. Thus, many estuaries are beginning to function differently compared with the natural state. Not only do freshwater abstraction schemes lead to negative downstream impacts, but they also have the capacity to influence the marine nearshore. Major regulations along the Zambesi river (Kariba and Cahora Bassa dams) are having a negative influence in coastal waters, with die back of mangroves and a collapse of the coastal prawn fisheries (Davies et al. 1993).

Salt Marshes

Salt marshes are temperate habitats, mostly associated with estuaries. African salt marshes occur in the Mediterranean, on the northwest and on the south and southwest coasts. Most African salt marshes are of limited extent; for example, total salt marsh area in South Africa is 1,700 ha, most of which is confined to five systems. There is clear zonation of the marsh flora from the subtidal to the top of the intertidal zone, *Spartina*, *Zostera*, *Sarcocornia*, and *Limonium* are the most typical genera. These African salt marshes are distinguished from those elsewhere by their warm temperate character and limited size. Associated fauna includes a variety of marsh crabs and shrimps, wading birds and, at high tide, fishes.

Mangroves

Total world mangrove area is 181,000 km² and 20% of this is in Africa, 15% on the west and 5% on the east coasts (Spalding et al. 1997). On the east coast, in the Red Sea and Madagascar, they form part of the Indo-Pacific province; there are 10 species and the most widespread are *Avicennia marina*, *Rhizophora mucronata*, and *Sonneratia alba*. On the west coast of Africa their affinities are with New World tropical regions and, of the seven indigenous

species, *Avicennia germinans*, *Rhizophora racemosa*, and *Laguncularia racemosa* are widespread. Distribution may be as much limited by aridity as temperature. Estuaries may harbor large mangroves systems but they can also occur as narrow strips along the coastline where rivers are absent. Associated fauna includes mudskippers *Periopthalmus*, mangrove snails *Cerithidea*, fiddler crabs, *Uca*, and oysters. Fishes and penaeid prawns utilize mangroves as nursery areas. Mangroves are important sources of detritus for estuarine food chains. Wood collection threatens mangroves in several areas (e.g., Mozambique).

Islands

Coastal islands form important rookeries for seals and nesting sites for seabirds. Colonies of Cape fur seals, Jackass penguins, Cape gannets, cormorants and other seabirds occur on numerous coastal islands, especially off South Africa and Namibia. Socotra is also an important breeding area for boobys and terns and also harbors six endemic species of birds and eight endemic reptiles. Red Sea islands support colonies of gulls and terns. Islands off Mauritania are important for migratory birds and support large breeding colonies of gulls and terns.

Utilization

Coastal resources are extensively utilized throughout Africa. These activities range from subsistence gathering of shellfish and mangrove wood to commercial exploitation of abalone, kelp, lobsters, and inshore fishes. Subsistence utilization can cause degradation and damage to coastal habitats. Dynamite fishing also threatens Africa's coastal ecosystems in some localities as it disturbs coral reefs and lagoon systems. Nonliving resources that are exploited include oil (Angola, Nigeria), diamonds in beach and nearshore sediments (Namibia), heavy minerals in dune sands (South Africa), sand, rock, and groundwater. The coast is also a focal point for recreational activities including swimming, surfing, angling, and diving. Indeed, the African coastline is becoming as important as its game reserves and ancient civilizations in attracting tourism. For most recreational activity, sandy beaches are focal points. One third of the coastline is considered to be under threat from developments and other human activities.

Conservation and Management

Marine and coastal resources contribute significantly to African economies, especially through fisheries (e.g., Namibia) and tourism (e.g., Mauritius). Current development trends and pressures from increasing urbanization and

industrialization are steadily degrading fragile ecosystems. Pollution, mining, and oil exploration are also threatening coastal ecosystems. Oil spills due to well blowouts have caused serious problems in the Niger delta, decimating “black water” biodiversity. In addition, toxic wastes from developed nations have been illegally dumped along the coasts of poor African nations. Many industries dispose of untreated wastes directly into rivers running into the sea. In the Red Sea and in north Africa there is an increasing risk of pollution as over 10^9 t of oil are transported through the area annually and there are limited maritime traffic regulations. The coastal countries of Africa are also susceptible to the problem of accelerated coastal erosion. This is driven by natural processes that are exacerbated by sealevel rise, upstream construction of dams, other coastal infrastructure, and clearing of mangrove systems.

In recent years, Africa has seen major political, economic, and social changes and has transformed from a rural society to a complex modern region whose ties to natural capital remain strong despite economic development. Population growth (2.8%) is almost twice the global average (1.5%), far in excess of the average rate of economic growth. In many African countries, national parks and conservation areas include parts of the coast as increasing human impact necessitates protection and management. Coastal reserves, which have been established in a number of areas to afford protection or control utilization, need to be expanded, to involve the local populations and to be better managed. Important resources/habitats requiring protection are dunes, mangroves, coral reef, and islands. Integrated coastal zone management, in its infancy or absent in most of Africa, is essential for the future wellbeing of these spectacular coasts.

Cross-References

- [Africa, Coastal Geomorphology](#)
- [Beach Processes](#)
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Africa, Coastal Geomorphology

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The African continent measures 30×10^6 km² and its relatively unbroken coastline is 30,000 km long, compared with the 70,000 km coast of Asia, which is only 1.5 times larger than Africa, and the 76,000 km coast of smaller North America (24×10^6 km²) with its numerous Arctic islands. Over long distances, the African coast is unbroken by sizable inlets, and its major river mouths, except the Congo, are either deltaic or blocked by sand barriers. Excepting Madagascar (587,000 km²), no large islands lie off the African coast.

Offshore, Africa’s continental shelf covers only 1.28×10^6 km² compared with 9.39×10^6 km² for Asia and 6.74×10^6 km² for North America. The shelf averages only 25 km in width, wider off southern Tunisia, Guinea, and major deltas, and reaching 240 km wide across the Agulhas Bank, but narrowing to 5 km off Somalia, northern Mozambique, and Kwa-Zulu. This narrow shelf and paucity of sheltering islands allow deep-water waves and surface ocean currents to approach unmodified close to the mainland shore where they are unusually influential in moving sediment (Orme 1996).

Explanations for Africa’s relatively smooth coastline and narrow continental shelf are to be found in the tectonic processes that triggered the rupture of Gondwana in Mesozoic time and in the geomorphic processes that have shaped the coast more precisely during later Cenozoic time.

Coastal Origins

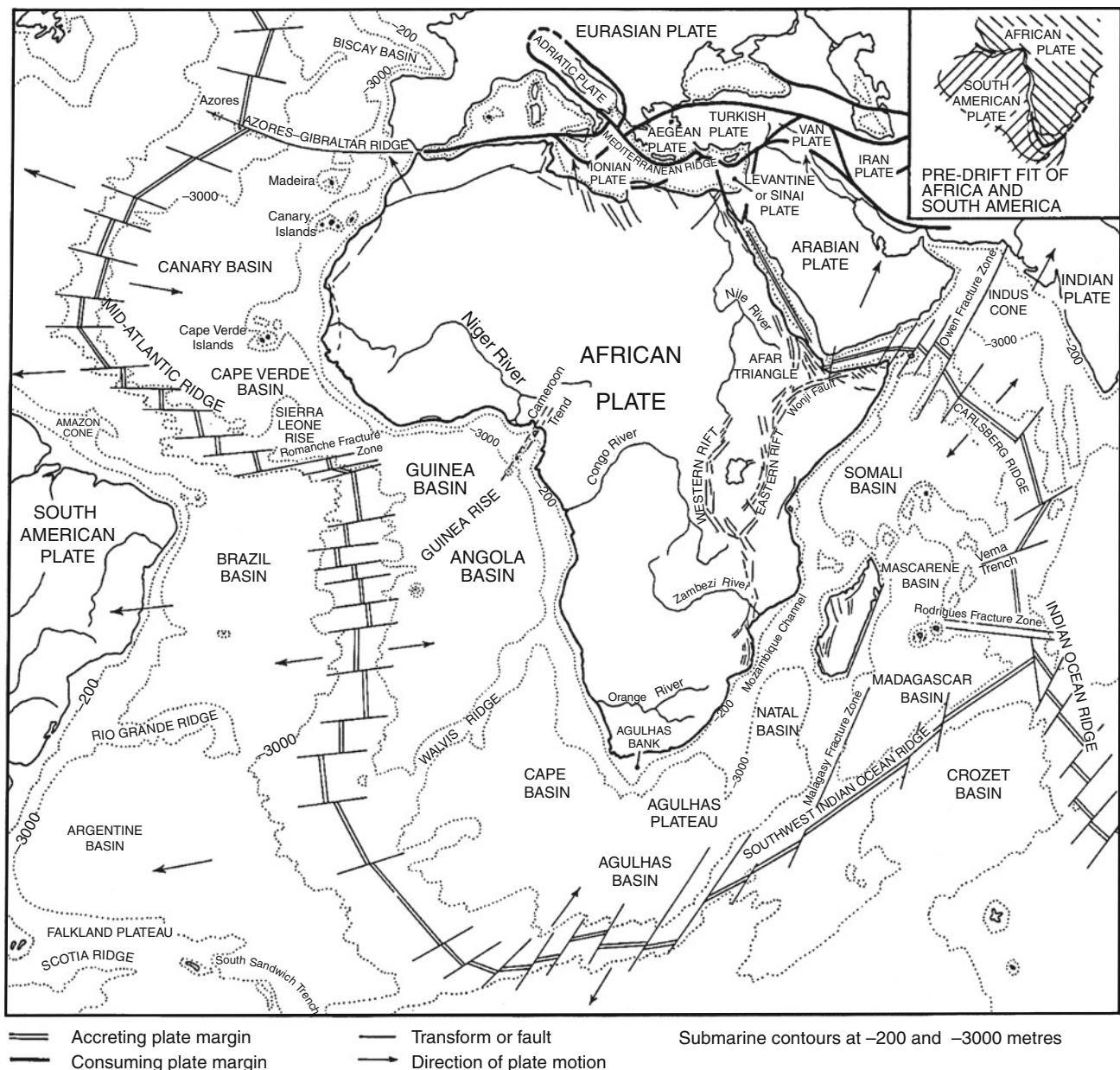
The broad outlines of Africa’s coastal margins may be explained in terms of plate-tectonic events over the past

200 Ma (million years before present). Africa's margins were initially blocked out by the rupture of Gondwana in Mesozoic time, and further modified during the progressive opening of the Atlantic and Indian Oceans, the gradual closure of the Tethys Ocean, and the Cenozoic opening of the Red Sea (Fig. 1; Orme 1996; Summerfield 1996).

Prior to these events, Africa's cratonic nucleus had been fused during the Panafrican–Brasiliano orogeny in later Neoproterozoic time (650–530 Ma) with cratons from other southern continents to form the supercontinent of Gondwana. Later, towards the close of Paleozoic time (330–260 Ma), Gondwana

became joined to the supercontinent of Laurussia to form a vast single landmass, Pangea, a suture marked prominently by the Appalachian (Alleghanian) and Atlas collisional orogens of North America and northwest Africa, respectively (Ziegler et al. 1997; Orme 2002). Compared with Gondwana, however, Pangea did not last long. Following Triassic crustal extension (248–206 Ma), North America began separating from Africa along the Appalachian–Atlas suture and, by mid-Jurassic time (180–160 Ma), an incipient north-central Atlantic Ocean, underlain by oceanic basalts spewing from a spreading center beneath the Canary Islands, lay off northwest Africa (Steiner

A



Africa, Coastal Geomorphology, Fig. 1 Tectonic relations of the African Plate. (From Orme 1982, with permission of Kluwer Academic Publishers)

et al. 1998). The later development of this coastal margin reflects variable rates of sea-floor spreading, epeirogenic flexuring, marginal basin development, and sedimentation.

The continent's western margin farther south was initiated by sequential events, which caused South America to separate from Africa over a period of 100 Ma, from early Jurassic to middle Cretaceous time (Uchupi and Emery 1991; Milani and Filho 2000). In the north, following Triassic crustal extension, sea-floor spreading in the early Jurassic (190–180 Ma) introduced ocean waters to the Guinea–Sierra Leone margin. In the south, the Malvinas (Falkland) Plateau, which had formerly wrapped around Africa's southern tip, began shearing westward in the early Jurassic (200–180 Ma), causing the continent's west-facing margin to unzip progressively northward from eastern Brazil. Later, in response to these stresses, right-lateral wrenching along a massive transform system began separating northern Brazil from the south-facing Guinea coast in middle Cretaceous time (120–100 Ma). Throughout this Atlantic margin, initial crustal extension and enhanced magmatism were followed by basin formation along the rifted margin and then, as sea-floor spreading moved Africa farther from the Mid-Atlantic Ridge, by widening oceanic tectonic and sedimentary regimes. Thus, magmatism at a common eruptive center on the Mid-Atlantic Ridge generated early Cretaceous volcanic rocks at Etendeka in northern Namibia and the Paraná flood basalts of southern Brazil, which on subsequent rifting and sea-floor spreading have since separated (Glen et al. 1997). The Liberian coastal margin off Cape Palmas reflects fracture zones that developed around 140 Ma, since when some 8,000 m of sediment have blanketed the shelf and slope. Farther south, the deep Angola and Cape basins developed during Cretaceous time to the north and south of the Walvis Ridge, respectively. South of this ridge, the western continental margin consists of downdropped basement blocks, aligned NNW and overlain by a prograded sediment wedge formed by debris brought down by the Orange River and other streams that have been eroding the Great Escarpment since it was first tilted upward in early Cretaceous time (Dingle and Scrutton 1974; Summerfield 1996). Islands such as Ascension and Tristan da Cunha in the central South Atlantic Ocean mark continuing hotspot volcanism on the Mid-Atlantic Ridge at the western edge of the African plate. At Africa's southern tip, the Malvinas Plateau sheared westward from the Hercynian Cape Fold Belt (280–230 Ma) along major transform structures during Jurassic time. Since then, and notably since epeirogenic uplift of the southern Great Escarpment in Cretaceous time, terrigenous debris has crossed the narrow continental shelf to blanket the continental slope and adjacent basins. Farther out, the submarine Agulhas Plateau was probably created by volcanism in fracture zones vacated by the Malvinas Plateau (Barrett 1977).

Following Permian tensions and Triassic saltwater intrusion, much of Africa's eastern margin was blocked out during earlier Jurassic time (200–160 Ma). In Mozambique, early tensional rifting produced N–S horst and graben that were later buried beneath debris from the Limpopo and Zambezi river systems (Dingle and Scrutton 1974). The N–S segments of coast south of the Zambezi and Limpopo river mouths reflect these structures, whereas the NE-trending coasts north of these rivers reflect right-lateral offsets of basement cratons, bringing Precambrian rocks close to shore in northern Mozambique. Some 6,000 m of Cretaceous and Cenozoic terrigenous sediment have since accumulated in offshore basins, including the massive Zambezi cone. Madagascar is a large fragment of continental crust that probably lay alongside western India before the latter began separating from eastern Africa in Jurassic time. Although Madagascar's place in Gondwana has been much debated, it probably translated southward to its present location from Tanzania and Kenya, with the reef-capped Comoro volcanic archipelago extruding along resultant fracture zones (Maugé et al. 1982; Ziegler et al. 1997). Farther north, the southeast margins of Somalia and Arabia also formed from the Jurassic rupture of Gondwana.

The Red Sea and Gulf of Aden coasts reflect relatively late intraplate separation of the African–Arabian portion of Gondwana along extensions of the East African Rift system. In early Cenozoic time, after prolonged stability, the region became the site of intensive magmatic and tectonic activity. By late Eocene time (35 Ma), vertical uplift had produced a massive Afro-Arabian dome which subsequently broke into Somali, Nubian, and Arabian segments. The Somali segment was bounded to the north and west by crustal flexures and tensional faults along which lateral displacement began in the Miocene, forming the discrete Somali plate. Since then, as the Arabian segment has moved north and east away from the Somali and African plates, new oceanic crust up to 200 km across has formed in the widening Gulf of Aden and Red Sea.

The Mediterranean coastal margin reflects a long and complex relationship with the European portion of the former supercontinent of Laurussia. The rupture of Pangea that began in late Triassic time, sometime before 200 Ma, permitted a large tropical ocean, the Tethys, to intervene between central Europe and north Africa at a time when the African plate lay wholly south of the Equator. As Africa subsequently moved northward and rotated anticlockwise across the Equator, transpressional tectonics reshaped the Tethys into several microplates and convoluted mid-Cenozoic (Atlas–Alpine) collisional orogens, which today form southern Europe and underlie the Mediterranean Sea. The effects of the Atlas orogeny are well revealed along the north African coast between Agadir and Gabes. Farther east, the situation is less clear because regional tectonism involves both transpressional and

transensional interactions between several microplates, activities that yield frequent earthquakes and volcanic events in southern Europe but not in north Africa. In the Aegean subduction zone, for example, the African plate is presently moving under the Aegean microplate at a rate of 2.7 cm yr^{-1} .

In summary, Africa's Atlantic and Indian Ocean coasts reflect divergent or passive plate margins inasmuch as the African plate has diverged from oceanic spreading centers from which the continent is now separated by vast expanses of post-Gondwanan basaltic ocean floor variably mantled with marine and terrigenous sediment. In embryo form, the Red Sea and Gulf of Aden coasts are also divergent margins but continuing crustal adjustments and volcanism belie the term passive. In contrast, the western Mediterranean margin is an active margin, but this categorization is less easily applied farther east.

Coastal and Offshore Geology

Tectonic origins apart, the present shape of the African coast also reflects the character of the rocks and structures introduced to the coastal zone during and since the breakup of Gondwana. The principal rocks influencing coastal features may be grouped into Precambrian basement, consolidated Phanerozoic platform covers, and poorly consolidated late Cenozoic sediments.

Precambrian rocks outcrop over 57% of Africa's surface, reflecting numerous sedimentary cycles whose deposits were intensely folded and fractured, metamorphosed, and often granitized during at least eight orogenic cycles. Gneiss, schist, quartzite, and migmatite are important but post-orogenic molasse deposits and tabular to strongly folded platform covers of sandstone, limestone, tillite, and volcanics also occur. Today, these basement rocks reach the coastal zone in the Anti-Atlas of Morocco, the Guinea coast between Monrovia and Accra, at intervals from Angola to Cape Province, and again in northern Mozambique, eastern Madagascar, and the Red Sea. Because intense fracturing favors deep weathering, these rocks rarely form high cliffs.

Phanerozoic rocks occur mostly as tabular platform covers occupying large basins between swells in the basement complex, but folded cover rocks form the Cape Fold Belt (Hercynian) of South Africa and the Atlas Fold Belt (Hercynian and Alpine) from Morocco to Tunisia. In Africa as a whole, the great extent of continental deposits and the paucity of marine sediment are noteworthy. The continental rocks include the Nubian Supergroup (Cambrian–Cretaceous) which straddles the Red Sea, and the 7,000 m thick Karroo Supergroup (Carboniferous–Triassic) of southern Africa whose tillites, shales, sandstones, and basalts form bold escarpments inland and sea cliffs at the shore.

In contrast, late Cenozoic sediments within the coastal zone are mostly marine sediments or fluvial deposits reworked by waves, currents, and winds. These materials are often poorly consolidated and thus erodible but, except in the Mozambique coastal plain, rarely extend far inland. More resistant late Cenozoic rocks include coral-reef limestones and aeolianites.

Offshore, there are about a dozen major basins around Africa, formed following the rupture of Gondwana and now partly filled with late Mesozoic and Cenozoic sediment. Where these basins straddle the present coast, both marine and terrigenous sedimentation have occurred, leading to epeirogenic seaward subsidence of the coastal margin which in turn favors more sedimentation (Orme 1996). The contrast between sedimentation in these basins and denudation in their hinterland has caused significant isostatic responses to loading and unloading, leading, for example, to 600 m of uplift inland from the southwest coast and adjustments to the Orange River below Augrabies Falls (Gilchrist and Summerfield 1990). Continued deformation and volcanism have complicated coastal evolution, notably in coastal extensions of the 7,000 km long East African Rift system and the Cameroon volcanic line.

Moving clockwise from Ras Asir at the Horn of Africa, the first basin encountered is the 7,000 m deep Somali Basin which merges south with the 8,000 m deep Kenya and 3,000 m deep Dar es Salaam Basins. When the Somali Basin formed along Gondwana's rifted margin, marine transgressions and carbonate deposition followed, but late Jurassic uplift of the Bur basement complex along NE-trending faults separated the interior of this basin from continuing carbonate deposition at the coast. Cenozoic rejuvenation of these faults has largely defined the east coast from Ras Asir to Tanzania (Orme 1985, 1996). Farther south, the 4,000 m deep Mozambique Basin has also been broken by N–S faults at the south end of the East African Rift system. The 7,000 m thick sedimentary cover of the Madagascar Basin has also been severely faulted. Along the west coast, a more or less continuous embayment of folded sediments and salt domes is represented by the 4,000 m deep Luanda Basin, the 3,000 m deep Cabinda Basin, and the 8,000 m deep Gabon Basin. The Niger Basin, the seaward extension of the Benue graben, contains up to 10,000 m of Cretaceous and Cenozoic deposits, including important petroleum resources beneath the Niger delta. The narrow Ivory Coast Basin is faulted to depths of 4,000 m and the Senegal Basin reaches depths of 7,000 m at the coast. The Tarfaya Basin farther north, bounded inland by the Zemmour Fault, contains 10,000 m of Mesozoic and Cenozoic deposits beneath Cape Juby.

Early evolution of the Mediterranean coast differed from other coasts because sediments in its developing marginal basins became involved in mid-Cenozoic orogeny, as seen

in the mountain arcs and deep sedimentary basins of the Algerian coast. Farther east, the 5,000 m deep Tripoli Basin is separated by NW-trending faults from the extensive 8,000 m deep Sirte Basin of Cyrenaica.

Relative Sea-Level Change

Since the rupture of Gondwana, continuing tectonic, isostatic, and eustatic forces have left their imprint on Africa's emerging coastal margins. For example, sea-level relations have been strongly impacted by Atlas tectonism along the continent's northwest margin during and since mid-Cenozoic time, while late Cenozoic rifting has impacted the Red Sea coast. Isostatic adjustments of Africa's plate margin to denudational unloading and sediment loading have also affected sea levels. Further, the eustatic high sea levels of mid-Cretaceous time caused shallow seas to transgress north Africa and link the Tethys with the Gulf of Guinea. After Paleocene regression, Eocene seas again flooded the Sahara and a shallow gulf persisted into Miocene time (~20 Ma) in Libya and Egypt.

Relative sea-level changes over the past few million years, and especially for the Quaternary, have been the focus of much research, in part because they aid prediction of future coastal behavior. In general terms, tectonism, isostasy, and eustasy may have combined locally to force Quaternary sea-level changes ranging from 300 m above to 200 m below present levels. Along the Moroccan coast, for example, a deformed Pliocene marine surface up to 20 km wide, the Moghrebien rasa, has been raised between 100 and 600 m above sea level, providing a useful measure of Alpine tectonism (Weisrock 1980). Flights of deformed Pleistocene coastlines also occur along the western Mediterranean coast, notably at Algiers, Bizerte, and Monastir, while raised coral reefs occur along deformed Red Sea and northern Somali coasts. Elsewhere, the stability of much of the African plate relative to high last interglacial seas (~125 ka) is indicated by little deformed coastlines and coral reefs a few meters above present sea level, notably along southern coasts (Orme 1973; Hobday and Orme 1975). Offshore, former sea cliffs, beaches, aeolianites, and river valleys, now submerged on the continental shelf, testify to low late Pleistocene sea levels (Orme 1974, 1976).

The last major global sea-level change, the Flandrian transgression which accompanied the melting of the last Pleistocene ice sheets and culminated around 5 ka (thousand years before present), initiated the present phase of coastal erosion and deposition. Locally, this transgression may have risen a few meters above present sea level, as suggested by the Nouakchottian deposits in Mauritania and Senegal, but such evidence for seemingly high Holocene sea levels may also be explained by local epeirogenic forcing or by climate-induced changes in hydrology and sediment delivery (Ausseil-Badie

et al. 1991). Relative sea-level rise during historic times is reflected in submerged Phoenician tombs and Roman ports along the Mediterranean coast, by submerged Arab legacies along the east coast, and more recently by tide-gauge records.

Coastal Processes

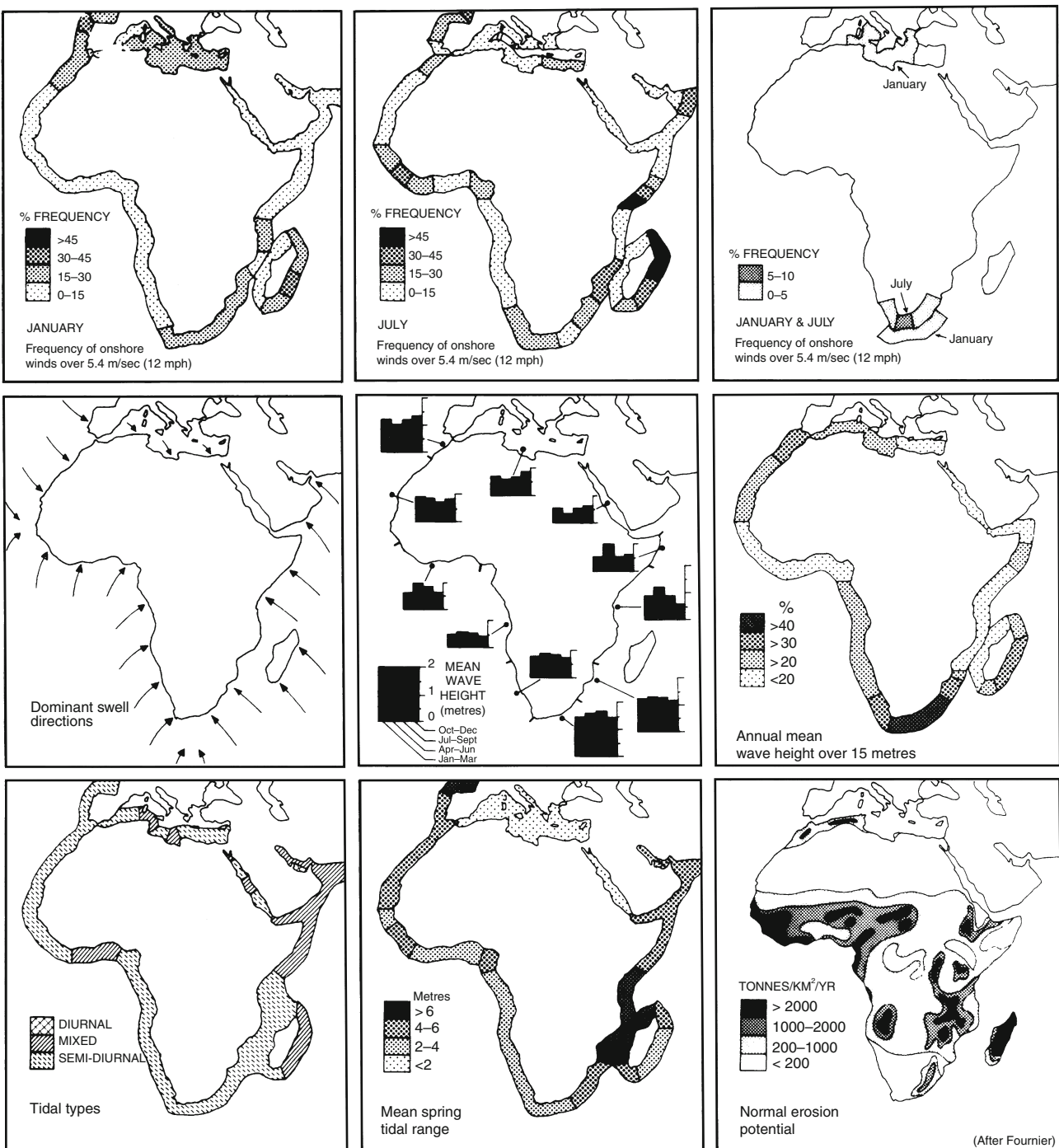
Climatic Factors

Directly or indirectly, climate affects most processes shaping the coast. As elsewhere, Africa's coastal climates reflect the impact of seasonal changes in Earth-Sun relations on the Intertropical Convergence Zone (ITCZ), air masses, and wind regimes. During the southern summer, the ITCZ shifts southward with the Sun to a zone running from the Guinea coast to Madagascar. Excepting the influx of cool moist maritime polar air masses from the Atlantic into the Mediterranean, most of north Africa is covered by warm dry outflowing air associated with continental tropical air masses over the Sahara and southwest Asia, promoting the dust-laden Harmattan over the Guinea coast and the northeast monsoon along the east coast, respectively. During this, the *jilaal* season in Somalia, warm northeasterly winds up to 30 km hr⁻¹ drive ocean waters and aeolian sands southward along the coast (Orme 1985). Farther south, warm rainy conditions and onshore winds linked to inflowing marine air or local thermal convection cells prevail.

During the northern summer, the ITCZ shifts northward to the Sahel. Hot wet conditions prevail over central Africa and the Guinea coast, Saharan aridity reaches the Mediterranean, and, although much of southern Africa is dry, cool maritime air masses and storms from the Southern Ocean impact the southern coast. This, the *hagaa* season in Somalia, sees southwest monsoon winds with mean velocities over 40 km hr move aeolian sand and surface waters northward along the east coast, while the hot desiccating dust-laden *khariif* wind invades the Gulf of Aden from northern Somalia.

Oceanic Factors

Excepting the Mediterranean and northwest coasts where northwesterly swells predominate, the African coast is dominated by southerly swells generated by storms in the Southern Ocean (Fig. 2). From Cape Vert to Cape Agulhas, these swells are mostly southwesterly, decreasing in height northward and from southern winter into southern summer. Along the Guinea coast, these swells are reinforced by southwesterly monsoon winds during the northern summer. Because of the orientation of Africa's west coast, the southwesterly swells generate mainly east-flowing longshore currents in the Gulf of Guinea and mainly north-flowing longshore currents along the west-facing coast farther south. This nearshore wave-driven circulation is reinforced offshore by the Guinea and Benguela Currents. Exceptions to these patterns occur in the



Africa, Coastal Geomorphology, Fig. 2 Selected processes affecting the African coast. (In part after Hayden et al. 1973; Davies 1977; from Orme 1982, with permission of Kluwer Academic Publishers)

lee of major promontories such as the Niger delta. Throughout the Atlantic coast, wave erosion and longshore sediment transport are strongest and most frequent when swells are highest – from December to March north of Cape Vert, and from June to September farther south (Davies 1977).

Along the east coast, from Cape Agulhas to Ras Asir, most swells also originate from the southwest but are strongly

refracted to reach the coast from the southeast (Fig. 2). This circulation is further complicated by reversing monsoonal winds and currents off Somalia, by westward-moving cyclones and tropical easterlies off Madagascar, and by the south-flowing Agulhas Current farther south. Under the influence of the southwest monsoon, for example, the Somali Current reaches surface velocities of 3.7 m s^{-1} and the

uppermost 200 m of its water column transports $60 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ northward. Along the Kwa-Zulu coast, littoral transport of coarser terrigenous sediment is commonly northward within the surf zone, but finer sediment that is flushed farther seaward is carried south with the Agulhas Current (Orme 1973). Onshore winds are stronger and more frequent during the southern winter, when winds exceeding 60 km hr^{-1} may drive powerful surf against the continent's southern tip.

Highest wave energy occurs around the southern tip, from Cape Point to Durnford Point, and moderately high wave energy prevails from the Orange to Limpopo river mouths, along Madagascar's east coast, and along Morocco's Atlantic coast (Fig. 2). Lower wave energy occurs in the Red Sea, the Mozambique Channel, the Gulf of Guinea, and eastern Mediterranean.

Semidiurnal tides predominate around Africa, with mixed regimes confined to the central Mediterranean, Red Sea, Guinea, Somali, and eastern Madagascar coasts (Fig. 2). Mean tidal range is less than 2 m throughout the Red Sea and Mediterranean, over 6 m along the Mozambique Channel, and 2–6 m elsewhere.

Hydrologic Factors

The continent's hydrologic regimes are important because they deliver freshwater and sediment to the coast, which in turn influence coastal ecology and sediment budgets (Orme 1996; Walling 1996). Sediment delivery is highest in areas of rugged relief and/or high rainfall, at least seasonally (Fig. 2). Thus, the many small rivers that drain steeply from the mountains of Algeria and Morocco to the Mediterranean may have only modest sediment loads but large sediment yields per unit area of their basins, often $1,000 \text{ tonnes (t) km}^{-2} \text{ yr}^{-1}$ (Milliman and Syvitski 1992). Sediment yields may also be high where open forest or savanna vegetation has been modified or destroyed by human activity, for example, along the coast of Guinea, east Africa, and Madagascar. Sediment yields are lower where luxuriant vegetation inhibits surface erosion or where rivers have long run-out reaches across coastal lowlands, and lowest where heavy rainfall is lacking, notably in desert areas.

Most larger rivers have lower sediment yields because they often cross different climatic regimes and broad floodplains, both of which afford opportunities for floodplain storage. However, because their drainage basins are large, they may deliver substantial sediment loads to the sea annually. Thus, the Congo and Limpopo, respectively, carry average loads of $43 \times 10^6 \text{ t}$ and $33 \times 10^6 \text{ t}$ to the sea annually (Milliman and Syvitski 1992). Of particular importance to the present coast are the reduced sediment loads in rivers that have been dammed upstream. Thus the Orange, which formerly transported $89 \times 10^6 \text{ t}$ of sediment annually, now moves only $17 \times 10^6 \text{ t}$ to the sea. The Zambezi, which once transported

$48 \times 10^6 \text{ t}$ annually, now carries only $20 \times 10^6 \text{ t}$, largely owing to construction of the Cabora Bassa Dam upstream. Most dramatically, annual sediment loads from the Nile and Volta, formerly $120 \times 10^6 \text{ t}$ and $19 \times 10^6 \text{ t}$, respectively, have diminished to virtually zero as a result of construction of the Aswan High Dam (completed 1964, capacity $162 \times 10^6 \text{ m}^3$) and earlier structures on the Nile, and of the Akosombo Dam (1961) on the Volta. These structures have major consequences for coastal sediment budgets.

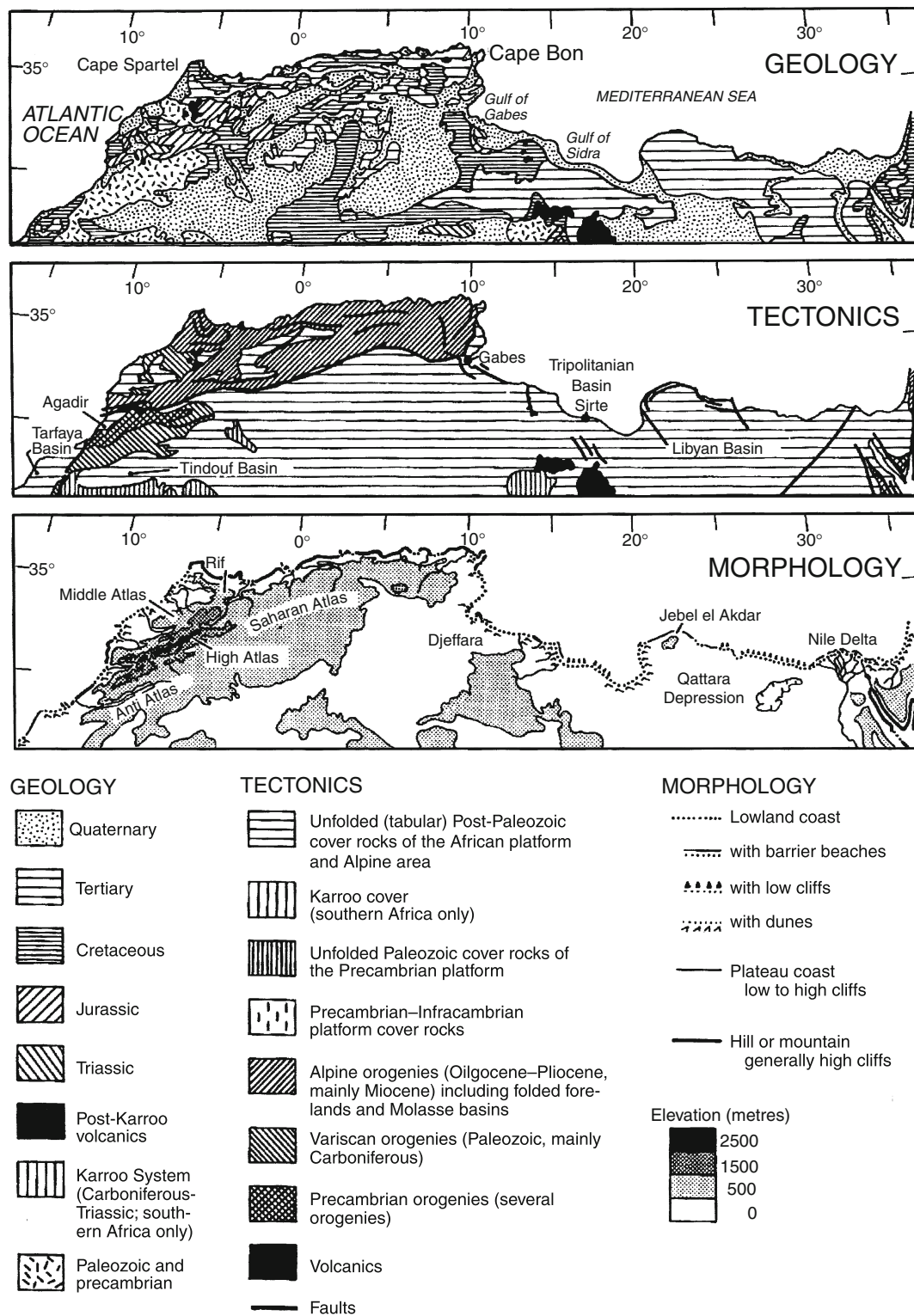
Despite high sediment yields from seasonally wet areas and high sediment loads from many larger undammed rivers, Africa's vast internal drainages and the relatively aridity and low gradient of most external basins restrict sediment yields to the coast (Walling 1996). Thus, compared with other continents, direct wave attack on coastal cliffs and carbonate reefs is a more important source of sediment. Further, the wide extent of seasonally dry climates means that, for much of the year, ocean processes predominate over fluvial processes around river mouths. Thus, many mouths are closed during drought and partially blocked by coastal barriers at other times.

Human Factors

Although the African coast has suffered less from human activity than that of Europe or North America, some historic impacts are noteworthy and the rate of human-induced change has accelerated in recent decades. Apart from the construction of major dams inland, noted above, land-use practices inland have often led to extensive soil erosion and increased sediment yields (Stocking 1996). In Kwa-Zulu, for example, recent sediment yields are two to three times the average for the past 5,000 years, owing in large measure to overgrazing and accelerated erosion in coastal basins (Orme 1973). Along many parts of the African coast, increased sediment yields have clogged estuaries while sediment plumes have been worked onshore to form new beach-dune ridges. Conversely, along Somalia's east coast, Pleistocene dunes have been reactivated and gullied by overgrazing of vegetation by cattle, goats, and camels, aggravated by tillage practices of an increasingly desperate refugee population (Orme 1985). Elsewhere, beaches and dunes have been mined for diamonds and heavy minerals, respectively, while beaches and wetlands near major ports have been consumed by harbor and urban development. Pollution and recreation are also beginning seriously to impact coral reefs in east Africa.

The Mediterranean Coast

Africa's 5,000-km long north coast is divisible into two contrasting segments (Fig. 3). The 3,300 km eastern segment from Sinai to the Gulf of Gabes features low cliffs of little deformed Cenozoic sedimentary rocks interspersed with



Africa, Coastal Geomorphology, Fig. 3 The Mediterranean coast: geology, tectonics, morphology. (From Orme 1982, with permission of Kluwer Academic Publishers)

barrier beaches, lagoons, and dunes. The 1,700 km western segment from Cape Bon to Cape Spartel reflects Alpine tectonism in its rugged coastal mountains and deep coastal

basins. Throughout this coast, prevailing northwest winds generate northwesterly swells and wind waves, strongest in winter, which in turn promote net littoral drift eastward.

With a mean tidal range of 0.2–0.5 m, but 1.8 m in the Gulf of Gabes, tidal forcing is weak. Wave energy, littoral drift, annual rainfall, and fluvial sediment inputs, diminish eastward.

The 12,500 km² Nile delta fronts 280 km of the Egyptian coast. Although fluvial sediment rich in Fe and Al from the Ethiopian Highlands may still be found in beaches east of the delta, water and sediment delivery by the Nile have been much reduced by irrigation diversions and dams built since the 1880s. Of the six or more distributaries used by the Nile some 2,000 years ago, only the Damietta and Rosetta branches still function. Further, because longshore currents continue to move sediment eastward, the delta's seacoast is suffering severe erosion, at rates up to 60 m yr⁻¹, leaving high-density magnetite, ilmenite, zircon, and garnet as black-sand placers while moving lighter minerals downdrift (El-Ashry 1985; Frihy and Komar 1991). A 40 km deep littoral zone, the *barari* (barren) fronts the delta, comprising barrier beaches and back-barrier marshes, salt flats, and lagoons of which the largest are Manzala (1,450 km²) and Burullus (560 km²). Over the past 8,000 years, delta subsidence has ranged from 1 mm yr⁻¹ Alexandria to 5 mm yr⁻¹ near Port Said (Stanley 1990). During historic time, while delta subsidence was offset by sedimentation, coastline location did not change much but, with sediment delivery now curtailed, several areas have subsided to below sea level. Farther east, the Sinai coast is a broad sandy plain with bold dunes and a 100 km long, 500 m wide barrier beach fronting Bardawil Lagoon along whose shallow hypersaline margins gypsum and halite readily precipitate. The ephemeral Wadi el Arish is the only sizable stream course to reach this coast.

Westward from the delta, dark Nile sediment is replaced by carbonate grains from local sources. Several beach-dune ridges backed by elongate sabkhas parallel the coast, reflecting episodic barrier formation during late Quaternary time. In Cyrenaica, the Jebel el Akdar limestone plateau generates higher rainfall and greater surface runoff, and reaches the coast in rocky cliffs and narrow terraces that leave little room for coastal plain. From Benghazi to Cape Bon, the coast is again dominated by barrier beaches, dunes, sabkhas, and salt marshes. The 150 km² Sabkha el Melah in southern Tunisia exemplifies Holocene evaporite sedimentation in which marine detrital limestones, euxinic beds, magnesium carbonates, gypsum, polyhalite, and halite have successively accumulated over the past 6,000 years (Busson and Perthuisot 1977).

From northern Tunisia westward, coastal character changes as mid-Cenozoic (Atlas) and neotectonic structures form prominent cliffs, marine terraces, and subsiding basins, and as the seasonally wetter, stormier climate mobilizes coastal and fluvial sediment. Quaternary tectonic and eustatic forces are reflected in suites of deformed marine terraces rising to over 200 m above sea level near Monastir and

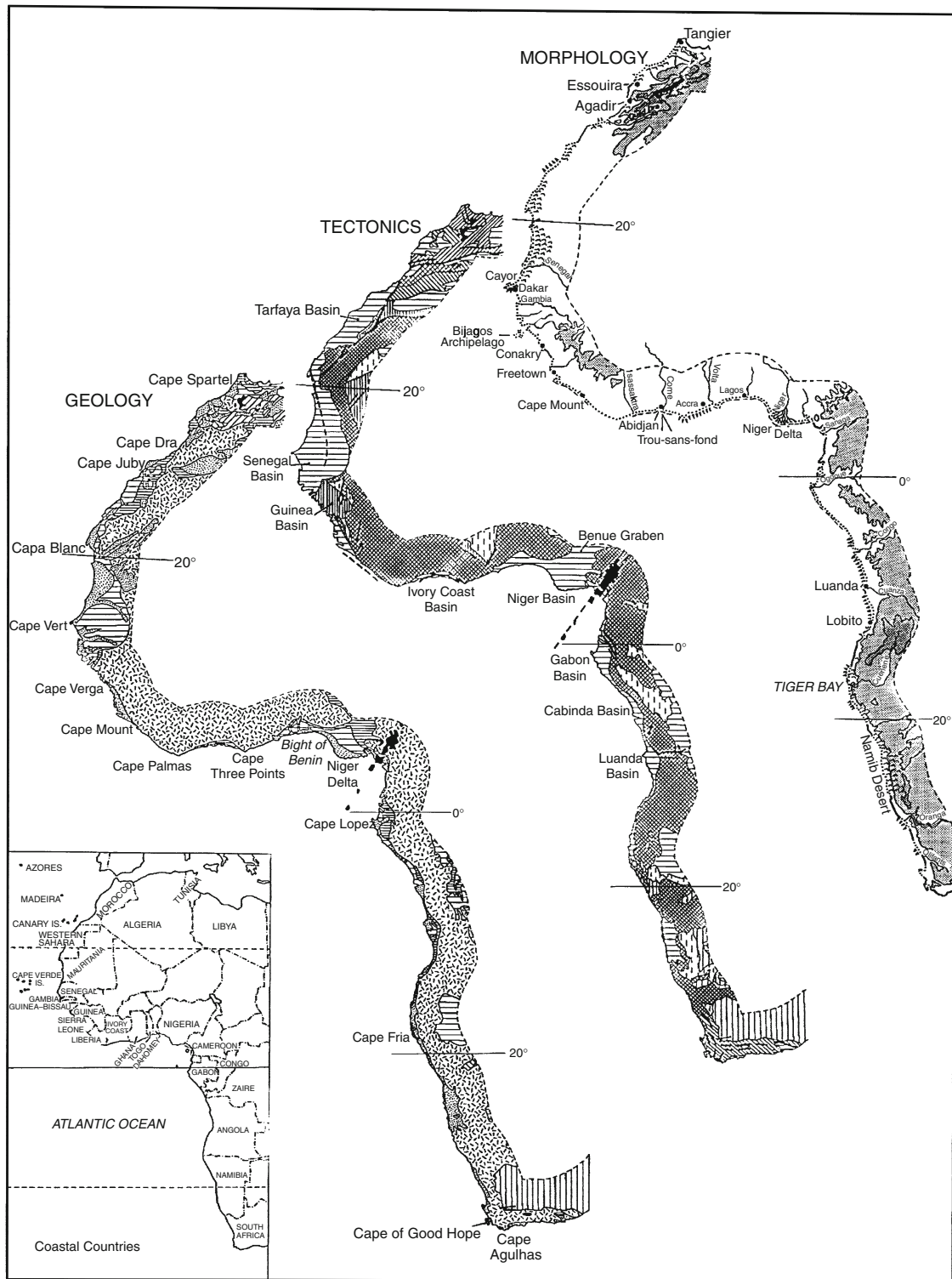
Bizerte, and in 400 m of littoral sediments beneath the subsiding Medjerda delta whose outlet has changed frequently in response to historic floods (Paskoff 1978). Farther west, along the Algerian Tell and Moroccan Rif, structural basins flooded with Quaternary sediment open through swampy lowlands into Skikda, Bejaia, and Arzew Bays. Wadis discharge abundant sediment to the coast in winter, to be partly reworked onshore in aeolian dunes. The Isser and Soummam rivers have a mean yield of 6.1×10^6 and 4.2×10^6 t yr⁻¹, respectively, while sediments carried by the Moulouya from the High Atlas contribute to impressive shingle beaches east of Nador, Morocco. Alpine massifs up to 1,000 m above sea level reach the shore in active sea cliffs involving crystalline basement, Mesozoic limestones and sandstones, and Cenozoic molasse and volcanic rocks, and then descend across a narrow shelf down the 2,000 m Habibas submarine escarpment.

The Atlantic Coast

The Atlantic coast of Africa may be divided into three segments: the semiarid to arid coast from Cape Spartel to Cape Vert characterized by the cool Canary Current, northwesterly swells and south-flowing littoral drift; the humid tropical coast from Cape Vert to beyond the Congo River, dominated by the Guinea Current, southerly swells and east-flowing littoral drift; and the sub-humid to arid coast from Angola south to the Cape characterized by the cool Benguela Current, southwesterly swells and strong north-flowing littoral drift (Fig. 4).

From Cape Spartel to Cape Dra, the Moroccan coast is underlain by Hercynian and mid-Cenozoic structures that have been planed off by later marine erosion, forming the Pliocene Moghrebian *rasa* and sequences of Pleistocene marine terraces, all since deformed. The continental margin offshore has seen episodic sedimentation and subsidence since the break-up of Gondwana. Winter floods transport large sediment loads from the Atlas Mountains down the Sebou, Rharrb, Sous, and other rivers to feed beach-dune complexes up to 80 m high which, shaped by northwest winds in winter and northeast trade winds in summer, enclose many lagoons and sabkhas.

From Cape Dra to Cape Vert, the coast overlies the Tarfaya and Senegal structural basins whose post-Gondwana cover rocks, planed by episodic marine transgressions, form low coastal cliffs and a relatively smooth continental shelf. Saharan sands blanket this coast along a broad front. Extensive late Cenozoic dune systems show that aeolian processes have long dominated the region. Saharan sands are driven seawards by the *irifi*, a hot easterly wind, to form the great red *draas* and superimposed barchans that trend obliquely across the Mauritanian coast between Cape Blanc and the Senegal River,



Africa, Coastal Geomorphology, Fig. 4 The Atlantic coast: geology, tectonics, morphology (see Fig. 3 for key). (From Orme 1982, with permission of Kluwer Academic Publishers)

notably in the Azefal and Akchar ergs (Kocurek et al. 1991). The Sahara now contributes $13 \times 10^6 \text{ m}^3$ of quartz sand to the continental shelf annually, but this amount may have been much larger in the past (Sarnthein and Walger 1974).

Late Pleistocene aeolianites occur to at least -50 m off Cape Vert while Saharan dust is found in deep ocean sediments far beyond the shelf. Coastal sands are reworked onshore by the *alisio*, a persistent northerly wind, to form

dunes ridges up to 2,000 m wide and 40 m high, which in turn trap sabkhas (*niayes*) seaward of Saharan draas in Mauritania and Senegal. Lacking rainfall and a mountain backdrop, stream flows and fluvial sediment discharge are now negligible, except during rare rain storms, along the coast between Cape Dra and the Senegal River, but relict fluvial sediments on the continental shelf, including detrital phosphatic placers derived from onshore phosphorites, show that fluvial processes were more important during Pleistocene time.

The 4,250 km² Senegal delta forms the outlet for a 196×10^3 km² drainage basin whose mostly summer rains average 1,381 mm annually. Despite significant evaporative loss, annual discharge through the delta averages $870 \text{ m}^3 \text{ s}^{-1}$, ranging from negligible during prolonged drought to $3,700 \text{ m}^3 \text{ s}^{-1}$ during floods (Guilcher 1985). The delta formerly entered a bay created by the Flandrian (Nouakchottian) transgression but, as the delta plain accreted, marine processes deflected the river mouth 120 km southward behind a massive barrier spit. The distal end of this barrier, the unstable 30 km long Langue de Barbarie, has been breached 25 times by storm waves or river floods since 1850. Although spring tidal range is only 1.2 m, the low gradient of the Senegal River allows tidal influences to penetrate over 400 km upstream. Southward littoral drift along the open coast ranges from 0.5×10^6 to $1.0 \times 10^6 \text{ t yr}^{-1}$, part of which is lost down Kayor submarine canyon off Cape Vert.

From Cape Vert to the Niger delta, cliffed coasts formed in Precambrian basement and Phanerozoic sedimentary rocks alternate with lowlands underlain by Cretaceous and Cenozoic sediments and fronted by beach ridges and mangrove swamps. Paleozoic rocks emerge along the south coast of Guinea-Bissau and at Cape Verga; basic intrusives form the Kaloum and Freetown peninsulas; Precambrian granite and gneiss underlie the coasts of Liberia and western Ivory Coast, and reappear with Paleozoic sandstone in low cliffs from Cape Three Points nearly to the Volta delta. In contrast, the subsiding and downfaulted Ivory Coast Basin and the larger Niger Basin are underlain by great thicknesses of Cretaceous and Cenozoic sediment overlain by coastal barriers, lagoons, and swamps.

The southwest-facing coast between Cape Vert and Cape Palmas, tropical rains of 2,000–3,000 mm yr⁻¹, enhanced by inland relief and southwest winds drawn onshore during the northern summer, is reflected in abundant discharge of water and sediment from short coastal rivers. Broad deltaic estuaries, such as those of the Casamance, Geba, Konkoure, and Scarcies rivers, reach the sea through intricate networks of tidal creeks, mangrove swamps, and supratidal mudflats or *tannes*. While most estuaries remain open throughout the year, barrier beaches locally front coastal swamps and lagoons in southern Sierra Leone and Liberia. Modern beaches and Holocene beach ridges farther inland, notably on Sherbro Island and the Turner Peninsula testify to

abundant sediment supplies, powerful onshore swells, and seasonally reversing littoral drift. Offshore, the continental shelf ranges from 200 km wide on the shallow Great Geba Flat around the Bijagos Archipelago to 20 km wide at Cape Palmas. This shelf contains many submerged river channels, deltas, coral reefs, and shoreline sequences to depths of 90 m, all evidence of low Pleistocene sea levels (McMaster et al. 1971). Fine sediment is readily swept southward across shelf by the surface Canaries Current and tidal ebb currents, and then returned northward by strong bottom countercurrents.

East from Cape Palmas, the south-facing Guinea coast causes southwesterly swells to promote east-flowing littoral drift, aided for fine sediment by the Guinea Current. Littoral drift reflects sediment availability. Thus, resistant Precambrian rocks east of Cape Palmas and again east of Cape Three Points restrict littoral drift to around $0.2 \times 10^6 \text{ m}^3 \text{ yr}^{-1}$ but, where Cenozoic basinal and recent fluvial sediments reach the coast east of Fresco and again east of Accra, littoral drift reaches $0.8 \times 10^6 \text{ m}^3 \text{ yr}^{-1}$, forming massive barrier beaches across the mouths of many rivers. Where sediment supplies have been curtailed, notably by construction of the Akosombo Dam (1961) across the Volta River, or where littoral drift is interrupted, for example, by jetties off the Vridi Canal and Lagos Harbor, downdrift beach erosion occurs. Offshore, the continental shelf is relatively narrow (10–70 km) and featureless but exceptions occur in the Trou-sans-Fond, a massive submarine canyon that heads at –10 m off Abidjan but descends to 2,000 m and traps 4,000,000 m³ yr⁻¹ of littoral drift; the large Volta submarine delta; and submerged beach–aeolianite ridges that indicate a late Pleistocene regression to the shelf edge at –120 m.

The Niger delta is the outlet for a 1.11×10^6 km² drainage basin whose main artery, the 4,460-km Niger River rises in the Guinea Mountains only 250 km from the Atlantic. Following mid-Cretaceous separation of South America from Africa in this locality, and subsequent crustal adjustments and sea-level changes, the delta began forming in Eocene time when the coastline lay near Onitsha, 250 km inland from the present coast. The basin began acquiring its present form during a later Pleistocene pluvial stage when the NE-flowing Soudanese Niger overflowed into the SE-flowing Nigerian Niger. The Flandrian transgression later drowned the outer 50 km of the delta's late Pleistocene distributary system. The delta now comprises 500,000 km³ of Eocene–Holocene sediment, up to 8 km thick, and covers 28,827 km² of which two-thirds lie above sea level (Hospers 1971). Present annual discharge of freshwater into the delta is about $200 \times 10^9 \text{ m}^3$ and of sediment about $18 \times 10^6 \text{ m}^3$, mostly as silt and clay that radiate unequally through a dozen major distributaries into the Gulf of Guinea. Depositional environments and sedimentary facies are distributed concentrically within the delta (Allen 1970). The deltaic plain comprises upper and lower floodplains, tidal

creeks, and mangrove swamps fronted by barrier beaches. The submerged delta comprises river-mouth bars, delta-front ramps, pro-delta slope, open shelf, and relict Pleistocene littoral deposits. Prevailing southwesterly swells strike symmetrically against the nose of the delta, promoting divergent littoral drifts east toward the Cross River and west toward the Benin River. Submarine canyons channel about $1 \times 10^6 \text{ m}^3$ of sediment from these drifts into fans beyond the delta foot. The delta mass is broken by arcuate gravity faults downthrown toward the gulf and also affected by compaction and subsidence of its constituent sediments, aggravated in recent years by extensive subsurface oil and gas withdrawal. Subsidence and mangrove clearance have in turn increased erosion to 40 m yr^{-1} along the exposed muddy coast northwest of the Benin River (Ebisemiju 1987).

The west-facing coast of Africa south of the Mount Cameroun volcanic pile (4,070 m) comprises several post-Gondwanan basins containing Cretaceous and Cenozoic sedimentary and volcanic rocks separated by Precambrian basement, all variably mantled by recent marine, fluvial, and aeolian deposits. The Niger Basin extends through the Cameroun volcanics into the swampy and deltaic Sanaga coastal plain. The Gabon Basin is fronted by barrier-lagoon complexes and north-oriented spits terminating at Cape Lopez. In contrast, the Cabinda Basin has active sea cliffs north and south of the Congo estuary. South of Precambrian outcrops near Ambriz, the Luanda Basin is fronted in part by 100–150 m sea cliffs cut in Cretaceous and Cenozoic rocks and in part by *restingas*, north-oriented barrier spits enclosing bays and lagoons, notably the Luanda spit (12 km long), Lobito spit (9 km long), and the spit enclosing Tiger Bay (37 km long, before breaching in 1962; Guilcher et al. 1974). These barrier spits form in response to powerful southwesterly swells and predominant north-flowing littoral drift. However, littoral processes are incapable of closing the mouth of the Congo River where powerful discharge annually jets up to $50 \times 10^6 \text{ m}^3$ of sediment seaward into a submarine canyon, denying most of this load to the littoral drift. This submarine canyon descends to 2,700 m before spilling across an abyssal fan with leveed distributaries to –4,900 m.

Farther south, although Precambrian basement and Cretaceous basalts form cliffs locally, the coast from southern Angola through Namibia to northern Cape Province is largely mantled by relict and active coastal dunes, including the Kunene Sand Sea and the 400 km long Namib Sand Sea (Lancaster 1996), and by relict alluvial fans. The perennial Kunene River along Namibia's northern boundary sheds much terrigenous debris onto the continental shelf. Similarly, the Orange River along the country's southern border, though blocked by sand during the dry season, has long been the source of the beach and dune sand that moves north with the dominant southwesterly winds and swells, prograding the coast and deflecting the mouths of the Swakop and Kuiseb

ivers. The diamantiferous marine-terrace deposits of the Sperrgebiet, ranging from 90 m above to –60 m below sea level, originate from switching late Cenozoic outlets of the Orange River. The shelf here varies from <30 km wide off rocky coasts to >160 km wide where terrigenous Orange River sediments have built a broad submarine delta. Where aeolian sand seas block lesser streams, however, shelf sediments are composed mainly of diatomaceous muds, foraminiferiferous sands, and skeletal clasts.

The Southern Tip

Between the Olifants and Tugela rivers, strong southerly swells and winter storms have fashioned a rugged cliffed coast interspersed with sandy bays from Precambrian and Paleozoic rocks of the Cape Fold Belt (280–230 Ma) and the thick formations of the Karroo Supergroup. Thus, massive sandstone-on-granite cliffs of the Cape Peninsula are separated from fold mountains farther east by the broad dune-mantled Cape Flats. From Cape Agulhas to Algoa Bay, the coast cuts across the easterly strike of the Cape Fold Belt, causing cliffed anticlinal ridges to alternate with synclinal corridors fronted by barrier beaches and extensive dunefields. In the Alexandria Basin, a transverse dunefield is moving inland along a 50 km wide front driven by WSW winds off Algoa Bay. Sand enters this dunefield at a rate of $375,000 \text{ m}^3 \text{ yr}^{-1}$ and leaves at $45,000 \text{ m}^3 \text{ yr}^{-1}$ for a net gain of $330,000 \text{ m}^3 \text{ yr}^{-1}$ (Illenberger and Rust 1988). Farther northeast, gently tilted rocks of the Karroo Supergroup are exposed in majestic cliffs fronting coastal plateaus.

Quaternary marine terraces rise at intervals to over 60 m along this coast, while beachrock, related to both high interglacial and Holocene sea levels, forms a broken pavement along much of the immediate shore. Beachrock typically consists of quartz sand and shell fragments cemented by micrite or aragonite overlain by laminated calcrete. Relict beachrock and aeolianite are often exposed along the coast and also form submerged ridges offshore (Fig. 5). Off Cape Agulhas, the continental shelf forms the 240 km wide Agulhas Bank, relatively featureless except for the shelf edge between –110 and 380 m which is cut by several canyons. The bank is formed from over 6,000 m of post-Gondwanan sediments lying on basement rocks and, during low Pleistocene sea levels, was exposed over $22,500 \text{ km}^2$ (Dingle 1973).

Compared to Africa's great rivers, the Gourits, Sundays, Great Kei, Mtamvuna, and other rivers are relatively short but, in descending swiftly from the Cape fold mountains and the Great Escarpment, they transport large sediment loads seasonally (Fig. 6). In the Mediterranean-type climate of Cape Province, maximum flows occur in the southern winter (May–August) but farther north, in Natal-Kwa-Zulu, summer



Africa, Coastal Geomorphology, Fig. 5 Pleistocene raised beach deposit overlying Pleistocene aeolianite, Durban Bluff, South Africa. (From Orme 1982, with permission of Kluwer Academic Publishers)

(December–March) thunderstorms are more important. Rivers draining the bold Drakensberg (3,482 m) enter the sea through Pleistocene channels excavated to -100 m below sea level, now partly filled with Holocene estuarine sediment (Orme 1974, 1976; Cooper 1993). Over the past 200 years, sediment yields have been accelerated by soil erosion but diminished by dam construction.

The Indian Ocean Coast

Africa's mostly low-lying east coast, from the Tugela estuary to Ras Azir, straddles post-Gondwanan coastal basins filled with several thousand meters of Mesozoic and Cenozoic sediments the 4,000 m deep Mozambique Basin in the south, characterized by north–south faults at the south end of the East African Rift system, and the 3,000–8,000 m deep Dar es Salaam–Kenya–Somali Basin continuum farther north (Fig. 7).

The onshore portion of the Mozambique Basin forms a coastal plain 300 km wide in central Mozambique but pinches out southward near the Tugela estuary and northward at Mocambo Bay. Its 2,100 km-long shoreline is characterized by long barrier beaches topped by parabolic and lobate dunes up to 180 m high, behind which lie shallow lagoons,

extensive swamps, old coastal barriers, and extensive aeolian cover sands (Orme 1973). In Kwa-Zulu, the massive coastal barrier, although veneered with Holocene sands, is superimposed upon remnants of a major late Pleistocene barrier-lagoon complex, the Port Durnford Formation, whose near-shore, beach, and swamp facies were probably deposited during the last interglacial around 125 ka (Fig. 8; Hobday and Orme 1975). A similar scenario seems likely along much of the Mozambique coast. The older coastal barrier was subsequently breached by rivers draining to lowered late Pleistocene sea levels and then invaded by the Flandrian transgression, for example, forming Lake St. Lucia in Kwa-Zulu and Delagoa Bay behind Inhaca Island in Mozambique (Fig. 9). Late Pleistocene estuaries excavated to more than -55 m below sea level are now choked by complex sequences of Holocene marine, estuarine, and fluvial sediment (Orme 1974, 1976). Offshore, beachrock and aeolianites towards the edge of the continental shelf testify to the magnitude of late Pleistocene drawdown.

Shorezone dynamics along the coast of the Mozambique Basin are conditioned by prevailing northeasterly and southwesterly winds, by powerful swells from the Southern Ocean, by longshore currents up to 2 ms^{-1} , by reversing coastal flows related to the Agulhas Current that parallel the shore with surface velocities up to 0.8 m s^{-1} , and by seasonally high



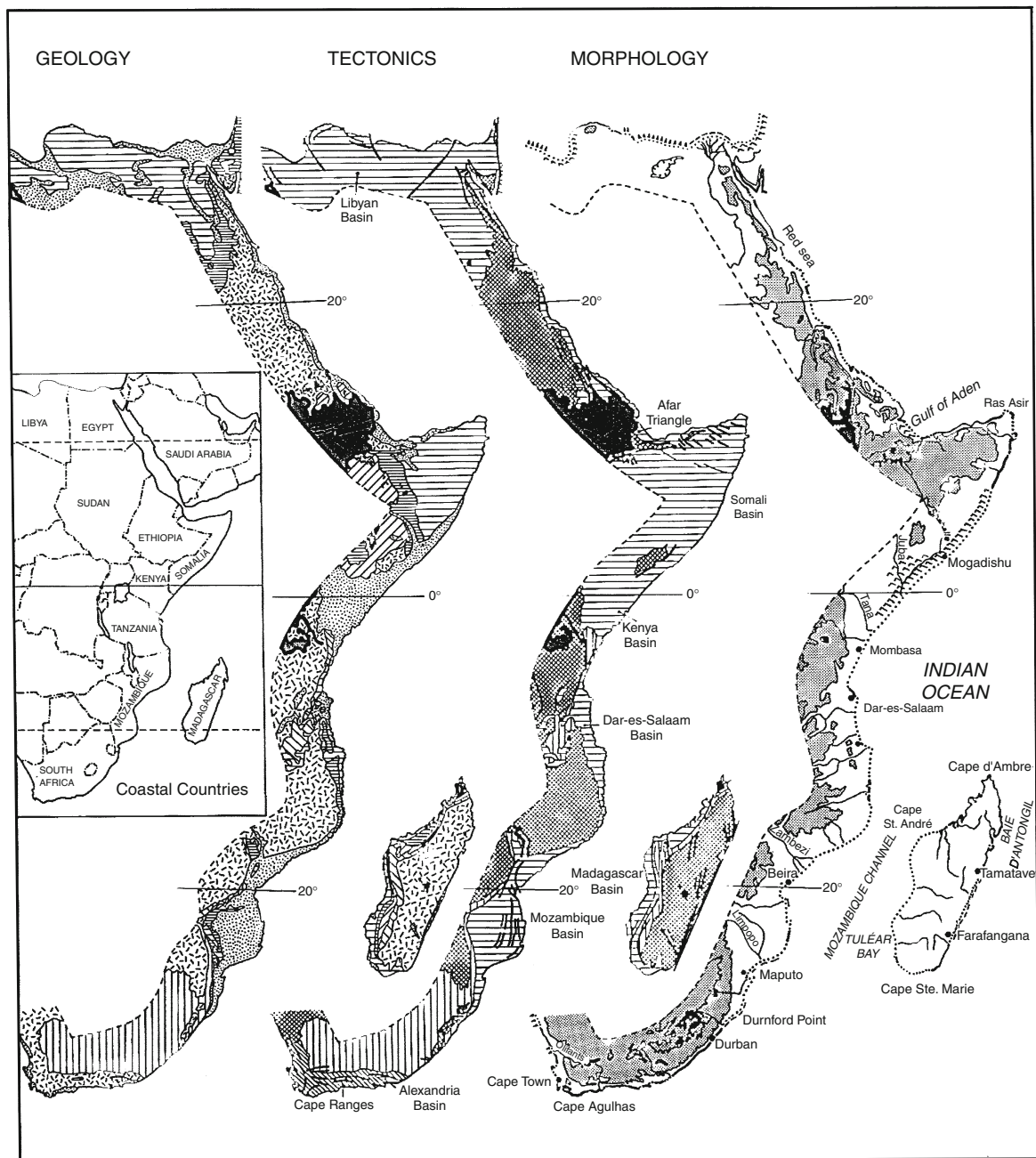
Africa, Coastal Geomorphology, Fig. 6 Mtamvuna river mouth and wet (summer) season sediment plume, Transkei-Kwa-Zulu border. (Photo: A.R. Orme)

discharges from coastal drainage basins. The Tugela River brings 10.5×10^6 t of sediment to the coast annually and its sediment plume, like those of neighboring rivers, may be traced far out into the Agulhas Current. Owing to unwise land use and increased erosion inland, recent sediment delivery rates to the coast have greatly exceeded prehistoric rates, leading, for example, to accelerated beach-ridge and dune formation north of the Tugela estuary (Fig. 10). The 7,150 km² Zambezi delta forms the outlet for a 1.33×10^6 km² savanna drainage basin and a 3,540 km mainstream. Discharge through the delta averages $7,000 \text{ m}^3 \text{ s}^{-1}$, and sediment that is not trapped within coastal swamps and lagoons, or transported coastwise as littoral drift, is flushed over the continental shelf onto the Zambezi cone. North of the Mozambique Basin, from Mocambo Bay to Kisware Bay, faulted Cretaceous and Cenozoic rocks underlie an embayed cliffed coast fronted by raised coral bluffs and fringing reefs. The continental shelf here is often only a few hundred meters wide, plunging to $-2,500$ m within 10 km of the shore.

The coastal zone of Tanzania and Kenya comprises numerous, deformed and faulted Pleistocene marine terraces supporting relict barrier beaches and coral reefs, as well as active Holocene barrier beaches, mangrove swamps, offshore bars, and coral reefs. Three late Quaternary beach

ridges surmount the lowest terrace between Dar es Salaam and Tanga, reflecting an abundant sediment supply worked onshore by strong swells, modified by wind action, and then stabilized by vegetal growth (Alexander 1968). Raised coral reefs rise to 140 m along the Kenya coast but their margins are often obscured where cliffing was inhibited by the presence of wide lagoons and coral growth during marine regressions (Hori 1970). The offshore islands of Zanzibar (1,660 km²), Pemba (984 km²), and Mafia are also relict coralline structures. All three islands are defined eastward by massive faults and Pemba is separated from the mainland by a 700 m deep graben. The Rufiji River, draining a 0.18×10^6 km² drainage basin, transports an average sediment load of 17×10^6 t to the sea annually (Milliman and Syvitski 1992).

From the Tana River for 1,600 km northward to Gifile in Somalia, although fringing reefs persist, the coast is dominated by massive aeolian deposition promoted by semiarid conditions, strong monsoon winds, and episodic supplies of fluvial sediment. The Tana delta is blocked seaward by dunes, while complex Quaternary dune complexes up to 200 m high have deflected the Shebele River for 400 km parallel with the present coast, allowing it to reach the sea through the Juba estuary only after heavy rains (Orme 1985). Paleodunes along this coast have been widely destabilized and gullied as a result



Africa, Coastal Geomorphology, Fig. 7 The Indian Ocean and Red Sea coasts: geology, tectonics, morphology (see Fig. 3 for key. (From Orme 1982, with permission of Kluwer Academic Publishers)

of overgrazing and poor tillage practices, most recently by a population swelled by refugees from strife farther inland. Dune stabilization efforts have so far met with only limited success. From Gifile to Ras Asir, Cenozoic carbonate rocks form rocky marine terraces and sea cliffs 100–300 m high. The 1,100 km long *guban* (sunburnt plain) coast of the Gulf of Aden is partly cliffed, partly backed by alluvial plains, sabkhas, and marine terraces. Thick Plio-Pleistocene marine sequences and coral reefs raised 180–280 m above sea level

around Berbera emphasize the recent tectonic emergence of this coastal zone. No perennial rivers reach Africa's east coast north of the Shebele River.

Madagascar

The coast of Madagascar reflects the structural and topographic asymmetry of this large island. The linear, fault-controlled east



Africa, Coastal Geomorphology, Fig. 8 Port Durnford Formation, Kwa-Zulu, showing basal fossiliferous interglacial swamp peat overlain by transgressive beach sands, buried beneath weathered aeolianites. (Photo: A.R. Orme)

coast is broken by only one major embayment, the Baie d'Antongil graben (Fig. 9). Basement rocks plunge from 2,000 m highlands onto a narrow coastal plain and shelf underlain by Cretaceous sediments. Persistent southeast trade winds, storms in the Southern Ocean, high average wind velocities, and west-moving cyclones from December to March all promote strong wave action. In response, barrier beaches and shallow linear lagoons have formed along 800 km of coast from Fénériver southward to Port Dauphin. These lagoons, separated from one another by low rises or *pangalanes*, are used by coastwise traffic.

In contrast, the west coast fronts a broad alluvial lowland underlain by thick post-Gondwanan sediments and the shelf widens to 150 km off Cap St. André and on the Leven Bank. Along the northwest coast between Cap St. André and rugged Cap d'Ambre at the island's northern tip, the Flandrian transgression breached a barrier system to produce complex rias, partly drowned cuestas, deltas, mangrove swamps, and islands. The Isles Glorieuses northwest of Cap d'Ambre, contain coral reefs, now 3 m above sea level, which formed around 150 ka. The semiarid southwest coast is dominated by active parabolic dunes and oxidized paleodunes shaped by prevailing southerly winds. The shelf is narrower here and in Tuléar Bay is notched by massive submarine canyons, that off the Onilahy River descending to −2,600 m. Coral reefs of

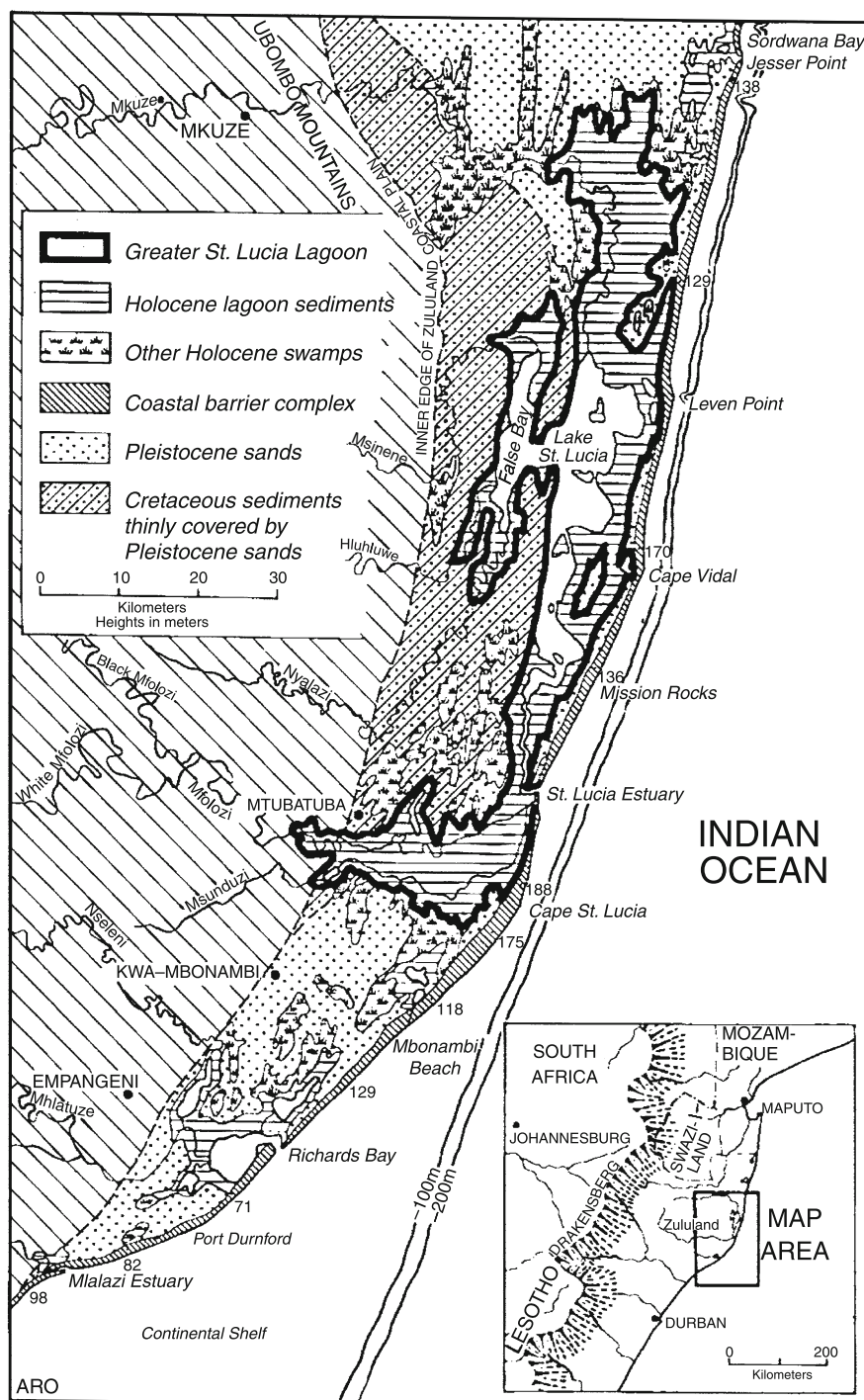
many ages and types occur: barrier reefs, fringing reefs, coral banks, and islets. Because muddy detritus inhibits reef growth, these reefs occur between major estuaries. Relative sea-level change since the last interglacial (~125 ka) is reflected in the barrier beach 8–12 m above sea level behind east coast pangalanes, in raised coral reefs at the northern tip, and in submarine coastlines and aeolianites elsewhere (Battistini 1978).

The Red Sea Coast

The Red Sea extends 2,000 km NNW from the narrow Bab el Mandeb, ranges from 150 to 300 km in width, and deepens to over −2,300 m along its central axis (Fig. 9). This embryo ocean occupies a complex rift system between the African and Arabian plates. Separation began with left-lateral trans-tension in Cretaceous–Eocene time but accelerated with a 7° counter-clockwise rotation of the Arabian block after Oligocene time. The Red Sea opened to the Indian Ocean in Miocene time but this link has not been continuous. It remained open to the Mediterranean until the early Quaternary. Then, with closure of the Mediterranean link and eustatic and tectonic changes at Bab el Mandeb, the Red Sea became isolated from time to time during the later Quaternary.

Africa, Coastal Geomorphology, Fig. 9

Greater Lake St. Lucia lagoon, Kwa-Zulu, on the site of a late Pleistocene barrier-lagoon complex, was reformed by the Flandrian transgression and has since been largely filled with Holocene lagoonal sediment



With potential evaporation approximating $2,000 \text{ mm yr}^{-1}$ but rainfall ranging from only 3 mm yr^{-1} in the north to 150 mm yr^{-1} in the south, episodic closure of the Red Sea from the Indian Ocean has had a dramatic impact on water volume and thus coastal environments.

Owing to the recency of separation, faulted Precambrian basement rocks lie close to the coast and dissected

mountains plunge toward a narrow ribbon of coastal plain underlain by later Cenozoic sediments. The northern Red Sea, together with its northern arms in the 10–25 km wide, –1,800 m deep Gulf of Aqaba and shallower Gulf of Suez graben, reflect mainly normal tensional faulting of continental margins. South of Ras Banas, however, the Red Sea is structurally more complex and falls stepwise into a narrow



Africa, Coastal Geomorphology, Fig. 10 The Tugela River drains the east face of the Drakensberg and much of Kwa-Zulu. Driven by prevailing southerly swells, its sediment plume provides abundant material for coastal beach and dune formation farther north. (Photo: A.R. Orme)

central trough where ascending volcanics have overlapped the continental margins.

The African coast of the Red Sea is characterized by a faulted staircase of coral reefs that form wedge-like plates up to 10 m thick, several kilo-meters wide, and range from 50 m above to –100 m below present sea level. Onshore, these reefs are mantled by aeolian sand, lagoonal evaporites, estuarine deposits, and alluvial gravels. The coastal plain widens only where major valleys reach the coast, as in the Tokar delta into which the Baraka River flows between June and September, or where structural troughs occur, such as the Dallol Basin, a saline depression that descends to –116 m below sea level where the Red Sea rift turns inland south of Massawa. Offshore faulted paleoreefs form the Dahlak Archipelago at the northern end of the Danakil horst. The paucity of blanketing terrigenous sediment allows modern fringing reefs to thrive in warm transparent waters. Reef margins are dominated to windward by *Acropora* associations, to leeward by *Porites* associations (Braithwaite 1982). Lateral reef growth of several centimeters annually even occurs in front of now dry early Holocene river deltas and abandoned *marsas* (*sharms*), corridors cut by wadis draining to low Pleistocene sea levels. Barrier reefs and atolls are less common but are found overlying

subsiding fault blocks from Port Sudan to Ras Hadarba (Mergner and Schuhmacher 1985).

Cross-References

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- ▶ [Sharm Coasts](#)

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Algal Rims

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History

Since the beginning of coral reef studies algal formations of various sizes and shapes, made of coralline algae and other organisms and generally associated with coral reefs, have been described under various names (algal ridges, crests, mounds, reefs, etc.). They were described as being developed upon the windward edge of reefs, but mention of algal mounds, crusts or rims developed on rock also exist in the early literature. A detailed description of algal rim structure was given by Tracey et al. (1948) on Bikini atoll.

Locations were mainly Pacific, but analogous formations were also described in the Atlantic, such as the “boilers” of Bermuda (Agassiz 1895). Later papers (Kempf and Laborel 1968; Gessner 1970; Glynn 1973; Adey and Burke 1976; Focke 1978; Jindrich 1983; Bosence 1984) dealt with algal and animal populations associated with Atlantic rims, and demonstrated their identity with Indo-Pacific rims, and their independance from coral reefs proper, since algal rims can develop directly upon littoral rock, or in regions where corals do not thrive.

Definition and Morphology

Algal rims are reef-like structures (bioherms or biostromes, following their size, development, and influence on local sedimentation processes) developing on the outer edge of reef-flats as well as on rocky coasts submitted to permanent, strong surf (trade winds). They exist both in Atlantic and Indo-Pacific regions but rim-like formations have also been described in the Mediterranean (Blanc and Molinier 1955).

They are built mainly by massive or encrusting coralline algae (mostly *Porolithon*), associated with Hydrocorals (*Millepora*), and Vermetid Gastropods (*Dendropoma*), with a few sturdy corals. Specific composition varies with exposure, slope, and nature of substrate (Focke 1978); coralline algae replacing vermetids when wave energy increases.

Rims are often built-up upon the inner parts of spur-and-groove systems by fusion of algal heads (Tracey et al. 1948).

In extreme surf conditions, upward water movement leads to algal terraced pinnacles or blowholes.

Rim populations gradually mingle seaward with populations of the outer slope, rich in coralline algae down to several meters deep but enriched in corals (such as Atlantic *Acropora palmata*).

In weaker surf conditions, the rim gradually passes into a spur-and-groove reef edge or to thin rim-like formations along rocky coasts. Algal rims or equivalent structures may develop in tideless as well as in tidal conditions.

An often overlooked fact is that structures with algal rim or rimmed terrace morphology may be formed not by constructional evolution but by an erosive process. The Bermudian “boilers” may thus be divided into “erosive boilers” generated by erosion of an emerged stack of soft rock, the edge of which is protected by a thin organic cover (original description of a Bermudian boiler) and into “constructive boilers” obtained by the upward growth of a coral patch, covered by a surface layer of vermetids (Ginsburg and Schroeder 1973).

Relation with Past Sea Levels

Some components of algal rims (*Dendropoma* vermetids, some *Lithophyllum*) have a narrow repartition range around MSL and their presence in cores or on elevated coastlines is a precise indicator of ancient sea levels, with an approximation of less than 1 m (Jindrich 1983; Laborel 1986; Laborel et al. 1994). Conversely, corals such as *Acropora palmata*, have less precise indicators (Lighty et al. 1982) unless a true rim may be recognized in the core.

Further study of algal rims is nevertheless needed to obtain better biological indications of past sea level stands.

Cross-References

- Bioherms and Biostromes
- Coral Reefs
- Sea-Level Indicators, Biologic

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Alluvial-Plain Coasts

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Definition

Alluvial coasts form under the influence of marginal-marine processes on sedimentary deposits of fluvial origin. These may be alluvial fans, alluvial plains, or terraces of low slope and elevation. These coasts are neither strongly outbuilding of retreating, and are not strongly influenced by tectonic or isostatic uplift or subsidence.

Introduction

Alluvial-plain coasts are systems “formed where the broad alluvial slope at the base of a mountain range is built out into a lake or the sea” and are part of a “neutral” coast (Johnson 1919, p. 188). The term “neutral coast” is little used in modern publications, but is still a useful classification of coastlines built by interaction of river and coastal processes where sediment input and relative sea level change are morphodynamically balanced, producing a coastline that neither

progrades nor retreats rapidly. Alluvial-plain coasts may be closely associated with deltas, estuaries, beach-ridge plains, barrier-lagoon systems, and in some settings glacial outwash plains. Variations in geomorphology and process responses among these coastal environments are determined by the rate of relative sea level change and the abundance or paucity of sediment supply. This sediment supply may come from local bluff erosion, alongshore from fluvial or other sources, or from offshore. Climate, wave and tidal energy, and underlying geologic controls such as relief, complexity of coastline shape, and tectonic setting are other controlling factors.

Discussion

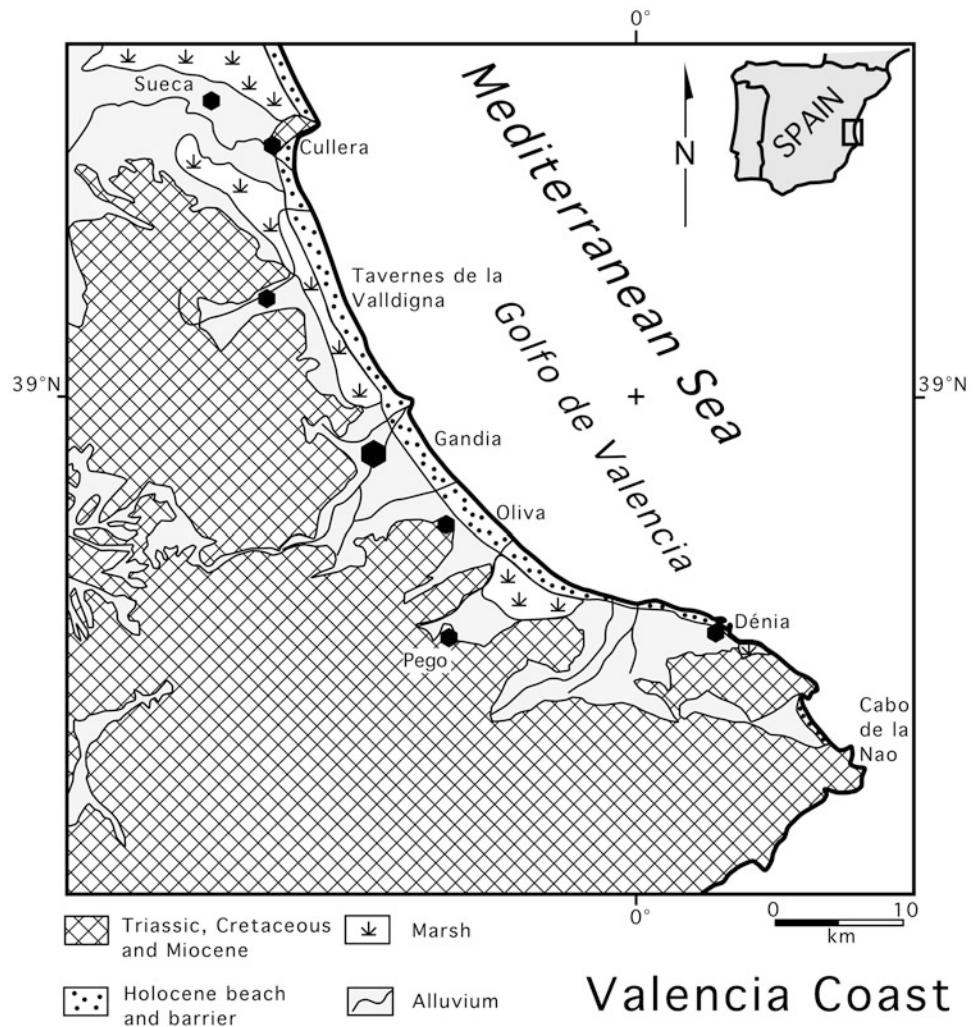
Tectonic setting starts from a plate tectonics viewpoint. Inman and Nordstrom (1971) provide a broad-brush classification of the tectonic activity and morphology of the world's coastlines. Alluvial-plain coasts are found primarily on passive margin coasts. In sequence stratigraphic terminology, alluvial-plain coasts are expected within a highstand systems tract (Posamentier and Vail 1988) or falling-stage systems tract (Plint and Nummedal 2000). These systems tracts form during rising to highstand sea level and highstand to falling sea level, respectively. However, the use of these interpretations clearly requires an understanding of rates of sediment influx and the influence of local basin subsidence or tectonic uplift. Boyd et al. (1992, Fig. 2) show a summary model of coasts of both outbuilding and retreat, with the neutral, alluvial plain coasts an intermediate between the two.

Modern coastal environments are graded to the late stages of Holocene sea level change, which varies around the globe, but has been rising at a gradually slowing rate in much of the Northern Hemisphere over the past 6000 years. However, during times of slow change, such as the last interglacial highstand alluvial-plain coasts may have been more predominant. Blum and Price (1994) show examples of this setting on extensive alluvial fans created at highstand on the Texas coast. A similar argument can be made for lowstand systems.

Low-relief alluvial plains are built by fluvial processes such as meandering stream pointbar accretion, deposition on levees, over-bank flood deposition, and a variety of channel avulsion processes that create oxbow lakes, flood chutes, and crevasse-splay deposits. In higher-relief coastal settings fan deltas may develop. Fan deltas are sloping alluvial sediments deposited in a sweeping arc where mountain streams flow out onto lowlands and into a sea or lake, with a substantial subaqueous extent (Prior and Bornhold 1989). Their subaerial portions are alluvial fans, dominated by braided stream and sediment-gravity flows in irregular flow conditions. Their subaqueous portions include fan-shaped deltaic and prodelta facies, usually with only a narrow shelf. Alluvial-plain coasts are influenced by littoral processes, either by the reworking by

Alluvial-Plain Coasts, Fig. 1

Alluvial-plain coast of eastern Spain, the Valencia Coast (After Viñals and Fumanal 1995, Fig. 1)



waves and tides of fluvial sediments brought into the coastal setting or by the gradual infill of preexisting embayments.

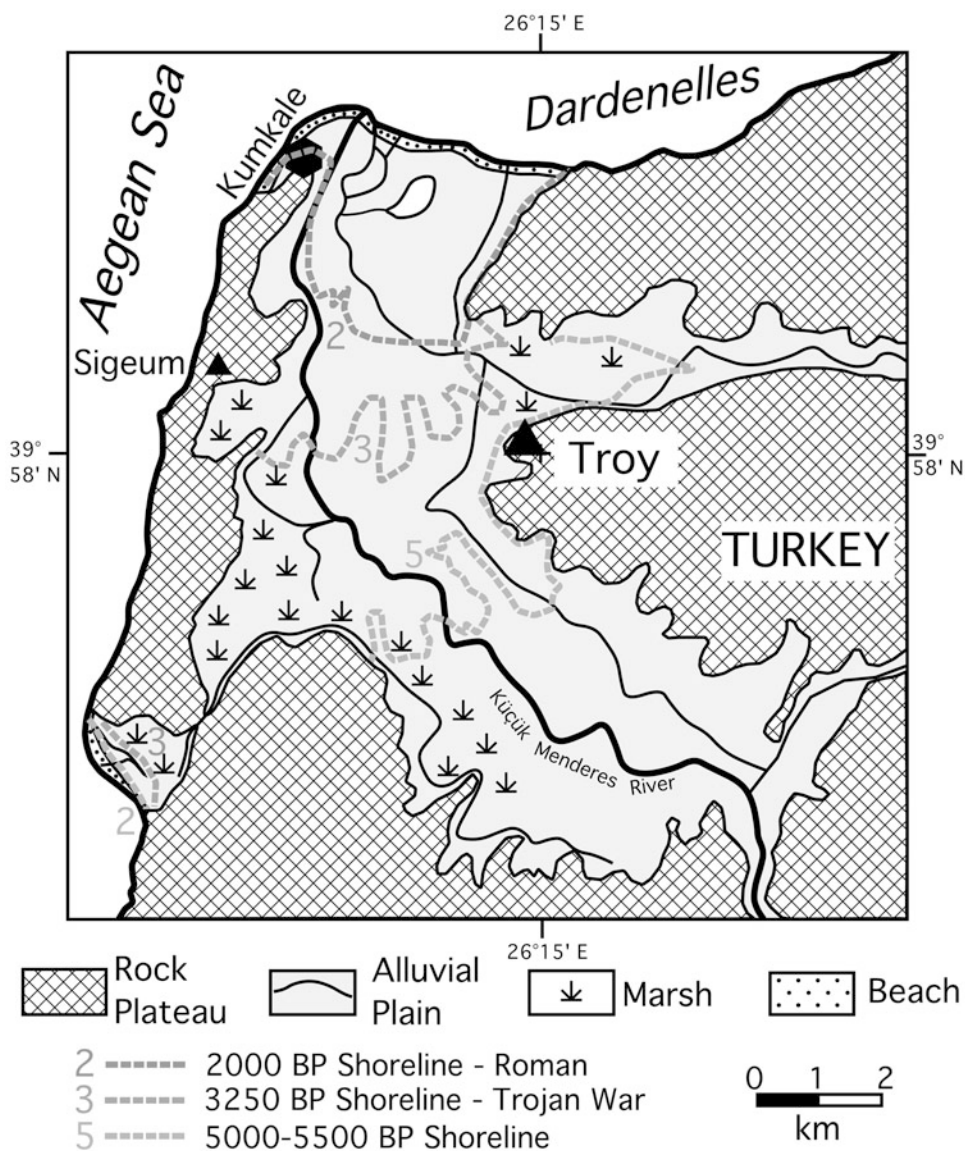
Wave-dominated alluvial-plain coasts occur on relatively simple coastlines, where there is sufficient fetch and wave energy to prevent the outward building of a distinct delta form. The coast may be somewhat tectonically influenced and of moderate relief, such as the Mediterranean Valencia coast of eastern Spain (Fig. 1) (Viñals and Fumanal 1995). The very low relief Texas coast has alluvial-plain coastal segments at High Island, and on the Brazos River delta (Rodriguez et al. 2000, 2001). The Peruvian coast is battered by higher wave energy and has steep cliffs in many sectors. However, valley mouths receive abundant coarse and fine sediment from rivers heading in the foothills of the Andes, modulated by the important climatic variability of the El Niño – Southern Oscillation as well as earthquake destabilization. Fluvial reworking and resurfacing creates alluvial-plain coasts near river mouths, while sand and gravel are reworked into well-developed beach ridge plains northward (down-drift) of the Santa, Piura, and Chira Rivers in

northwestern Peru (Richardson 1983; Sandweiss 1986; Belknap and Sandweiss 2014).

Embayment-infill alluvial plains occur on complex coastlines with higher relief. Sediments are trapped in coastal compartments with relatively limited fetch and lower wave energy. The coastal valleys of western Turkey (Fig. 2) and Greece contain examples of this type of coast. The coastlines in these embayments have prograded many kilometers in historic times, with the result that famous archaeological sites such as Troy and Thermopylae are now well inland of the coastal settings described in antiquity (Kraft et al. 1980, 1983, 1987). Figure 2 shows the progradation of the Küçük Menderes alluvial plain 7 km from a shoreline within a protected embayment at 5000–5500 yrs. BP, more than 5 km since the Trojan war 3250 BP, and more than 3 km since Roman times 2000 BP. These reconstructions are based on interdisciplinary coastal geology and geomorphology, geoarchaeology, historical, and other written records (Kraft et al. 2003). Recognition of the potential for coastline change is crucial to the interpretation of archaeological sites on alluvial-plain coasts.

Alluvial-Plain Coasts, Fig. 2

Alluvial-plain coast of northwestern Turkey, the Trojan plain of the Küçük Menderes River (After Kraft et al. 2003, Figs. 1 and 4–6)



Alluvial-plain coasts associated with large rivers may grade into delta systems, representing an inactive lobe of a delta cycle (Coleman 1988), where the coast is distant from the primary river influx. The Huanghe (Yellow River) entering the Bo Hai Sea of northeastern China has an active modern birdsfoot delta, but there are also broad stretches of alluvial-plain coast on the Bo Hai coast that formed by coastal reworking of former fluvial/deltaic environments (Saito et al. 2001). Marsh and swamp environments, precursors to the coals found in ancient sequences (e.g., Fielding 1987), are commonly associated with delta plain and alluvial-plain coasts.

Recognition of alluvial-plain coasts in ancient rock sequences may be challenging, particularly when attempting to distinguish them from delta plain and barrier-lagoon

systems. However, the Cretaceous Mesaverde Group littoral facies provide clear examples of alluvial-plain coasts on the margin of the interior seaway of Colorado and Wyoming (Hollenshead and Pritchard 1961; Kraft and Chrzastowski 1985). This succession includes transitions from fluvial delta plain (Menefee Fm.) with associated coal and carbonaceous shales, to the Lookout Point Fm. sandstone, in a regressive succession, overlain by the Cliff House Fm. Sandstone, in a transgressive succession. Determining the relative influences of eustatic sea-level fluctuations and rates of sediment supply, as well as climate, subsidence, and local process variations, can be even more challenging in the ancient rocks than in Holocene systems, where geomorphology and geologic setting are more directly traced and more completely understood.

Summary and Conclusions

Alluvial-plain coasts may be underrepresented in the literature because they are combined with delta, barrier, or estuarine environments. They represent intermediate conditions of balance between sediment accumulation and sea-level change. They are neither strongly prograding, like deltas, nor transgressed like estuaries. They reflect primary fluvial processes of deposition, but also significant reworking by littoral processes. They may be recognized by the lack of distinct barrier and lagoon environments and close juxtaposition of alluvial and beach systems.

Cross-References

- [Barrier](#)
- [Beach Ridges](#)
- [Coastal Seafloor Geomorphological Features, Classification](#)
- [Deltas](#)
- [Estuaries](#)
- [Paleocoastlines](#)
- [Sandy Coasts](#)
- [Sequence Stratigraphy](#)

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Altimeter Surveys, Coastal Tides, and Shelf Circulation

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Definition

A conventional radar altimeter aboard a satellite is a nadir-looking active microwave sensor. Its signal pulse, transmitted vertically downward, reflects from the ocean surface back to an altimeter antenna. The round-trip time and the propagation speed of the electromagnetic waves are used to compute the range between the antenna and the ocean surface.

From the altimeter-measured range, the instantaneous sea surface relative to a reference surface, such as an ellipsoid, can be determined if a satellite orbit relative to the reference surface is known. With the knowledge of the oceanic geoid, the sea surface topography relative to the geoid due to ocean dynamic circulation including the temporal averages can be

obtained. Although the geoid is not well determined yet, repeated observations can provide a measurement of the temporal variability of the sea surface height since the geoid can be treated as time-invariant for oceanographic applications.

Oceanographic applications of satellite altimetry require an accuracy of a few centimeters. This requirement constrains not only the altimeter sensor but also pertinent atmospheric and oceanographic corrections that have to be made, the satellite orbit, and, in some applications, the reference geoid. One-second data are provided to the science community as the Geophysical Data Records (GDR). So far, the accuracy of altimetric sensors has steadily improved to a few centimeters. The atmospheric corrections consist of ionospheric delay and wet and dry tropospheric delays; the oceanographic effects are composed of the sea state bias, inverse barometric response, the elastic ocean tides, solid Earth tides, and pole tides. The radial uncertainty of the orbit has long been one of the largest error sources in satellite altimetry until the TOPEX/Poseidon (T/P) mission. So far, the root mean square accuracy of the orbit used for producing GDRs is better than 1.5 cm. At present, the geoid errors occurring at various spatial scales are one of the largest error sources and the major obstacle to the derivation of the mean current from altimeter data.

Because almost all radar altimeter missions use a repeating orbit, repeat track analysis or collinear analysis becomes a conventional approach for application of altimeter data in physical oceanography. A mean at a location is calculated from data available, and then the mean is removed from the data to produce sea surface height anomalies relative to the mean. Thus, the repeat track analysis eliminates the geoid and its errors, the mean sea surface, and a portion of the orbit error. There are three variants from the standard approach. One is to subtract a selected instantaneous sea surface instead of the mean sea surface; another calculates the along-track slope deviations instead of sea level deviations. For a tidal analysis, collinear differences are usually calculated to bypass the uncertainty associated with the mean sea surface.

While satellite altimeters perform well over the open ocean, the reflected signals in coastal waters have artifacts due to the presence of land within the instrument footprint. To remedy these artifacts, waveform retracers have been developed, including empirical ones and physically based ones (Gommenginger et al. 2011).

Conventional radar altimeters are pulse-limited. New altimeter sensors have been developed. For example, a delay-Doppler altimeter (also called as synthetic-aperture radar (SAR) altimeter) applies coherent processing to groups of transmitted pulses and thus utilizes the full Doppler bandwidth to make the most efficient use of the power reflected from the ocean surface. The SAR altimetry has high along-track resolution and can measure closer to the coast

than the conventional pulse-limited altimeter. Both Cryosat-2 and Sentinel-3 missions operate SAR mode altimeters.

Introduction

Satellite altimetry produces global measurements of instantaneous sea surface heights relative to a reference surface and is one of the essential tools for monitoring ocean surface conditions and understanding physical oceanographic processes. Altimetric sea surface height measurements are one of the most important data for ocean forecasting. Research using altimetric data has significantly improved global ocean tide models (Stammer et al. 2014) and greatly advanced our understanding of mesoscale circulation variability in the deep ocean (Fu and Cazenave 2001).

In the past decade or so, coastal altimetry has also advanced to a great deal (Vignudelli et al. 2011). Coastal altimetry products have been developed, which provide valid data points within ones of kilometers from the coast. Among those products are the Innovative Processing System Prototype for Coastal and Hydrology Applications (PISTACH, Mercier et al. 2010), the Prototype for Expertise on AltiKa for Coastal, Hydrology and Ice (PEACHI, <http://www.aviso.altimetry.fr/en/data/products/sea-surface-height-products/global/experimental-saral-product-s-peachi.html>), the SAR Altimetry Coastal and Open Ocean Performance (SCOOP, <http://www.satoc.eu/projects/SCOOP/>), and the X-TRACK (Birol et al. 2016). This article reviews advances related to applications of satellite altimetry to coastal tides, storm surges, as well as coastal and shelf circulation.

Coastal Tides

Coastal tides are mainly forced by adjacent deep-ocean tides which themselves are generated by the tide-generating potential associated with the motion of both the moon and the sun. The tidal waves in coastal seas propagate as trapped Kelvin waves, subject to significant amplifications in magnitude and reductions in wavelength. Shallow water tide constituents are generated due to nonlinear effects.

Several empirical methods can be used to derive tides from altimetric time series. These include but are not limited to harmonic analysis, response analysis, and the inverse method. The first two methods are applied to each geographical location, and the last method considers the spatial correlation as defined by a priori spatial covariance in the tidal signal.

Both harmonic and response analyses are purely temporal analyses, common in the analysis of hourly tide gauge data. The methods are suitable for either coastal seas (Woodworth and Thomas 1990; Han et al. 1996) or deep oceans (Cartwright and Ray 1990), not constrained by large satellite track intervals. However, an assimilative model in which altimetric

tidal information is optimally blended with a dynamical model would be a great benefit in providing spatial interpolation and extrapolation of tidal fields (Han et al. 2000; Egbert and Erofeeva 2002).

In the past decade, global tide models have further advanced and their accuracy over the shelf has been significantly improved. The root-sum-square (RSS) errors for the eight major tide constituents at 195 tide gauge stations over the continental shelf are as small as 5 cm (Stammer et al. 2014). At a smaller number (56) of coastal tide gauge stations, the RSS errors for the eight tide constituents are as small as 7 cm. Nowadays, global tide models are still used to correct for satellite altimetry by coastal altimetry users. It is cautioned that global tide model errors can be large in coastal waters. Where possible, coastal altimetry users should use validated local tide models with sufficient accuracy. Accurate, high-resolution coastal tide models are needed for coastal altimetry. More accurate bathymetric information is critical for improving coastal tidal simulations. Coastal tide models need to be smoothly blended with global tide models for widespread use by the altimeter community (Ray et al. 2011).

Storm Surges

Storm surge is the main factor that causes extremely high sea level and catastrophic coastal flooding, such as in New Orleans during Hurricane Katrina and in the Philippines during Typhoon Haiyan. Coastal tide gauges have been the main tool for monitoring storm surges. Hydrodynamic models have been used for storm surge forecasting.

In the past decade, there have been studies showing the utility of satellite altimetry for observing storm surges, e.g., in the Gulf of Mexico during Hurricane Katrina (Scharroo et al. 2005), off Newfoundland during Hurricane Igor (Han et al. 2012), off the US eastern coast during Hurricane Sandy (Lillibridge et al. 2013; Chen et al. 2014), in the North Sea during Cyclone Xaver (Fenoglio-Marc et al. 2015), and off the northeastern coast of the Gulf of Mexico (Han et al. 2017). Antony et al. (2014) showed that the multi-mission satellite data significantly enhance the chance of detecting storm surges in the Bay of Bengal. It is opportunistic for an altimeter to capture storm surge, while a constellation of altimeter missions may significantly enhance the chance. Relative to a conventional altimeter, a wide-swath altimeter has much high capacity of capturing storm surges (Turki et al. 2015; Han et al. 2017).

Subinertial Coastal and Shelf Circulation

Another important application of satellite altimetry in physical oceanography is to subinertial variability over a continental shelf and in coastal waters. One of the key components of the geophysical corrections is the removal of oceanic tides.

In coastal seas, tides are often larger than subinertial variability and have horizontal scales of about 100 km owing to the local coastline and bottom topography. During the early years of accurate altimetry, global tidal models were clearly not adequate for shallow waters; for example, the M_2 amplitudes and phases in Geosat GDRs are significantly different from both tide gauge and altimeter data over the Scotian Shelf and the Grand Banks (Han et al. 1993) and in the Amazon mouth (Minster et al. 1995). The tide model in T/P GDRs was found not sufficient over the Canadian Pacific shelf (Foreman et al. 1998).

Because the semidiurnal and diurnal constituents are aliased into fluctuations with longer periods, their errors may obscure subinertial fluctuations of interest. For example, the K_1 has an aliasing period of 173 days in the T/P data, close to a semiannual cycle. One way to reduce the tidal correction errors is to use regional tidal models that have higher accuracy (Han et al. 1993; Ray et al. 2011). The separation of the low-frequency variability of interest from the aliased tidal constituents can also be achieved using harmonic analysis. The results may be subject to notable statistical uncertainties, depending on the length of data, magnitudes of tidal errors, and error correlation.

The altimetric data can be used to calculate sea surface slopes and thus to derive geostrophic surface currents. There have been increasingly more applications of satellite altimetry for coastal circulation, such as in the Mediterranean Sea (Bouffard et al. 2011), the Caspian Sea (Kouraev et al. 2011), the Black Seas (Ginzburg et al. 2011), the Barents and White Seas (Lebedev et al. 2011), as well as off the coasts of North America (Emery et al. 2011) and of China (Han and Huang 2008).

The geostrophic currents at depth can be determined when combined with interior density structure. Han and Tang (1999) combined the T/P data with density and wind data to study seasonal variability of current and transport in the Labrador Current. The method was also used to study interannual (Han and Tang 2001) and multiyear variability of the Labrador Current (Han et al. 2010).

The spatial sampling resolution is critical for shelf circulation. Dynamic circulation models are needed for a space-time interpolation to make the best use of the information. The assimilation of altimeter data into numerical models is one of the most promising approaches to study subinertial coastal and shelf circulation. The availability of near-real-time satellite altimetry data is very useful for nowcasts and forecasts of coastal ocean conditions.

Conclusions

Application of satellite altimetry for sea level and currents in coastal waters remains more challenging than in the deep ocean (Han 1995). These challenges are being tackled actively from several fronts. Dynamic features in coastal

seas are strongly influenced by the complex coastline and bottom topography and have temporal scales of several days to weeks and spatial scales of ones to tens of kilometers. A conventional satellite altimeter is not good at sampling these small-scale features. Altimeter waveforms are subject to land contamination upon crossing the land-sea boundary; useful data are not available close to the coast. Atmospheric and oceanographic corrections are more uncertain for coastal waters than for the deep ocean. SAR and interferometric altimeters have been used to improve spatial sampling resolution; waveform retracking algorithms have been applied to alleviate the land contamination associated with conventional altimeters. Special editing techniques such as for wet tropospheric effects and correction models such as for dynamic atmospheric pressure effects have been applied to improve atmospheric and oceanographic corrections.

High temporal and spatial resolutions are very important, particularly in the coastal seas. Unfortunately, fast (more frequent) sampling must be at the expense of large orbit interval (cross-track separation), and a reasonable compromise must be made to best meet mission requirement.

A wide-swath interferometric altimeter provides high-resolution two-dimensional image of sea surface topography within each swath. An appropriately formed constellation of satellite altimeter missions can improve both the spatial and temporal resolutions.

Dynamical models with data assimilation are useful for optimally blending the model dynamics with altimetry. Only dynamical models may be able to effectively derive ocean interior conditions from their sea surface manifestation observed by an altimeter.

Cross-References

- [Changing Sea Levels](#)
- [Natural Hazards](#)
- [Remote Sensing of Coastal Environments](#)
- [Sea-Level and Climate Change](#)
- [Storm Surge](#)

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Antarctica, Coastal Ecology and Geomorphology

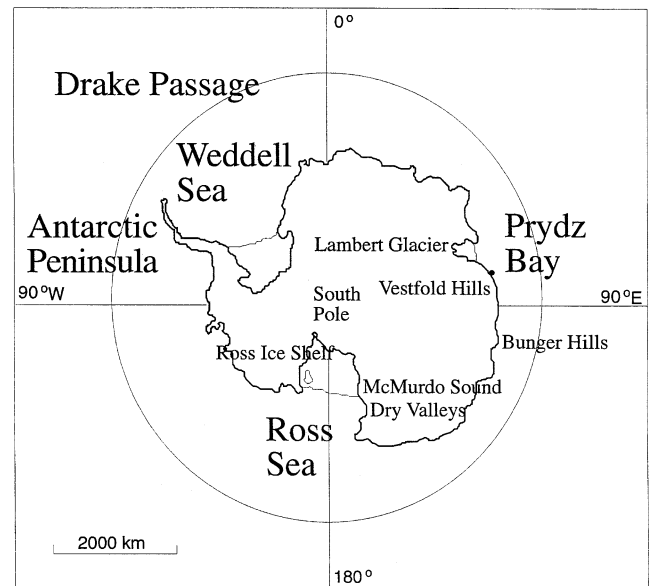
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Ecology of Coastal Antarctica

Antarctica covers an area of 14 million km² and has a coastline over 32,000 km long (Fig. 1). However, more than 80% of this coastline is comprised of ice cliffs formed by either glaciers or the polar plateau. On average, the continental shelf is both narrower and deeper than other continents, averaging more than 600 m deep.

Marine life in Antarctic waters is controlled by the presence of sea ice. This ice, which can be more than 2 m thick, grows out from Antarctica each year until it covers approximately 20 million km²; more than doubling the area of the continent. It reaches its maximum extent by late October but by late January it has retreated to the coast. The ice affects



Antarctica, Coastal Ecology and Geomorphology, Fig. 1 Map of Antarctica

Antarctic life in two main ways. Most importantly, it reduces the amount of light reaching the underlying water column and so reduces the primary productivity and secondly, together with icebergs, the sea ice also disturbs and scours much of the shallow coastal zone.

The first explorer to sail south of the Antarctic Circle was secondly Captain James Cook in 1773. He noted the abundant seal and whale life but concluded “that no good would come of it” (Antarctica). His observations on the abundant seal and whale life, however, were to have a profound impact on the ecology of the Southern Ocean as it led to the focus of the whaling and sealing industries moving from the Northern to the Southern Hemisphere. Over the subsequent 150 years these industries effectively removed the higher order predators from the ecosystem and only now is there evidence that seal and whale populations are beginning to recover from their early savage onslaught. The first record of penguins was much earlier and surprisingly, by Sir Francis Drake in 1644. He commented

Wee found great store of strange birds, which could not flie at all, nor yet runne so fast as they could escape with their lives; in body they are less than a goose and bigger than a mallard, short and thicke sett together, having no feathers, but instead thereof a certaine hard and mattered downe; their beaks are not much unlike the bills of crows, they lodge and breed upon the land, where making earthes, as the conies doe, in the ground, they lay their egges and bring up their young, their feeding and provision to live on is the sea, where they swimm in such sort, as nature may seeme to have granted them no small prerogative of swiftnesse.

Many of the early explorers had biologists aboard and by the beginning of the twentieth century most of the important components of the marine ecosystem had been described.

Research over more recent years has concentrated on the significance of the microbial ecosystem, specific adaptations to the unique characters of the Antarctic environment, and the effects of human impacts, such as global warming, the ozone hole, over fishing, and pollutants.

The base of all marine food chains is comprised of algae. In the open ocean this is comprised exclusively of phytoplankton. In coastal areas seaweeds and benthic microalgal mats are also important. In Antarctica, a fourth and often dominant component is contributed by the sea ice algae.

Phytoplankton in the waters around Antarctica are comprised of the same basic algal divisions as elsewhere. Diatoms are particularly prominent but summer blooms are often comprised of phytoflagellates such as Prymnesiophytes (*Phaeocystis* in particular), Cryptophytes, Prasinophytes, and dinoflagellates.

The largest size class of phytoplankton, that is, the net plankton (20–200 μm), is dominated by diatoms and these comprise most of the large spring algal blooms. The nanoplankton (2–20 μm), unlike most other areas is also dominated by diatoms, *Fragilariopsis cylindrus* in particular, but *Phaeocystis* and Parmales can also form large blooms. Cyanobacteria (i.e., blue green algae) characterize the picoplankton (0.2–2 μm) in most other areas but these are almost completely absent from Antarctic waters.

Each year as the ice melts and the pack ice retreats, it leaves behind it a surface layer of slightly fresher water. Algal cells fall out of the melting ice and contribute to a major annual algal bloom which tracks the retreating ice edge back to the continent. When it reaches the continent a major summer coastal phytoplankton bloom results. This bloom can account for more than 60% of the annual primary productivity and has a similar biomass concentration to phytoplankton blooms elsewhere in the world (up to 10 $\mu\text{g Chla l}^{-1}$). At other times of the year the phytoplankton biomass is very low ($<0.1 \mu\text{g Chla l}^{-1}$) but comparable to many other remote ocean sites. In most oceanic areas of the world the availability of nutrients (i.e., nitrate, phosphate, and silicate) limits the growth of algal blooms. In Antarctica, however, this is rarely the case and nutrients are nearly always present in excess. Here phytoplankton growth is usually limited by low light levels and deep mixing. Limited iron (Fe) availability is now also thought to reduce phytoplankton productivity over large areas of the Southern Ocean, but this is unlikely to be the case in coastal areas where iron is available from both the sediments and wind-blown dust. Early explorers often noted the teeming wildlife of Antarctic waters and concluded that it was an area of high productivity but in fact more recently we have found that the reverse is the case. The explorers typically sailed south as the ice was retreating in spring and early summer, a time that coincides with the one time in the year when the melting ice creates the conditions that allow a transient algal bloom. This bloom in

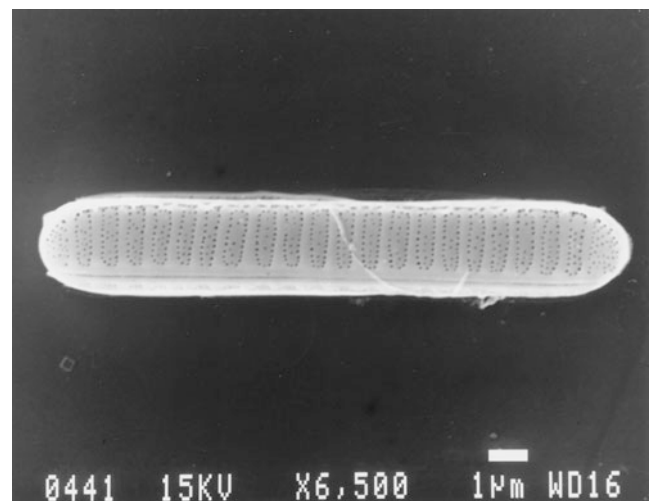
turn attracted the abundant zooplankton, birds, whales, etc. The rest of the year the Southern Ocean is remarkably unproductive.

Seaweeds make only a minor contribution to Antarctic primary productivity, first because there is little rocky substrate and second because of the disturbance caused by sea ice and icebergs. Diversity is relatively low and is dominated by red algae. Marine benthic algal mats have not received much attention although it is likely that they contribute up to 50% of the primary production in many shallow marine areas. They are dominantly comprised of diatoms.

When the sea ice reforms in autumn phytoplankton cells are taken up into the ice. The few species that are able to survive do well and grow to be released as an inoculum when the ice again melts the following spring.

The algae form communities either at the top of the ice, within the interior or on the bottom. In late winter and spring in coastal areas the most important communities grow as either encrusting or hanging communities on the bottom of the ice. It is only these bottom communities that are available as food to grazing invertebrates and fish. The algal biomass growing on the bottom of the ice is so high (380 mg Chla m^{-2}) the under-surface of the ice often appears dark brown to black and has a biomass approaching the theoretical maximum for the available sunlight. In coastal areas the dominant species are the diatoms *Nitzschia stellata*, *Entomeneis kjellmannii*, *Berkelaya adeliense*, and *Thalassiosira australis*.

Further off shore in the pack ice, the ice communities are dominated by the diatoms *Fragilariopsis curta* and *F. cylindrus* (Fig. 2).



Antarctica, Coastal Ecology and Geomorphology, Fig. 2 Scanning electron microscope photograph of *F. cylindrus*, an important diatom species in both the phytoplankton and sea ice. It also comprises a major proportion of many bottom sediments

With the arrival of warmer temperatures in late spring, melt pools develop on the surface of the ice. These are predominantly colonized by a dinoflagellate, *Polarella*, and chrysophytes. These organisms are released into the water column when the ice finally melts or breaks out in late December or January.

The seafloor, sea ice, and the water column all contain abundant and diverse communities of bacteria and protozoans. These organisms, together with the phytoplankton, comprise the elements of the microbial loop (Fig. 3). The loop is so named because it describes the route taken by the organic material which is created by the phytoplankton. Phytoplankton use the sun's light energy to synthesize organic matter which they use to grow and reproduce. These cells are grazed on by heterotrophic protists, which in Antarctic coastal waters are mostly dinoflagellates, amoebae, and choanoflagellates. These organisms either die or are attacked by bacteria or viruses, which then decompose to become nutrients for the phytoplankton. It has been estimated that as much as 50% of the energy, which has been harvested from the sun by the phytoplankton, is circled around this loop and thus not made available to multicellular grazers.

The sea ice algae is directly grazed on by small crustaceans such as amphipods, krill, and copepods. These are able to consume up to 2% of the total biomass per day. The other main grazer on sea ice algae is the ice fish, *Pagothenia borchgrevinki*. Much of the primary production of the sea ice falls to the bottom when the ice begins to melt. Here it becomes available as food, together with the benthic algal

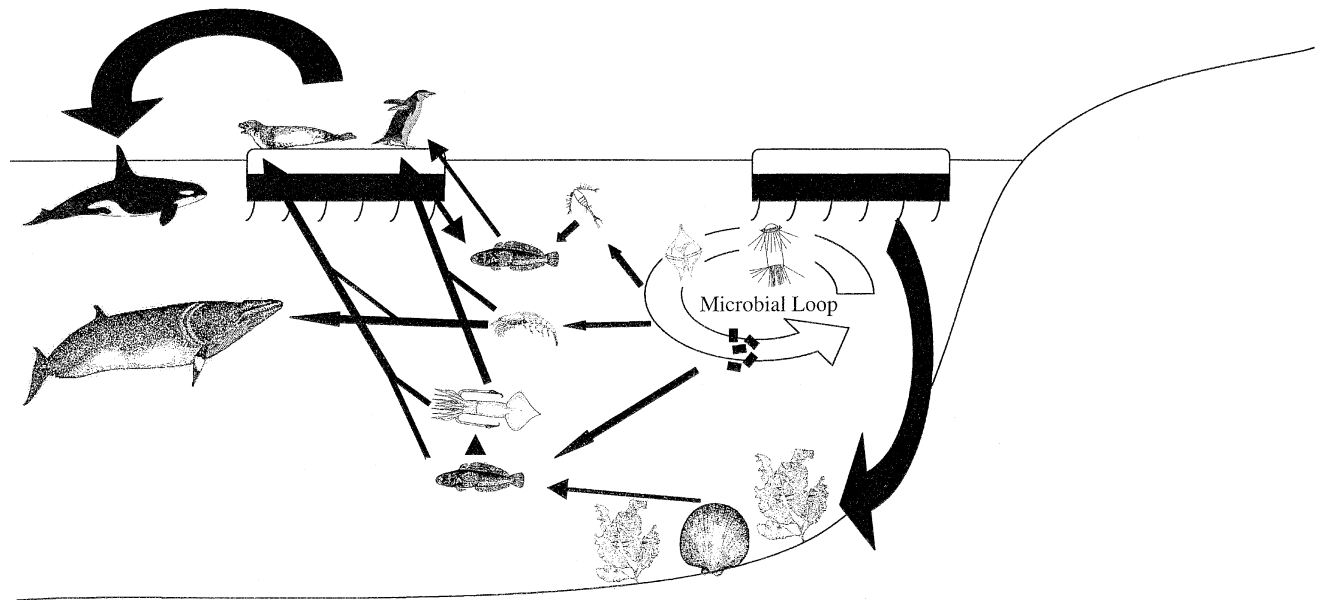
mats, to the benthic invertebrate community. This community is comprised of a diverse range of sponges, starfish, sea urchins, sea slugs, worms, and shellfish.

The best known invertebrate of the Southern Ocean is the Antarctic krill, *Euphausia superba*. This shrimp-like crustacean, which is up to 20 cm long, forms incredibly large aggregations of up to 100 million individuals and is arguably the most abundant multicellular organism on earth. However, it is not usually a creature of coastal areas; there, it's smaller cousin *Euphausia crystallorophius*, takes over. These creatures together with smaller zooplankton such as copepods and fish larvae comprise the main herbivorous grazers of the phytoplankton.

Zooplankton

The herbivorous zooplankton is composed largely of heterotrophic protists (e.g., dinoflagellates), pelagic crustaceans (e.g., copepods, amphipods, and euphasids), larval fish, and salps. Compared to most other areas the biodiversity of the Antarctic coastal zooplankton community is relatively low although the biomass is high. Diversity tends to increase with depth. Salps, gelatinous zooplankton, are very significant filter feeders in pelagic waters around Antarctica but their role in coastal communities is unknown.

Biomass and reproduction is strongly seasonal with the peaks coinciding with the development of the maximum phytoplankton bloom soon after the disappearance of the



Antarctica, Coastal Ecology and Geomorphology, Fig. 3 Simplified food web diagram of Antarctic coastal communities. Arrows indicate that an organism is eaten by the animal that is being pointed at. Diagram shows the importance of the microbial loop and sea

ice communities to the whole ecosystem. The large arrow from the sea ice to the bottom indicates that most ice algal production falls to the bottom where it is grazed by invertebrates

secure. Zooplankton biomass in McMurdo Sound in February has been estimated at 1.5–3.4 g dry weight m^{-3} , a biomass toward the top of the range for coastal communities elsewhere.

Fish and Squid

Compared to most other areas the diversity and biomass of Antarctic coastal fish is very low. There is only one common truly pelagic fish, the Antarctic Silverfish (*Pleuragramma antarcticum*), and this commonly comprises more than 90% of the pelagic fish biomass. This species has been found in the diet of all fish that live above the bottom, Weddell Seals, whales, and both Emperor and Adelie Penguins. The diet of *Pleuragramma* itself is comprised largely of copepods, mysids, and other fish.

There is a community of fish associated with the underside of the sea ice. This community is characterized by *Trematomus* spp. and *P. borchgrevinki*. These fish feed directly on the sea ice biota. Other fish species live and feed near the bottom.

Many species of Antarctic fish have developed remarkable adaptations to help them survive in the Antarctic marine environment. The body fluids of most fish have a salinity which would freeze at approximately -0.7° . Since the water temperature is usually around -1.8° they would freeze solid unless they had a mechanism to deal with it. Many, such as *P. borchgrevinki*, have evolved antifreezes comprised of glycoproteins. These depress the freezing point of the fish by inhibiting the growth of ice crystals as contact will cause damage as they get larger. Fish which do not possess antifreezes, such as *P. antarcticum*, must avoid coming into contact with ice crystals as contact will cause snap freezing.

Squid almost certainly comprise a critical component of the Antarctic coastal ecosystem and yet very little is known about them. Approximately, 72 species occur south of the Antarctic convergence but only a relatively small number are thought to live near the coast. They are known to feed on krill but also feed on a variety of other zooplankton and fish species. They themselves comprise an important component of the diets of several whale species (e.g., Sperm Whale, Killer Whale), seal species (Weddell Seals, Ross Seals, and Elephant Seals), and birds (particularly Emperor Penguins and albatrosses).

Benthic Communities

Antarctic coastal benthic environments are characterized by both stable low temperatures and the effects of ice. Seasonal changes in water temperature are small compared to most temperate areas with a range of only $2\text{--}3^{\circ}$, from -2° in winter to approximately $+1^{\circ}$ in summer. Melting ice in spring and summer reduces the salinity of surface waters. Conversely,

forming ice in autumn and winter expels salt producing water of higher salinity, which then sinks to the bottom.

Sea ice can affect benthic communities down to at least 10 m. The moving ice scours and abrades littoral and sublittoral areas. Icebergs disturb the seafloor down to depths of over 150 m. These bergs, driven by currents and winds can scour rocky surfaces and plough into sediments.

Benthic microalgal mats grow in most shallow nearshore environments. These mats, which are mostly dominated by diatoms, can account for up to 50% of the annual primary productivity. The biomass of these mats is strongly seasonal and is governed by the amount of light reaching the bottom. In winter, when the sun either does not rise or is low on the horizon for only a few hours each day biomass is at a minimum. The amount of light reaching the bottom is usually further reduced by a cover of sea ice up to 2 m thick.

Over 700 species of seaweed are known from the southern Ocean of which approximately 35% are thought to be endemic. Seaweeds are uncommon down to a depth of 5 m due to the scouring action of the sea ice. In McMurdo Sound *Iridaea cordata* is dominant at shallow depths, *Phyllophora antarctica* at intermediate depths and *Leptophyllum coulmanicum* below 20 m. At all locations the diversity increases with depth.

The equable but cold marine environment of Antarctica has led to the development of benthic communities characterized by high biomass but low growth rates. Biodiversity is similar to benthic communities elsewhere but endemism in most major groups, for example, echinoderms, bryzoa, polychaetes, amphipods, etc., is between 50% and 90%.

Epibenthic (i.e., those living on the surface rather than within the substrate) invertebrate communities are also strongly influenced by the actions of sea ice. The upper zone, 0–15 m is almost devoid of animals but is briefly colonized by detritus-feeding echinoids (*Sterechinus neumayeri*) and starfish (*Odonaster validus*), isopods, and small fish. An intermediate zone, 15–33 m is characterized by abundant soft corals, sponges, anemones, hydroids, and ascidians. Fish, mostly *Pagothenia bernachii* and *Trematomus centronotus*, are also conspicuous. The lower zone, below 33 m and down to approximately 200 m is characterized by a diverse community of sponges. Infaunal communities (i.e., those living within the sediment) are dominated by polychaetes (worms) and crustaceans; mollusks, unlike temperate areas contribute less than 5% of the total number of individuals. The number of species found is relatively low, usually less than 100, but the number of individuals can be enormous with over 155,000 individuals per meter square found in McMurdo Sound.

A characteristic of Antarctic benthic communities is the high proportion of suspension feeders. This is thought to be a response to the short period of phytoplankton availability during the short summer ice-free period. This has also led to

an increasing prevalence of scavenging and opportunistic feeding strategies compared to elsewhere.

Penguins and Other Sea Birds

The abundance of birds in many areas of Antarctica gives a deceptive impression of their biodiversity. Thirty-eight bird species are known to breed south of the Polar Front but only ten actually breed on the Antarctic continent. This level of diversity is very low compared to most other natural areas. Penguins comprise approximately 90% of the Antarctic avian biomass. Four species, the Emperor (*Aptenodytes forsteri*), Adelie (*Pygoscelis adeliae*), Gentoo (*Pygoscelis papua*), and Chinstrap (*Pygoscelis antarctica*) breed on the Antarctic continent, but of these the Gentoo and Chinstrap Penguins are restricted to the Antarctic Peninsula area. The Emperor Penguin is the largest of the penguins. It alone of all the vertebrates winters on the continent. All other birds and mammals migrate north for the winter. The Adelie Penguin is the most abundant penguin and colonizes areas of bare rock on the coast or on nearby offshore islands. Colonies can be up to several hundred thousand in size. All penguin diets are dominated by krill, although squid and fish supplement the diets to varying degrees depending on location and seasonal abundance.

Most of the remaining birds breeding on the Antarctic continent belong to the petrel family (Procellariiformes). These include the Wilsons Storm Petrel, Snow Petrel, Antarctic Petrel, Cape Petrel, Antarctic Fulmar, and Giant Petrel. The Southern Skua, a large gull-like bird is an aggressive predator at many bird colonies. Although rarely taking adult birds, it feeds on eggs, chicks, and sick individuals. The only other species is the Blue Eyed Cormorant, which is restricted to northern areas of the Antarctic Peninsula.

Bird distribution patterns and breeding cycles are strongly influenced by the short intense pulse of marine primary production between December and February. As this is the only time when food is abundant, the breeding cycle of all birds, except the Emperor Penguin, must be rushed to coincide with maximum food availability. It is essential that the chicks are fledged by the end of summer so they can migrate north before the growing sea ice prevents them access to their food supply.

Seals and Whales

Six species of seals are found around Antarctica, the Weddell Seal, the Ross Seal, the Crabeater Seal, the Leopard Seal, the Antarctic Fur Seal, and the Elephant Seal. However, of these only the Weddell Seal actually breeds on the Antarctic continent. This species, which breeds close to the continent on the fast ice and feeds mostly on fish, squid, and krill, forms groups of up to several hundred. Leopard Seals, together with the Killer Whale, are the main top-order predators. These species both feed on penguins and other seals. The Ross and Crabeater seals (the world's most abundant large mammal)

spend their entire lives in the pack ice and never venture to shore. The Elephant and Fur seals are mostly creatures of the sub Antarctic and only juveniles or vagrants make appearances on the Antarctic coast.

Twelve species of toothed whale and seven species of baleen whale are present in the Southern Ocean. None are endemic to the region and most have widespread migration patterns. As a result of the whaling impact most of the larger whales are uncommon, with Blue Whale numbers probably down to only a few hundred. However, the smallest of the baleen whales, the Minke Whale, was never seriously harvested and is possibly more abundant than ever. This species is common in coastal areas. All baleen whales feed on zooplankton, predominantly on krill.

Of the toothed whales, the Killer Whale is the most abundant. It is common throughout the pack ice zone and when the ice retreats from the coast in summer, is also a common visitor to coastal areas. It feeds on most larger animals including fish, penguins, and seals. There are no known human deaths attributable to Killer Whales although there are numerous stories of near misses.

The ecosystem of the Antarctic is often seen as pristine and untouched but this is not the case. The whaling and sealing industries of the nineteenth and early twentieth centuries massively depleted the numbers of top-order predators and effectively removed an important component of the ecosystem. Similarly, in the 1970s, over-fishing led to the commercial extinction of most of the more abundant fish species. National bases have degraded areas of the coastline and dumped waste into the nearby ocean. PVCs, DDT, and other toxic compounds have been found in the tissues of most animals. Introduced diseases (e.g., Newcastle disease, canine distemper) have caused widespread deaths among bird and seal populations. The ozone hole and global warming threaten to make even more significant changes. In spite of all these factors, Antarctica is still one of the least impacted ecosystems.

Antarctic Geomorphology

Antarctica is simultaneously both the highest and driest continent on earth. It covers an area of approximately $14 \times 10^6 \text{ km}^2$ and has a coastline approximately 32,000 km long. The average elevation is over 2,300 m but this is largely made up of an ice cap that is over 4,000 m thick near its center close to the south Pole. The ice cap contains 90% of the world's freshwater reserves and covers more than 95% of the continent. The remaining ice-free areas are mostly small, scattered, and located near the coast. The Antarctic coastline itself is predominantly ice bound. Over 38% of its length is comprised of ice cliffs where the ice cap terminates abruptly at the sea. A further 44% of the coast is associated with floating ice shelves, large areas where the polar ice cap extends from the continent and floats on the sea. The largest

of these, the Ross Ice Shelf, covers an area of 536,000 km²; an area larger than the U.S. state of Texas or the country of France. About 13% of the coast is occupied by glaciers and only 5% of the coastline is ice-free and even in these areas sea ice covers the sea surface for up to 11 months of the year.

Much of the land surface of Antarctica is currently below sea level, with the ice cap grounding at depths of up to 1,200 m. Were the ice cap to melt, the land surface would rebound by up to several hundred meters but more than 50% of the current land surface would still be under water and the Antarctic Peninsula would become a chain of islands.

Geological investigations of Antarctica have been closely associated with its exploration history. Of particular interest was the position of the South Magnetic Pole. The British explorer, James Clarke Ross, had this objective in mind when he entered what is now known as the Ross Sea in 1841. However, as he sailed southeast he was confronted by the Transantarctic Mountains blocking his way to the Magnetic Pole. Disappointed, he sailed south reaching 76°S and in the process discovering amongst other things the Ross Ice Shelf and Mount Erebus. Most of the subsequent expeditions of the late nineteenth and early twentieth centuries, which often had funding and backing from scientific associations such as the British Royal Society or Royal Geographical Society, took geological and geographical scientific parties along. Both of Robert Falcon Scott's expeditions included geologists as did those of Shackleton, Mawson (himself a geologist and also searching for the South Magnetic Pole), Drygalski, and others. On Scott's fateful return trip from the Pole, the explorers never considered discarding their heavy load of geological specimens even though they probably contributed to their demise. The single most important event in the scientific understanding of Antarctica was the International Geophysical Year in 1957. This led to the establishment of a large number of permanent bases on the Antarctic continent with the specific aim of investigating the structure and evolution of Antarctica.

The present shape of the Antarctica continent is profoundly influenced by its geological past. Antarctica was an integral part of the ancient supercontinent of Gondwana. This continent started to break up in the Jurassic (approximately 200 million years ago) but the last major segments, that is, Australia and South America, did not finally break from Antarctica until the Oligocene, approximately 40 million years ago. For virtually all of this time Antarctica was situated over the South Pole. The northward movement of Australia and South America away from Antarctica led to the development of Drake Passage beneath South America and the Macquarie Passage beneath Australia. This opening ocean led to the progressive thermal isolation of Antarctica and the onset of continental glaciation. As the Southern Ocean widened, Antarctica continued to cool and a permanent ice cap

developed, a feature of Antarctica that has been present ever since. The stability of the ice cap, however, has varied and there is now strong evidence for major periods of ice cap shrinkage and expansion through the Pliocene and early Pleistocene (5.1 million years ago).

The erosive and depositional action of ice is almost entirely responsible for the topography, landforms, and continental shelf bathymetry of Antarctica. While melt water streams are seasonally present in some areas they have been inconsequential in either landscape evolution or marine processes. In Antarctica, ice either covers the surface or is responsible for its features. Less than 5% of the surface area of Antarctica is ice-free. These ice-free areas are small and clustered around the coast. The largest of these is the Dry Valleys of the McMurdo Sound area followed by the Bunge Hills and Vestfold Hills of eastern Antarctica. These areas typically show many of the classical features of recently glaciated terrains, including moraines, striated surfaces, drumlins, etc. Many of these areas are also characterized by rich lake provinces. The Vestfold Hills, for instance, an area of only approximately 400 km² contains more than 200 lakes, many of which are hyposaline reflecting a marine origin and subsequent stranding by isostatic sea-level rise. Lakes in the Vestfold Hills area are less than 12,000 years old but some lakes in the Dry Valleys may be as old as 500,000 years. Some of these lakes are permanently ice covered (e.g., Lake Hoare and Lake Fryxell in the Dry Valleys), while some are so saline that they never freeze (Deep Lake, Lake Stineer, Vestfold Hills). One of the more unusual features of many of these lakes is that they are permanently stratified (meromictic), that is the water column is layered with zones of increasing salinity with depth. These layers have typically been undisturbed for thousands of years.

Soft sediment shores are uncommon around Antarctica. This is largely because of the absence of significant river action. Small local deltaic features are sometimes present but these are always of very limited extent. Sandy beaches are particularly uncommon in Antarctica as there is rarely sufficient water action to provide the necessary sorting.

The continental shelf of Antarctica is generally narrow, varying in width between 64 and 240 km wide. However, its most distinguishing features are its irregular topography and its great depth, which averages over 500 m; a depth which is more than five times the global average. The outer shelf is typically 400–500 m depth but, unusually, has a landward sloping profile that deepens toward the coast. Depths of 100–1,200 m are common close to the coast in many areas around Antarctica. The reasons for this unusual profile have been speculated on since the late 1920s. It has been shown that glacial isostasy, that is, the effect of the heavy ice cap depressing the crustal surface, can account for up to 150 m of this depth but the remainder is almost certainly due to the

effects of glacial erosion at times of low sea-level stand and glacial advance. In some areas during periods of lower sea level, glaciers carved channels far out onto the continental shelf. These drowned fjords are often beneath the effects of modern iceberg scouring and contain basins with restricted circulation. This has led to the preservation of very well-preserved late Quaternary sequences that have a remarkably good time resolution. Such sequences can be found off the Antarctic Peninsula and on the MacRobertsonland Coast. These glacial troughs are usually associated with modern major ice streams and outlet glaciers and they are likely to be the result of multiple glacial advances. Geomorphic landforms associated with these troughs include megascale glacial lineations, striations, and drumlins.

Sedimentation on the continental shelf is dominated by biogenic and glaciomarine processes. Biogenic sediment is comprised of the remains of animals and plants, predominantly from the plankton, while glaciomarine sediment is sediment-derived originally from glaciers but subsequently deposited in the ocean. In offshore areas up to 50% of the sediment can be of biogenic origin. This is overwhelmingly composed of diatom remains. Diatoms are microscopic plants, 2–100 [μm] in diameter, with siliceous cell walls. Diatom diversity in these sediments is relatively low due to extensive silica dissolution, a factor that becomes more pronounced with increasing water depth. Dominant species include *F. curta*, *F. cylindrus*, *Fragilariopsis kerguelensis*, *Thalassiosira lentiginosa*, and *Eucampia antarctica*. Other biogenic components include minor quantities of silicoflagellates, Parmales, and foraminera. The remaining sediment is predominantly of glaciogenic origin, although a small component is also of eolian origin.

Around Antarctica, icebergs influence sedimentation up to a distance of several hundred kilometers from the continent. Sediment entrained in a basal debris zone is transported towards the coast. Sediment enters the sea by one of three principal processes; either at ice shelves, ice cliffs, or via glaciers. Ice at the edge of ice shelves calves into icebergs, which drift out to sea and drop their sediment often far from the coast. Most of the sediment, however, is deposited in the grounding zone (i.e., where the ice shelf begins to float). Ice at ice cliffs is eroded and drops its sediment load in close proximity. Most sediment reaches the coast by glaciers, which calve rapidly into small sediment-laden icebergs. It has been estimated that 62% of the ice transported to the coast of Antarctica is calved as icebergs from ice shelves, 22% from glaciers, and 16% from ice cliffs (Drewry and Cooper 1981).

Most of the continental shelf, at least down to 400 m, is affected by icebergs. These regularly plough into the sediment, disrupting any bedding and causing the sediments to become well mixed and unlaminated. These sediments are sometimes referred to as iceberg turbate. Calcareous

sediments are rare on the Antarctic continental shelf but have been recorded on the shelf break in Prydz Bay. Here the dominant component is barnacle shells.

In Antarctica glaciers are by far the most important contributors of terrestrial sediment to the marine environment. Compared to many glaciers in temperate or Arctic areas, they are predominantly sluggish-moving, very cold, and usually extend out to sea as a floating ice tongue. Most Antarctic glaciers are classified as “cold glaciers,” that is the temperature of the ice is below the pressure melting point. This results in a glacier that is dry and frozen to the bed. On the Antarctic Peninsula, where the temperatures are higher and the precipitation is much greater they are more dynamic and the glaciers extend to near the mouths of short fjords. These glaciers which often undergo substantial melting in summer are often classified as “subpolar” glaciers. In cold glaciers, where the ice is frozen to the bed, there is comparatively little erosion of bedrock or deposition of sediment. Furthermore, there is rarely any glaciofluvial sediment.

Glaciers enter the sea via fjords, deep usually linear embayments with shallow sills at their entrance. In most temperate areas the glaciers have retreated leaving a deep marine embayment. In Antarctica, however, the glaciers mostly reach the sea and then extend as a floating ice tongue. Minor glaciers are present in areas where the ice cap has retreated such as in the Vestfold Hills and Dry Valleys. The world’s largest glacier is the Lambert Glacier in eastern Antarctica. This glacier drains approximately 25% of the eastern Antarctic ice cap and occupies a trough up to 80 km wide and 800 km long. This glacier has retreated in the geological past and would have left the world’s largest fjord.

Global warming resulting from the current accumulation of greenhouse gases is already having a major impact on Antarctic glaciers and ice shelves. Temperatures over the Antarctic are predicted to rise faster than most other places on earth. This is leading to an increase in the rate of disintegration of major ice shelves, particularly evident on the Antarctic Peninsula. A major focus of recent research has been to establish the natural limits of Antarctic climate fluctuations in the recent and more distant past. Minor warming and cooling trends of a few degrees have been identified since the end of the last ice age (i.e., over the last 10,000 years) which have led to small scale advances and retreats of the ice cap and glaciers. However, during the Pleistocene (i.e., 1.6 million to 10,000 years) the ice cap at times extended to the edge of the continental shelf. Periods of increased temperature, comparable to those predicted for Antarctica, have been identified in the Pliocene (i.e., 1.6–5 million years ago). During these times there is strong evidence of major ice cap retreat; indicating a possible warm future for Antarctica.

Further recommended reading is given in the following bibliography.

Cross-References

- [Arctic, Coastal Ecology](#)
- [Arctic, Coastal Geomorphology](#)
- [Glaciated Coasts](#)
- [Ice-Bordered Coasts](#)
- [Paraglacial Coasts](#)

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Aquaculture

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Definition and History

Aquaculture can be most simply defined as underwater agriculture. It is the rearing of aquatic plants and animals through some type of intervention by humans. Aquaculture is conducted in freshwater and saltwater. The term mariculture is often used in relation to marine aquaculture. Aquaculture involves plants and animals produced for human food, as ornamentals, for bait, and in recent years, as sources of nutritional supplements and pharmaceuticals. Some plants and animals are also grown as foods for other aquaculture species. For example, shrimp hatcheries typically grow one or more species of algae to feed brine shrimp (*Artemia salina*), which are fed upon by larval shrimp.

While there is some production of rooted aquatic plants, echinoderms (e.g., sea cucumbers), tunicates (e.g., sea squirts), amphibians (frogs), and reptiles (turtles and alligators), the majority of the world's aquaculture production comes from seaweeds, molluscs (e.g., abalone, clams, oysters, scallops), crustaceans (e.g., shrimp, crawfish), and finfish.

Aquaculture is widely acknowledged to have been born in China, perhaps as long as 4,000 years ago (Avault 1996) when carp culture was developed. The first written document on aquaculture, a very short book by Fan Li, entitled *Fish Breeding*, appeared in 475 BC in China (Borgese 1977). Carp culture was also developed in Europe, possibly during the Middle Ages. Tilapia, a type of fish native to north Africa and

the Middle East, is depicted on the tombs of the Pharaohs in ancient Egypt. Whether tilapia was being cultured or not is unclear. Oyster culture appears to have been practiced during the days of the Roman Empire, while shrimp were being produced in China by AD 730 (Borgese 1977). Native Hawaiians constructed fishponds and apparently developed a primitive form of aquaculture at least several hundred years ago. Thus, aquaculture in one form or another has a long history.

By the mid-nineteenth century, trout culture had become a well-developed art in both Europe and North America (Kirk 1987; Stickney 1996). In the United States, brook trout were being produced and at least a few commercial fish farms had been established.

Overfishing in the United States was first reported in the 1750s, but it was not until 1871, when Spencer F. Baird convinced Congress to create the US Fish and Fisheries Commission (Stickney 1996), that any attempt to rebuild wild fisheries was made. Baird enlisted the best fish culturists of the time and put them to work learning how to spawn an array of fishes and invertebrates with the goal of restocking both the inland and marine waters of the nation. In addition, the Commission shipped fish to foreign countries (e.g., establishing populations of rainbow trout in Europe and both rainbow trout and chinook salmon in New Zealand). Fish such as brown trout and common carp were introduced to the United States by the Commission.

While in most cases the stocking of eggs or newly hatched larvae by the Commission did little more than provide food for wild fishes, a significant amount of information and technology were developed, remnants of which can be seen in the most up-to-date of modern hatcheries. Techniques were developed for successfully producing fingerlings of many recreationally and commercially important species, including largemouth bass, channel catfish, walleye, striped bass, and Pacific salmon to name but a few.

The Commission was responsible for producing and distributing the eggs and larvae of hundreds of millions of fishes and invertebrates. Attempts were made to establish Atlantic salmon on the Pacific coast and Pacific salmon in the waters of New England. Both attempts failed, though not because of a lack of effort. Millions of eggs were shipped over a number of years to accommodate the program. Pacific salmon were ultimately established in the Great Lakes, but not until the 1980s.

Ultimately, attempts to enhance marine fisheries through stocking by the Commission and its successor organizations were terminated, though new enhancement programs, largely by the various individual states, have been initiated in recent years. Stocking programs for inland waters, by both state and federal agencies have continued.

Modern commercial aquaculture can arguably be attributed to have its origins in one or more of several countries and decades within the twentieth century. For instance, breakthroughs that made possible the development of commercial

shrimp culture occurred in Japan during the 1930s, though the industry did not develop significantly until the 1980s after which considerably more information had been developed in Japan, Taiwan, the United States, and other nations; information that was critical to move the technology beyond the research and demonstration phase.

The late 1960s and the 1970s serve as a benchmark period during which commercial aquaculture developed rapidly and began making significant contributions to the world's foodfish supplies. Fish farmers in the southern United States initially produced buffalo (*Ictalurus* sp.), but soon turned to channel catfish (*Ictalurus punctatus*). While the potential for rearing catfish to food size profitably was demonstrated in the 1950s (Swingle 1956, 1958), commercial activity did not expand from producing and selling catfish fingerlings to growout of foodfish until the mid- 1960s (Wellborn and Tucker 1985).

During the same two decades (1960s and 1970s), commercial craw-fish culture developed in Louisiana and rainbow trout production increased dramatically, particularly in the Thousand Springs area of Idaho (Stickney 1996).

Marine shrimp research with various species in the former genus *Penaeus* (which has recently been split into several genera) was being conducted in both the United States and Taiwan, which led to two different types of larval rearing systems, both of which have been employed extensively. Taiwan and Japan were leading shrimp producing nations in Asia until the 1980s when Thailand, China, Indonesia, the Philippines, India, and others developed active industries. In the Western Hemisphere, the shrimp industry began in the 1980s and was centered in Ecuador, which continues to dominate Latin American production, though many other nations are now involved. The 1970s saw a great deal of interest in the culture of freshwater shrimp, *Macrobrachium rosenbergii*, but for various reasons, there are very few farms currently producing that species.

Also, during the 1980s, the feasibility of producing salmon in net-pens placed in the marine environment was demonstrated (Stickney 1994). Much of the initial work was conducted on the west coast of the United States with coho (*Oncorhynchus kisutch*) and chinook (*Oncorhynchus tshawytscha*) salmon, and some commercial production occurred in both the United States and Canada. However, it soon became apparent that Atlantic salmon (*Salmo salar*) was more amenable to culture and the industry concentrated on that species, with Norway becoming the dominant salmon-growing nation. Chile became a major salmon-producing nation in the 1990s. Canada continues to produce significant quantities of salmon, and Maine has become the dominant salmon producing state in the United States, though there is still some production in the state of Washington. Large hatchery programs to produce Pacific salmon for release into the wild continue in the Pacific Northwest.

Many other aquatic species have been commercially produced within the past few decades. At present, there are well over 100 species under culture around the world, with more being developed each year. In Europe, sea bass (*Dicentrarchus labrax*) and sea bream (*Sparus aurata*) are being produced in the Mediterranean region, while Atlantic halibut (*Hippoglossus hippoglossus*) are cultured in Norway and Scotland along with the previously mentioned Atlantic salmon. Plaice (*Pleuronectes platessa*) and sole (*Solea solea*) are also among the fishes cultured in Europe.

Popular marine fishes in Japan are salmon (*Oncorhynchus* spp.), red sea bream (*Pagrus major*), and yellowtail (*Seriola quinqueradiata*). Milkfish (*Chanos chanos*) are produced primarily in the Philippines, Thailand, and Indonesia.

Tilapia (*Oreochromis* spp.) are being reared in both salt and freshwater throughout the tropical world and indoors in some temperate areas. Their fast growth and excellent flavor make them excellent aquaculture species.

Among the invertebrates being reared around the world are oysters (e.g., *Crassostrea* spp. and *Ostrea* spp.), including of course, pearl oysters, mussels (primarily *Mytilus* spp.), abalone (e.g., *Haliotis* spp.), and scallops (e.g., *Pecten* spp. and *Patinopecten yessoensis*).

Then there are the plants, dominated by red and brown seaweeds, but including green seaweeds (all of which are forms of algae), but also including higher plants such as water chestnuts, watercress, and such ornamental plants as water lilies, to name but a few. A wide variety of microscopic single-celled algae are produced as food for larval stages of aquacultured animals and, in the case of species such as *Spirulina*, as nutritional supplements for humans (*Spirulina* is also used as a food additive to produce color in fish). Extracts from aquatic plants are also being developed into pharmaceutical products.

Global Production

Seaweed aquaculture comprises nearly one-quarter of total world production (New 1999). Seaweeds play a large role in Asian cuisine, particularly that of Japan. Extracts from seaweeds, including agar and carageenin, are components of products utilized by people around the world. Seaweed extracts are used in automobile tires, toothpaste, ice cream, pharmaceuticals, and a wide variety of other commonly employed products.

By 1996, total world aquaculture production exceeded 34 million metric tons (mmt), with nearly 11 mmt coming from seaweeds (FAO 1998). A breakdown of production by type of organism is presented in Table 1.

The most widely produced finfish are various species of carp, with the vast majority of that production occurring in China. In fact, China was responsible for 45.7% of the world's

Aquaculture, Table 1 World aquaculture production for various groups of organisms in 1996 (data from FAO 1998)

Organism group	Production (mmt)
Freshwater finfish	14.4
Diadromous finfish ^a	1.7
Marine finfish	0.6
Crustaceans	1.1
Molluses	8.5
Seaweeds	7.7
Miscellaneous ^b	0.1
Total	34.1

^aIncludes fishes that spend part of their life cycle in freshwater and part in saltwater or that can live in either medium throughout their lives. Examples are salmon, eels, and milkfish

^bIncludes amphibians, reptiles, tunicates, and other minor species

Aquaculture, Table 2 World aquaculture production in 1996 by continent and the top 10 nations in terms of total aquaculture production (data from FAO 1998)

	Percentage of world total
Continent	
Asia	88.9
Europe	6.0
North America	2.3
South America	1.6
Oceania	0.3
Former USSR	0.4
Africa	0.4
Nation	
China	45.7
India	7.4
Japan	7.0
Republic of Korea	4.5
Indonesia	3.6
United States	3.6
Philippines	3.2
Taiwan	2.8
Spain	2.6
France	2.2

aquaculture production, excluding seaweeds, in 1996. When Asia is compared with other regions, the dominance of that continent in terms of aquaculture production is even more impressive (Table 2). Comparison of the top 10 countries in terms of aquaculture production also demonstrates the dominance of China, but also the important contributions of other Asian nations (Table 2).

The world's capture fisheries peaked and now hover around 80 mmt, while demand continues to increase. Aquaculture expansion has been identified as the only means of meeting that demand, and currently available data show that aquaculture production has rapidly increased in recent years.

New (1999) developed a comprehensive overview of global aquaculture production that includes a comparison of data between 1987 and 1996. Production over the period covered increased from 13.4 mmt in 1987 to 34.1 mmt in 1996. That rate of expansion cannot be maintained indefinitely, however. Many of the best places to conduct aquaculture are currently being utilized, while various constraints mitigate against aquaculture development in regions where climate might be conducive to good growth of aquatic species. Water supply and quality are constraints in many areas. In the coastal zone, the high cost of land and competition with other users often mitigate against using it for aquaculture.

One approach is to establish fish farms in the open ocean. The costs and logistics associated with that idea present significant problems, but at least some success has been realized. Enhancement stocking, first undertaken by the US Fish and Fisheries Commission over a century ago, is currently being re-evaluated. Revitalization of the red drum (*Sciaenops ocellatus*) fishery in Texas through an aggressive stocking program coupled with strict fishing regulations (including a ban on commercial fishing) provide an indication that such programs have potential and may be cost effective. One approach would be to pay for the programs through license fees, which would apply to commercial and/or recreational fishermen depending upon the species employed.

Much of the increase in aquaculture production over the years, at least in developed nations, can be attributed to increased knowledge that has led to improved technology. One example is the channel catfish industry in the United States. When the industry was in its infancy, a profit could be made if the fish were sold at \$1.10 per kg. Forty years later, a profit can still be made when catfish prices to the producer are in the range of \$1.65–1.80 per kg. The reason that catfish farmers can still realize a profit given the small increase in price per unit weight over nearly a half century of production is related to increases in production per hectare, which is a function of improvements in management of water quality, improvements in the food provided, and to a lesser extent, selective breeding programs.

In the 1950s, when catfish farming was in its infancy, a production level of about 500 kg/ha was considered credible. By the late 1960s that had increased to 1,500 or even 3,000 kg/ha. Currently, some catfish farmers produce as much as 10,000 kg/ha in ponds. Without those increases in production, the catfish industry would have probably collapsed.

Culture Systems

The vast majority of aquaculture species are grown in ponds, though a number of other types of culture systems are also in use.

They include linear and circular raceways (the latter are also known as tanks), cages, and net-pens. Specialized culture systems for molluscs and seaweeds include bottom culture in bays, pole culture, net culture, basket culture, and raft culture (Stickney 1994).

Culture systems are often classified as extensive or intensive, though in reality the intensity of culture occurs along a continuum and relates to the complexity of the culture system. An example of a highly extensive system would be the culture of oysters in natural waters by spreading shell (cultch) material on the bottom and allowing larval oysters (spat) to settle and grow. Very little intervention by humans, other than preparation of the bottom and the spreading of shell is involved.

At the other extreme are high-intensity closed-recirculating water systems. Such systems employ raceways as culture chambers. Water is continuously exchanged in each raceway. Exiting water is treated, at a minimum to remove solids (usually by settling or mechanical filtration) and convert ammonia (NH_3 or NH_4^+) and nitrite (NO_2^-) to less toxic nitrate (NO_3^-) in a unit called a biofilter. Additional components that maintain temperature in the desired range, remove foam, convert nitrate to nitrogen gas (N_2), and remove other nutrients, may also be present. Most closed systems, receive supplemental aeration to maintain a high level of dissolved oxygen in the culture chambers. In truly closed systems, new water is added only to replace that lost to splashout, evaporation, or to flush solids from the system. Systems that have the same basic components but in which some percentage of the water is routinely or continuously changed by adding new water, are called semi-closed.

Between the extremes are the other systems mentioned above. Culture ponds, unlike most farm ponds, are most commonly rectangular in shape, are usually no more than 2 m deep, have a drain structure of some type incorporated into them and are, in most cases, located in an area where there is a reliable source of water. While some culture ponds depend on rainwater runoff to fill them, most ponds, at least in the developed world, employ well water or water from a stream, lake, or reservoir for filling and maintenance. Culture ponds come in many sizes, but most used for commercial production range from about 0.1–10 ha, with those between about 1 and 5 ha being the most commonly seen. Small ponds are used most often for breeding and rearing of early life stages, while the growout commonly occurs in larger ponds.

If two or more species are reared in the same culture unit—a process known as polyculture—that unit is most likely a pond. The Chinese have a very long history of carp polyculture, which involves different species with different feeding habits. Different species are stocked to consume phytoplankton, zooplankton, benthic invertebrates, and rooted aquatic plants. Benthic feeding carp may specialize on worms and insect

larvae or molluscs. Today, the Chinese may also provide agricultural byproducts and prepared feeds. They, like aquaculturists in many other countries, maintain a high level of pond fertility to encourage the growth of natural food organisms. Cattle, swine, and poultry manure are common sources of fertilizer. Livestock are routinely seen being reared adjacent to ponds into which they deposit fresh manure. Rearing livestock adjacent to fishponds is common in many countries, particularly in the developing world.

The majority of raceway systems are of the flow-through variety. That is, water is continuously exchanged. Depending on stocking density and water quality, the exchange rate may range from several times an hour to several times a day.

Cages are relatively small structures with ridged walls, commonly made of welded wire or rubber-coated wire. Cages are used to contain fish in water bodies that are not enclosed (bays, rivers), and in those which are too large, too deep, or have some physical impediment to harvest (such as standing dead timber). Many reservoirs and lakes are good examples. Cages are not usually more than one to a few cubic meters in volume. Cages may be taken to the shoreline to facilitate harvesting.

Net-pens are large cages where nylon netting is employed for the sides and bottom instead of ridged material. The standard design most commonly used in protected marine waters where the net-pens are not exposed to high waves is typically $20 \times 20 \text{ m}^2$ on a side and can be 20 or more meters deep. Both cages and net-pens are provided with floats to keep them at the surface. Cages usually sit low in the water and are fitted with tops, while the sides of net-pens typically extend above the water surface sufficiently to prevent fish from jumping over the open top. Walkways around the perimeter of each net-pen allow access for feeding and harvesting.

Recently, there has been a move toward rearing fish in offshore locations. Net-pens designed for the open ocean are usually fully enclosed by netting because they may be subjected to storms. Some are designed so they can be totally submerged throughout the growing cycle or during stormy periods.

Oysters, mussels, clams, and scallops are less susceptible to predation by starfish and other organisms if they are grown off the bottom. Poles, lines, and baskets have been used in conjunction with off-bottom culture. In Spain, for example, ropes to which oysters or mussels are attached are suspended from large rafts.

Enhancement, as previously mentioned, is an option that deserves further consideration. While the culturists associated with the US Fish and Fisheries Commission in the latter half of the nineteenth century failed in their efforts to increase the numbers of fish in the sea, advances since that time greatly increase the potential for success.

Hatchery technology has developed to the point where the larvae of many species of interest can be reared from egg to

appropriate stocking size economically. Enhancement efforts need to be conducted in an environmentally sound manner, however. Fish and invertebrates should not be stocked in such numbers that they overwhelm the available food supply; they should not cause declines in the populations of other valuable species in the community; and their genetic composition should reflect that of wild populations, to name but a few considerations.

Getting Started

Several steps should be taken in advance of actually establishing an aquaculture facility. First, the prospective culturist should determine what species will be produced and type of water system to be employed. A suitable location should also be identified. All three factors tend to be co-dependent. Sometimes, the prospective aquaculturist will own property selected as the site of the new facility. If that property is in a subtropical area and there is no suitable source of cold water, it would be undesirable to attempt to produce trout or salmon. Similarly, if the property is distant from the coast and there is no convenient source of saline groundwater, producing marine fish might not be wise. The exception to those examples might be employment of a closed water system within which temperature and salinity could be controlled. Economic analyses of each option under consideration should be conducted to help determine the best course of action.

Once the above matters have been resolved, it is necessary to produce a business plan. Such a plan will be required by lending institutions and venture capitalists. The plan should include a complete financial analysis showing all costs associated with constructing, equipping, and operating the facility over a period of several years, along with an estimate of profit or loss for each year projected. All estimated costs and income should be based on information that can be documented. The culturist should provide a listing of the permits that may be required for the facility as well as a timetable for obtaining those permits and the fees that are associated. Conceptual drawings, or even better, detailed engineering blueprints, should be included with the business plan.

The prospective aquaculturist should determine what, if any, local, state, or national governmental permits are required. In some instances, the permitting process can be long and arduous. It may also be costly, particularly if the applicant is required to battle opponents within the legal system and conduct an extensive environmental impact study prior to establishing a facility. In general, inland aquaculture sites and those located along the coast that employ recirculating technology or other water systems that do not impact the marine or estuarine environment are more easily permitted than those that are established in public waters.

Facility Management

On the surface, aquaculture is a biological discipline. In reality, like terrestrial farming, aquaculture involves a variety of sciences as well as engineering, business management, and such skills as plumbing, welding, carpentry, and electrical wiring. In the science arena, while expertise in biology is critical on any fish farm, an understanding of chemistry may be even more important. As the size and intensity of aquaculture operations increase, so does the need for additional and better-trained personnel. A subsistence culture operation in a developing country might be operated by an individual or a family, with perhaps a few additional laborers. A large pond operation that involves a hatchery, larval rearing, and production facilities, or an intensive raceway operation may require a large number of unskilled laborers as well as employees proficient in various specialties. Some of the technical aspects associated with operation and management of an aquaculture facility are discussed in the following subsections.

Water Quality

Maintenance of good water quality is critical in any aquaculture facility. There are virtually hundreds of thousands of chemicals that can be found dissolved in water. In addition, there are physical characteristics such as temperature and transparency that can affect performance of organisms in an aquaculture facility. It is neither possible nor necessary to monitor all aspects of water chemistry. In most culture systems, after an initial screening of the water supply for toxicants (e.g., herbicides, pesticides, high levels of trace metals), routine monitoring involves only a few variables. Water temperature and dissolved oxygen, and in marine systems, salinity are typically measured at least several times a week and records of them are maintained. Some culturists employ sensors that constantly monitor those parameters and download the information into computers. Water transparency monitoring is also conducted in pond systems and provides a simple means of determining the state of plankton blooms. This is particularly important in systems that employ fertilization to provide natural food.

In high intensity systems, ammonia may be routinely monitored. Nitrite levels might also be tracked, particularly during colonization of biofilters with bacteria, since the types of bacteria that convert nitrite to nitrate tend to lag behind those that convert ammonia to nitrite. Thus, if only ammonia is measured, the culturist may get the impression that the water is not toxic, when in fact, it may contain a high level of nitrite. Toxic nitrite levels have occurred in heavily stocked catfish ponds in the United States during the late summer, so routine monitoring during that critical period is often undertaken. While ponds are not equipped with biofilters, the bacterial conversion of ammonia to nitrite and nitrate does occur.

Water temperature is of critical importance because it controls metabolic rate in the species being cultured. Aquatic animals can often be categorized as being warmwater or coldwater species. Warmwater animals are those that have an optimum temperature for growth of 25 °C or higher, while coldwater species have an optimum below about 15 °C. There are some animals, such as walleye (*Stizostedion vitreum vitreum*) and northern pike (*Esox lucius*) that have an optimum between 15 °C and 25 °C. Such animals are sometimes referred to as midrange species.

Most species have a broad tolerance for environmental temperature. Many species of fish live over a range of temperatures from near 0 °C to somewhat above 30 °C. Channel catfish and largemouth bass (*Micropterus salmoides*) are but two examples. Many coldwater species, such as trout and salmon, cannot tolerate warm temperatures, while many tropical fishes, such as tilapia, die at temperatures where trout and salmon thrive. Tilapia grow most rapidly when the water temperature is about 30 °C, show significantly reduced growth at about 25 °C, and stop growing around 20 °C. Below the latter temperature tilapia lose their disease resistance and will die when temperatures fall much below 15 °C (Avault and Shell 1968).

Terrestrial animals live in an atmosphere that contains some 20% oxygen, while aquatic species live in a medium that contains only a few parts per million (ppm) of the same essential element. Depending on species, the minimum level of dissolved oxygen in the water that should be present to ensure that the animals have a sufficient supply varies from about 3 ppm (e.g., for catfish) to 5 ppm (e.g., for salmon and trout). From a practical standpoint, it is always desirable to have dissolved oxygen at saturation.

The amount of oxygen that water can hold (the saturation level) varies with temperature, salinity, and altitude. As each of those variables increases, the ability of water to hold oxygen decreases. Thus, warm, saline water at high altitude would hold less oxygen than cool, freshwater at sea level. In virtually all locations where aquatic animals are cultured, dissolved oxygen saturation level reaches or exceeds 5 ppm.

Dissolved oxygen follows a diurnal (daily) cycle. During daylight hours, phytoplankton and other types of vegetation in the aquatic environment produce oxygen as a result of photosynthesis. As a result, the oxygen level in water tends to increase during the daytime. Oxygen is also introduced into water through diffusion from the atmosphere, a process that is enhanced through wind turbulence or mechanical aeration. At night, both plants and animals respire, thereby consuming oxygen during a period when no photosynthesis is occurring (though diffusion is still operating). The result is a cycle in the availability of oxygen in the water. The cycle is also influenced by weather conditions and season of the year. During cloudy days, the rate of photosynthesis is reduced (because of reduced light level), while during clear weather

the rate of photosynthesis is accelerated. Recall that temperature plays a role in the saturation level of dissolved oxygen in water; thus the seasonal variation. In most cases, problems associated with low dissolved oxygen occur during warm weather after a period of cloudy days. This often occurs during the late summer or early fall in temperate areas (the time just prior to harvest).

Aquaculturists routinely measure dissolved oxygen in pond systems early in the morning, particularly during the time of year when depletions are likely. Dissolved oxygen level typically peaks in the late afternoon before dusk and gradually falls during the night, and begins to increase after dawn. The lowest dissolved oxygen level will usually occur just prior to dawn, so the aquaculturist needs to check conditions at that time and determine which, if any, ponds may be approaching critically low oxygen levels.

When oxygen levels are low, the culture animals may alert the observant culturist. Fish may swim at the surface with their mouths open. They appear to be gulping for air. Shrimp may climb above the water level on rooted plants to expose their gills to the atmosphere. By the time such signs are noted, the animals may have already been severely stressed. It is much better to routinely measure dissolved oxygen during the night and take action to prevent severe depletions before they occur.

Under moderate stocking regimes only a few ponds on a given aquaculture facility will experience oxygen depletions on the same day and it is not necessary to have each pond provided with emergency aeration equipment. In many cases today, however, stocking densities are so high that oxygen depletions may occur with some frequency in nearly every pond. Such facilities may have aeration equipment permanently assigned to each pond. Paddlewheel aerators are commonly seen in culture ponds. They may be turned on only when the oxygen level falls below a specific level, or they may routinely be run during the night, or in some cases, continuously.

Mechanical aeration is not the only means of aerating ponds. New, oxygen-rich water can be added, though that can be expensive. Mechanical aeration is not inexpensive, however. It requires the equipment and the electricity or fuel required to operate that equipment.

Aquatic animals may have a wide range of tolerance for salinity (euryhaline) or may be restricted to a rather limited salinity range (stenohaline). Salinity is measured in parts per thousand (ppt). The salinity of freshwater is typically about 0.5 ppt, while full strength seawater is usually about 35 ppt. The optimum salinity for an aquatic species may be constant throughout the life cycle or it may change at various life stages. Salmon, for example, spawn and their eggs hatch in freshwater where the larvae live for a few weeks to 2 years before migrating to the ocean. Fish that spawn in freshwater but grow to adulthood in seawater are called anadromous.

Those that spawn in the ocean and migrate to freshwater to grow to adulthood, such as eels, are catadromous.

While most aquatic organisms survive and grow well at a constant optimal salinity, it may be necessary to change the salinity when rearing certain species. It is important to know when those salinity changes should be made and the extent of the changes.

Ammonia is produced by many aquatic species as a metabolic waste product. Ammonia enters the water through the gills. As previously indicated it can be present in the water in two forms: ionized (NH_4^+) or unionized (NH_3). It is the unionized form is highly toxic to aquatic animals. The ratio between the two forms is determined by factors such as water temperature and pH. As either or both temperature and pH increases, the amount of unionized relative to ionized ammonia increases. Under most conditions, total ammonia concentration should not exceed about 1 mg/l. That will ensure that the level of unionized ammonia is not excessive.

Water pH can undergo fairly dramatic changes diurnally, particularly when there is a strong algae bloom present. As plants and animals respire at night in the absence of sufficient light to promote photosynthesis, the increased level of carbon dioxide in the water causes the pH to drop unless the system is well buffered. During the daytime, carbon dioxide is removed from the water by photosynthetic organisms and pH may rise. Carbonate (CO_3^{2-}) and bicarbonate (HCO_3^-) ions make up the foundation of the buffer system. They can absorb hydrogen ions (H^+) or in the case of bicarbonate, act as a source of H^+ . If the supply of bicarbonate becomes exhausted, pH can change dramatically. Poorly buffered systems can be as acidic as pH 6 in the early morning and as basic as pH 10 in the afternoon if the buffer system is weak. In recirculating systems, pH will tend to drop over time as organic acids become concentrated in the water. Most aquatic species cannot tolerate such swings in pH. Calcium carbonate in the form of limestone or crushed oyster shell can be employed in the water system to help maintain pH by providing a source of carbonate ions.

Routine measurement of pH and alkalinity (a measure of carbonate and bicarbonate ions) every few days is common on aquaculture facilities. Hardness (a measurement of the calcium plus magnesium level in the water) may also be monitored. Some species, such as the euryhaline red drum that can be found in estuaries as well as in the coastal ocean of the southeastern Atlantic and Gulf of Mexico coasts of the United States, can survive and grow well in hard freshwater, but performance is not as good in soft freshwater.

Genetics and Reproduction

Most species being successfully produced by aquaculturists can be maintained throughout their life cycles. Closing the life cycle has been relatively simple for some species, but very difficult for others. Aquatic animals with large eggs, such as

salmon, trout, catfish, and tilapia—each of which has eggs 3 mm or larger in diameter—readily spawn in captivity. Their eggs can be incubated relatively easily and the newly hatched fish will consume prepared feeds as first food after yolk sac absorption.

Many aquatic animals of aquaculture interest have very small eggs and larvae so small as to be nearly invisible to the naked eye. The larvae tend to be very fragile and often have extremely limited swimming ability. If exposed to water currents in a hatchery raceway, they may become impinged on screens and killed. For many species, providing a complete feed that will be accepted has proved difficult so it is often necessary to culture living zooplankton as first food.

The most simple approach is to allow the animals to spawn naturally. Providing the proper conditions for spawning is not always possible, and in some cases, culturists still do not know how to duplicate what the animals experience in nature. Hormone injections are sometimes used to induce spawning. In the case of marine shrimp, removal of an eyestalk from females can trigger ovulation.

Fertilized eggs may be allowed to incubate naturally or they may be taken to a hatchery for artificial incubation. Various types of specialized hatching facilities have been developed. Not only is the survival rate generally higher in a hatchery than in a pond, but the culturist can get a good estimate of the number of offspring that have been produced since they are easy to observe in the hatchery as compared with an outdoor pond situation.

Only a few aquatic animals have been truly domesticated through many generations of selective breeding. Most are still only a step or two removed from their wild counterparts. In some cases, for example, in conjunction with hatcheries that produce fish for stocking into nature, a conscious effort is made to maintain the genetic integrity of the wild stock. For animals bred to be reared in captivity only, selection of brood stock for rapid growth, improved body configuration, disease resistance, and other desirable qualities may be employed.

The use of genetic engineering in conjunction with aquaculture species is in its infancy, but there has been some research activity in that controversial arena. Researchers have successfully incorporated selected genes from one species into another, but there do not appear to be any commercially produced transgenic aquatic animals in the market.

Nutrition and Feeding

Feed represents the largest variable cost for most types of aquaculture. The exception is mollusc culture where wild phytoplankton is fed upon in most instances (cultured algae is used in mollusc hatcheries and fertilization of ponds is widely used in China and other countries to produce natural food). Many fishes and crustaceans are provided with commercial feeds. When prepared feeds are employed, they often

represent nearly half of the variable costs associated with operating the aquaculture facility.

Most fishes and crustaceans will either accept prepared feeds upon first feeding or can be weaned from natural foods within a few weeks after first feeding. Prepared feeds may be a mixture of ground and mixed feedstuffs that are fed directly, or the ingredients may be formed into pellets. Feed pellets can be made in a variety of sizes and shapes; they can be dry, moist, or semi-moist; they may be hard or soft, and they may either sink or float depending upon the species being fed, local ingredient availability, and the type of equipment used in feed manufacture.

For some species, the nutritional requirements have been determined in some detail. Examples are salmon, trout, catfish, tilapia, carp, and to a lesser extent shrimp, striped bass, red drum, and a number of others (Wilson 1991; D'Abramo et al. 1997). Depending upon the species, requirement information may be available on dietary energy, protein and amino acids, lipids and fatty acids, carbohydrates, vitamins, and minerals.

Feeds may contain animal protein from a few percent (channel catfish) to one-third or more of the diet (some salmonids), or they may be made up entirely of plant feedstuffs with or without added vitamins and minerals. Most of the animal protein in aquaculture feeds comes from fish meal, though poultry byproduct meal, and meat and bone meal are among the other options. Locally available ingredients are typically employed when possible. Fish feeds in Asia, for example, often contain relatively high percentages of rice bran, which has limited nutritional value but is readily available. Soybean meal is a major ingredient in many fish feeds in the United States. Diets may also contain corn meal, cottonseed meal, and/or peanut meal depending on availability and cost. A wide variety of other ingredients are also used.

Aquatic animals on prepared feeds are usually fed once or twice daily during the growout period. The total amount of feed provided on a daily basis is usually not more than 3–4% of the biomass of the animals being fed. Young animals are usually fed at a higher rate (as much as 50% of body weight initially, and declining as the animals become more proficient at finding feed and their relative growth rate slows). Young animals may be fed more frequently. Some carnivorous species may be offered small amounts of feed as often as every few minutes to keep them satiated and reduce cannibalism.

Feeding may be done by hand, though increasingly, automated feeding systems are being employed. Such systems have been developed for virtually every type of water system. Some feed automatically when activated by timers, while others may be activated when animals go to the feeder. The latter, called demand feeders, are used in conjunction with some finfish rearing operations and feature a rod or plate that is suspended in the water which when bumped by the fish causes a few feed pellets to be released.

Diseases

Aquatic animals are susceptible to a wide variety of diseases. Organisms responsible for disease in aquatic species include fungi, bacteria, nematodes, cestodes, trematodes, as well as parasitic protozoans, copepods, and isopods. Some can cause death while others may stress the affected animal to the point that it becomes more susceptible to additional diseases. Not all aquatic animal diseases are caused by other organisms. Some, such as gas bubble disease, are caused by water quality problems (in this case gas supersaturation). Nutritional deficiency diseases also exist. For example, vitamin C deprivation can lead to decalcification of bone, which can lead to spontaneous fracturing of the vertebral column.

While a variety of disease treatments have been developed and are in widespread use around the world, only a very small number of drugs have been approved for use on aquatic organisms in the United States. The US Food and Drug Administration must approve each drug utilized on foodfish. Because the industry is relatively small, the market for drugs often does not justify the financial investment required by the pharmaceutical companies to obtain clearance.

Controversies in Aquaculture

In recent years, aquaculture has come under increasing criticism from environmentalists. Particularly heavy criticism has been leveled against aquaculture being conducted in public waters. The conduct of aquaculture on private land has received less opposition, except in cases where effluents from such facilities enter public receiving waters.

Salmon net-pen culture in Washington and Maine, and marine shrimp culture in Texas are two examples of aquaculture activities that have been targeted by anti-aquaculture groups. The net-pen industry in Puget Sound, Washington was initially objected to because upland landowners felt the net-pens represented visual pollution. That objection failed to gain traction in the courts, but many other criticisms followed (Stickney 1990). Included were claims of:

- Waste feed and feces causing dead zones under net-pens
- Nutrients released from decaying feed and feces leading to eutrophication
- Antibiotics in feed entering the food chain leading to the development of resistant bacterial strains
- Net-pens interfering with recreational and commercial fishing
- Net-pens interfering with navigation
- Fish in net-pens transmitting diseases to wild fish
- Escapees from net-pens interbreeding with wild fish, or in the case of nonnative Atlantic salmon successfully colonizing and displacing wild salmon
- Net-pen aquaculture operations being noisy and producing noxious odors.

Many of the claims have at least some basis in reality or hold the potential of becoming problems, though each can be effectively dealt with if the net-pen facility is properly constructed and placed in the proper location (Parametrix 1990).

If located in areas where tidal exchange and currents are insufficient to carry away waste products, an anoxic zone can occur. In Japan, uncontrolled development of net-pen facilities in several bays led to so-called self-pollution that resulted in heavy mortalities. Strict regulations on the number of net-pens per unit area are now in force in Japan and the problem has been resolved.

Salmon farmers, as other aquaculturists, are dedicated to maintaining the highest possible water quality and other environmental conditions for the animals under their care. Degradation of the culture environment causes stress and can lead to disease. If disease does occur, an approved antibiotic may be employed, but only for 10 days at recommended rates, so large amounts of such pharmaceuticals are not used.

Atlantic salmon have been known to escape but they cannot interbreed with Pacific salmon (the situation is different in Maine where Atlantic salmon are native). Many attempts were made to establish Atlantic salmon in the Pacific Northwest by the US Fish and Fisheries Commission over a century ago. All of those attempts failed. Atlantic salmon that have escaped into the waters of the Pacific Northwest in North America do not seem to compete well with native Pacific salmon.

Proper siting will help avoid problems with nutrient concentration and interference with other users. Excessive noise is not a valid criticism of net-pen operations and there would only be noxious odors in the event of a fish kill that was not immediately cleaned up.

The Texas shrimp farming industry has come under criticism for releasing sediment and nutrient-laden water into adjacent public waters. In one part of the state the suspended solids load from shrimp pond effluents has been blamed for heavy siltation of navigable waters, while nutrients in the effluent are blamed for harmful algae blooms. There has also been concern expressed that diseases in cultured shrimp could spread to wild populations. The industry is dealing with the situation by constructing wetlands through which water exiting the production ponds passes. Solids can settle and nutrients can be taken up by the wetland vegetation. The water may then be recirculated back to the ponds rather than released to the environment. Significant improvements in water quality have been achieved and production costs reduced when constructed wetlands are employed.

The other major criticism of the shrimp industry in Texas and other states involves the use of exotics. Most of the shrimp raised on farms in the United States are exotics from Asia or Latin America. Research on native species was all but halted in the 1970s, but many feel that attempts should be

initiated to solve some of the problems so the industry can move away from exotic or nonindigenous species.

The development of extensive pond culture for shrimp has led to destruction of mangrove areas in many tropical countries. Some nations have recognized the importance of mangroves as habitat for marine organisms and wildlife and as buffers that can dampen the effects of storms on upland regions. As a result, mangrove habitat destruction has been outlawed or significantly reduced due to governmental regulation. In addition, mangroves grow in acid soils, which are not as conducive to supporting aquaculture operations as other soil types.

Many countries have brought in nonindigenous species and introduced them into culture. Besides Atlantic salmon and marine shrimp as previously discussed, examples include tilapia in North, Central, and South America as well as throughout much of tropical Asia; Latin American species of marine shrimp in the United States; various species of carp in the United States and Latin America; and walking catfish (*Clarias* spp.) in Europe. For species which have only recently been introduced into aquaculture as exotics, there is concern about escapement and establishment of wild populations that could compete with and perhaps displace native species. The rapid spread of common carp throughout North America beginning in the nineteenth century and the dispersal and establishment of tilapia in parts of Asia (where tilapia are now considered native though they have been present in Asia only since the 1930s) and Latin America are good examples.

While transgenic fish are not apparently being commercially produced for sale, researchers have been able to incorporate growth hormone producing genes from one species into another and obtained an increased growth rate. Transgenic fish with so-called antifreeze genes that might, for example, allow tilapia to survive at much lower temperatures than normal have been discussed. This type of activity is controversial and there are many stories in the media that have convinced at least a portion of the public that transgenic fish are threats to the environment as well as to the health of consumers. A great deal of debate surrounding this activity will occur before transgenic fishes become commonplace in the market.

Cross-References

- [Economics of Beaches](#)
- [Environmental Quality](#)
- [Estuaries](#)
- [Human Impact on Coasts](#)
- [Mangroves, Ecology](#)
- [Mangroves, Geomorphology](#)
- [Mangroves, Remote Sensing](#)
- [Water Quality](#)

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Archaeology

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Definition

Humans have inhabited coastal areas and made use of their resources for hundreds of thousands of years. The richness of the coastal zone, the overlapping of maritime, littoral and inland resources, and the ease of transportation along it have led people to cluster along shores, including the people of today who are destroying much of the record of past inhabitants. However, coastlines are also the most dynamic environments on earth. Tides and currents effect daily and seasonal changes in the shore margin; storms dramatically resculpt the shore; isostatic, eustatic, and tectonic changes (cf. *Isostasy*, *Eustasy*, *Tectonics*, and *Neotectonics*) can result in former coastlines being located far inland or far out to sea.

In trying to understand the human use of coastlines, it is necessary first to discover where the coast was at the time the site one is interested in was occupied. Due to the ever-changing relationship between the land and the sea, sites of past coastal people may be well inland, well submerged under the sea or still coastal. To discover past coastal sites, it is essential that archaeologists work in cooperation with geologists and other coastal scientists, and take into account both local and global factors which influence relative sea level (cf. *Archaeological Site Location*). Once sites are located and are determined to be cultural rather than natural features, archaeologists turn to more culturally interesting questions about human adaptation to the coastline and how it changed with time. For example, were these people simply living by the sea, like summer beach residents today or in Pompeii, or were they using marine resources? If they were using marine resources, was their primary focus on industrial or alimentary needs? If the latter, were they littoral hunter-gatherers, who made use of coastal and nearshore resources, or were they maritime hunter-gatherers, who had boats from which to gather creatures of the deep sea?

To understand human adaptation to coastal environments, archaeologists in all parts of the world have looked particularly closely at diet, at the season of occupation of sites, and at relationships between coastal and inland peoples. Changes in adaptation can be forced by changes in the environment, by cultural change and development or, most likely, by a complex interaction between culture and environment. In the short term, archaeologists consider the human response to disaster: for example, if their settlement is destroyed by a *tsunami* (*q. v.*), do people move away or do they reestablish themselves in the same place? One can examine changes through time in coastal occupation, both those forced in some way by environmental changes, for example, rising or falling sea levels, changes in ocean temperatures, which affect littoral or maritime resources, and those caused by social, political, or demographic changes in the human populations of the coast. In the following pages, studies from various parts of the world will be used as exemplars of current examinations of these issues.

The vast majority of coastal archaeological sites are shell middens, piles of shells left behind by people eating mollusks and/or using their shells for industrial or ceremonial purposes. Other sites consist of modifications of the shoreline to enhance the productivity of the sea or humans' ability to harvest its resources.

Historical Development

The archaeology of coastlines has a long history, with some of the earliest studies in archaeology focusing on the Danish *kjokken-möddings*, or shell middens, of the Mesolithic period. These were recognized as cultural by 1837, and shortly

thereafter, the Royal Danish Academy of Sciences established a commission to study them, directed by Jens J.A. Worsaae, who was the world's first professional prehistoric archaeologist. In the 1850s, Worsaae and his colleagues published six volumes on their studies, which established that the Danish middens were cultural and had developed through time, that the only domesticated animal the people possessed was the dog, and that these middens were occupied in all seasons of the year but summer. They also identified the plants of the paleoenvironment and studied the distribution of cultural features within the mounds. Finally, they initiated experimental archaeological studies: they gave bones to dogs to chew on to determine why there was a preponderance of bird long-bone mid-shafts in the middens. This work was highly influential, stimulating scientists to study local shell middens on both coasts of the United States, and in Japan, Brazil, and Southeast Asia. By 1911, shell middens had been examined from the Aleutian Islands to South Africa (Trigger 1989). The study of coastal sites has not slowed down, since: a 1991 bibliography of coastal and maritime archaeology, admittedly incomplete, contains 2800 entries (Kerber 1991).

In 2011, Andrea Balbo and colleagues edited a special edition of *Quaternary International* with articles reviewing shell midden research around the world. In their introduction to the volume (Balbo et al. 2011), they reviewed the history of shell midden research, its current state, and possibilities for the future. The articles and their geographical foci are listed below:

Quaternary International vol. 239, issues 1-2: 2, 2011		
Authors	Geographic area	Chronology
Balbo et al.		
Rabett et al.	Vietnam	
Alvarez et al.	State of the Art	
Wagner et al.	Brazil	8 ka–0
Colonese et al.	Mediterranean	300–6 ka
Habu et al.	Japanese archipelago	14–1 ka
Ramos et al.	Gibraltar Strait	250–5 ka
Briz et al.	Tierra del Fuego	1 ka–0
Szabó and Amesbury	Tropical island Asia-Pacific	2 Ma–0
Erlandson et al.	NW Pacific	16 ka–0
Saunders and Russo	Florida	13–3 ka
Orquera et al.	SW Patagonia and Tierra del Fuego	12 ka–0
Gutiérrez-Zugasti et al.	Atlantic Europe	35 ka–0
Rick	Channel Islands California	4.5 ka–0
Demarchi et al.	Red Sea and Scotland	9 ka–0

Rabett et al. provide an overview of shell midden research but then turn to freshwater middens. Alvarez et al. review shell midden research overall without any geographic focus, while six of the contributions focus on area reviews (Wagner et al., Colonese et al., Habu et al., Ramos et al., Briz et al., Gutiérrez-

Zugasti et al.). The remaining articles, which primarily present new research (Szabó and Amesbury, Erlandson et al., Saunders and Russo, Orquera et al., Rick) or new methods of analysis (Demarchi et al.), will be discussed further below.

Following up on these studies, in an ethnoarchaeological and archaeological study of shell middens in Senegal, Hardy and colleagues (2016) extensively reviewed reports on mid-Holocene shell middens and mounds around the world. Their references, added to the ones above, will be invaluable for anyone interested in shell midden research.

Finding Coastal Sites

Finding coastal archaeological sites can be very difficult due to isostatic, eustatic, and tectonic forces that change the location of the sea relative to the land. The west coast of North America is an area in which all of these forces have been active to varying degrees during the period of human occupation. Fedje and Christensen show that, in the Queen Charlotte Islands, coastlines dating from 13,000 to 9500 BP [uncorrected radiocarbon dates] are deeply drowned while those dating from 9200 to 3000 BP are stranded in the rainforest up to 15 m above modern levels. “Coastlines have been approximately coincident with the current position for only the last two to three millennia and for a century or two centered around 9400 BP” (1999, p. 635). Modeling these *paleocoastlines* (*q.v.*) allowed them to discover a number of coastal archaeological sites in both drowned and uplifted zones. Since then, 20 additional raised beach sites have been found, and the search for drowned shorelines with sites has continued (Fedje et al. 2005). From southern Washington to northern California, on the other hand, late prehistoric coastal sites have been drowned due to earthquake-induced subsidence (cf. *Subsiding Coasts*) along the Cascadia subduction zone, seriously biasing our understanding of Washington's prehistory (Cole et al. 1996). While many sites thus drowned may be lost, the authors have found a number of examples buried under tidal mud along coastal *estuaries* (*q.v.*). In southern California, Waters et al. (1999) have identified a complex series of terraces along streams that are caused both by global patterns of sea-level fall and rise and by local climatic factors, possibly in response to an *El Niño-Southern Oscillation* (*q.v.*) event. These events have caused pre-4000-year-old sites to be deeply buried and consequently difficult to find. The later sites, which are closer to the surface, have an advantage over many southern California sites in being below the bioturbation layer. Waters and his colleagues have found large late marine-oriented sites in these channels, which are changing our understanding of late prehistoric cultural development in this area.

Determining Site Function

Both Hardy et al. (2016) and Marquardt (2010) caution against the facile interpretation of the functions of shell mounds. Through their ethnographic research, Hardy et al. point out that “In the Saloum Delta, perceptions of the middens are not static and they change and evolve as the middens develop; they can be dumps, villages, cemeteries or burial sites, religious places, foraging locations and possibly even monuments; no interpretation excludes any of the others. We suggest that this fluidity in the meaning and use of middens, that we have detected challenges some of the fixed social interpretations that can be ascribed to middens and that rather than assigning specific meaning that suggests deliberate, pre-planned action sometimes over thousands of years, middens are more likely to have had many uses at different times during the course of their construction and active lives” (2016, p. 30). Similarly, Marquardt avers that only by using carefully sedimentological analysis can one determine the function of Southeastern US shell middens and mounds.

Natural/Cultural

In many parts of the world, shell midden sites have been quarried out for fertilizer or to use as fill. Thus, many midden deposits may be lost or exist in places far removed from their original deposition. Even in remote places like the Shumagin Islands south of the Alaska Peninsula, people have moved midden from one island to another to serve as fertilizer for gardens. Thus, merely finding onshore shell deposits, even when it can be determined that they are of human creation, does not mean that one has found an archaeological site *in situ*.

In Sri Lanka, the lowering of sea level has led to shellfish dying and their shells being washed and blown into windrows on beaches, lagoon and lake bottoms, sand dunes, and headlands. These are often difficult to tell from cultural deposits although artifacts found among the shells suggest that people were responsible for some of the deposition (Katupotha 1995). In southwestern Louisiana, Henderson et al. (1998) also wished to determine whether shell-rich deposits on coastal beach ridges were natural or cultural. In this case, the archaeological deposits were found lying on or actually constructed out of reworked *chenier* (*q.v.*) deposits. The authors discovered a suite of taphonomic, sedimentologic, and stratigraphic methods, which served to distinguish the cultural deposits from the natural *chenier* deposits.

Butler and her colleagues have been working during the past decade to discover how natural deposits of salmon differ from those left by human hunter-gatherers. This is a particularly important problem in the Pacific Northwest where the harvesting and storage of vast quantities of anadromous

salmon are thought to underlie the development of the cultural complexity for which the area is famous. Butler has looked at a particular site, the Dalles Roadcut on the Columbia River, and concluded that the deposition is probably cultural; has considered the varying natural decay of different salmon bones; and has investigated whether fish bones that have passed through mammalian digestive tracts bear recognizable marks (cf. Butler and Schroeder 1998). All of these studies aid archaeologists in determining whether salmon deposits are natural or cultural.

An ethnoarchaeological study of fish middens on the coast of Senegal showed that caught and filleted fish mimicked natural fish deposits as the heads were not removed in filleting and, therefore, the entire skeleton remained (Van Neer and Morales Muñiz 1992). On Lake Turkana in Kenya, on the other hand, investigators could distinguish base camps, fish processing camps, and fish waste discard sites from each other and from natural death assemblages by the nature and number of bones remaining (Stewart and Gifford-Gonzalez 1994). Both of these ethnoarchaeological studies can aid archaeologists in interpreting prehistoric shell deposits.

Related to the problem of distinguishing natural from cultural deposits is that of determining why species representation and shell size of shellfish found in middens might change through time. In looking at middens on the West Coast of Southern Africa, Jerardino (1997) determined that changes in species were due to humans exploiting different tidal zones, while size changes in individual species were due to a combination of water conditions and human exploitation.

Industrial and Ceremonial

The study of industrial and ceremonial uses of coastal resources, particularly shells, has increased tremendously over the past decade. In the 1990s, working in the Sinai, Bar-Yosef Mayer (1997) demonstrated that shells entered sites as raw materials for beads rather than as food remains, while in the Arctic, Savelle (1997) notes that whale bones are often more indicative of housing needs than diet. Most investigators look at shellfish gathering as an enterprise that requires few or no tools aside from a container in which to carry the collected goods. However, Johnson and Bonsall (1999) note that in both the Shumagin Islands in Alaska and in Scotland, tools are found in shell midden contexts which appear to have been used in the collection and processing of shellfish. Wedge-ended rods of stone or sea mammal bone are used for prying limpets from rocks and, possibly, for digging the animal out of its shell, while small splinters of bird bone are used to pry periwinkles and other small snails loose from their shells, developing worn points in the process. Jerardino (2016, p. 215) expands the recorded use of levers to South

America, Africa, Australia, New Zealand, and the Pacific coast of North America.

Shell technologies have been studied from Upper Paleolithic Spain, where *Patella* shells were used both to process pigments and to tan hides (Cuenca-Solana et al. 2013), to Baja California, Mexico, where single-piece shell fishhooks dated to the terminal Pleistocene/early Holocene transition were found in levels which “contained a diverse assemblage of fish remains, including deepwater species, indicative of boat use” (Des Lauriers et al. 2017). In the Pacific, Pawlik et al. (2015), found a *Tridacna* shell adze in a site in the Philippines which showed that shell technology emerged in the early middle Holocene and implied an early human interaction between the Philippines and Melanesia. On the other end of the timescale, Kononenko et al. (2010) showed that turtle bone cleavers from Wuvulu Island, Papua New Guinea, which were traded to Germans in the nineteenth century were well-used and ready for discard rather than being manufactured for the tourist trade.

Marquardt and Kozuch (2010) published a fascinating article on the industrial and ceremonial uses of the lightning whelk (*Busycon sinistrum*) in the North American southeast and midsouth. In the same area, Wallis and Blessing (2015) looked at a large pit feature in Florida to understand feasting among small-scale hunter-gatherers. The remains included saltwater taxa which must have come from the Gulf or Atlantic coasts more than 100 km away. Also in Florida, Pluckham et al. (2016) identified stepped pyramids of shell from a terminal Late Woodland site. In Georgia, in addition to examining individual shells to determine the seasonality of use (see below), Thompson and Andrus (2011) also identified both gradual deposition of shells as food waste and short-term large-scale deposition of shells which might indicate that the rings became monuments. To the west, Gamble (2017) urged archaeologists to consider the ritual significance of shell mounds in southern California; while Whalen (2013) reinterpreted the huge caches of marine shell at Casas Grandes (or Paquime) in northwest Chihuahua, Mexico as “animate objects,” which formed “a vast repository of supernatural power that was a central part of the community’s ritual system.”

Four recent studies looked at shell beads: Tata et al. (2014) used experimental archaeology to analyze beads from Vale Boi, Portugal, and concluded that while species changed, technology remained the same over a long period of time; Christiani et al. (2014) studied the site of Vela Apila (Corcula Island, Croatia) to understand Late Upper Paleolithic and Mesolithic personal adornments dating from 19,500–8150 cal BP. Diversity of shell species decreased to one type in the Mesolithic indicating that the site became an important procurement and processing center for *Columbella rustica*. On the Pacific Coast of North America, Coupland et al. (2016) found unequal distribution of shell and stone disc beads from 4000 cal

BP leading them to conclude that material wealth-based inequality began earlier than previously thought on the Northwest Coast, while Smith et al. (2016) discovered that marine shell beads from the northern California, Oregon, and Washington coasts were deposited in a rockshelter in southcentral Oregon during the early Holocene.

Coastal Adaptation

Once coastal sites have been located and identified, archaeologists attempt to discover what people were doing at the sites. This involves figuring out what foods they were eating, by examining both the food remains left at sites and the bones of the humans, which reflect the nature of their diets. In looking at the food remains, it is necessary to figure out what the dietary implications are of, for example, ten bivalve shells versus two salmon bones versus one sea lion femur. Archaeologists are also interested in discovering whether people were using the site year-round or only seasonally, whether they were littoral gatherers or maritime hunters and fishermen, and what their relationship with other groups might be. Finally, archaeologists look both at individual sites occupied for a long period of time and at the pattern of sites within an area to discover how human adaptation to the coastline changed through time. Another line of discovery is provided by ethnoarchaeological and experimental studies (Briz Godino et al. 2011)

Diet

The most direct means of discovering what people ate is to examine their bones. Analyses of trace elements and stable isotopes can distinguish between diets rich in marine foods and those rich in terrestrial foods. Low Ba/Sr and Ba/Ca ratios and high $\delta^{13}\text{C}$ values in bones indicate a diet rich in marine foods, while the reverse is true for terrestrial diets. High $\delta^{13}\text{C}$ values also occur when plants such as maize, which use a C4 metabolic pathway, are eaten, but maize and marine foods in the diet can be distinguished by examining $\delta^{15}\text{N}$ and $\delta^{34}\text{S}$ values.

Gilbert et al. (1994) examined modern plants and animals of the western Cape of South Africa in order to determine marine and terrestrial signatures for Ba/Sr and Ba/Ca ratios, finding distinct differences in both edible tissue and bone. Reanalyzing bones, which had already been tested for $\delta^{13}\text{C}$ values, led to the conclusion that both methods worked, but that stable isotopes revealed more subtle differences.

Aufderheide and his colleagues (1994) had the advantage of working with mummified human remains from northern Chile. Here, they were looking at recent immigrants to the coast from the highlands and discovered that the community appeared to have split into two different groups, one right on the coast subsisting almost entirely on marine foods, and the

other in the mid-valley eating both marine foods, probably supplied by the coastal community, and valley bottom foods. These results were supported by the presence of external auditory canal exostoses, formed due to prolonged cold water diving, in the coastal population.

In looking at the south Florida Calusa, Hutchinson et al. (2016) “combine data on the zooarchaeological and archaeobotanical remains from the archaeological sites with those from stable carbon and nitrogen isotopic ratios of archaeological human bone, and modern and archaeological plants and animals. These multiple lines of evidence confirm that marine-based protein and terrestrial C3 plants provided a large and reliable portion of the diet in southwestern Florida as early as 4000 years ago and up to European contact.”

Human remains are not always available in the sites or levels where archaeologists would like to find them, and the prehistoric people’s descendants are often not pleased to have their ancestors subjected to scientific analysis. Cannon et al. (1999) have demonstrated at the site of Namu on the British Columbia coast that dog bones can serve as surrogates for human bones in analyzing diets. It is probable that in most prehistoric coastal settlements the diet of dogs was very similar to that of their masters, thus providing archaeologists a much wider sample of bones to analyze for dietary information. Thornton et al. (2011) looked at the Angoon-Killsnoo area where seven of nine sites had herring bones, spanning 1800 years, pointed out that herring remains are found in Northwest Coast sites by 8–9.9 kya, and noted that this continued productively was possible due to managing people and marinescapes.

Another approach to dietary reconstruction is to estimate the amount of food represented by the remains present in a site. James Savelle has been particularly active in deriving meat utility indices for marine mammals (cf. Diab 1998). This also relates to issues of variable preservation, discussed above, and to recovery methods. Fish bones are often under-represented in archaeological assemblages because of a lack of screening of excavated materials or of using screens with too large a mesh. Ross and Duffy (2000) have demonstrated that a 1 mm mesh screen results in considerably greater recovery of fish bones from an Australian site and have also developed deflocculation and flotation techniques, which make sorting the screened remains practical in the real world: sorting remains from a 100 ml dry screened sample took 20 h, whereas the deflocculated and floated sample was sorted in only 2.5 h. Moreover, the chemical added to water for flotation was sugar and an excellent deflocculant was baking soda, making the procedure inexpensive and safe as well. The result of the 1 mm screening was to verify the reports of southeast Queensland Aborigines that both fishing and shellfishing had been important to them since time immemorial in contrast to prior archaeological investigations in which the paucity of fish remains recovered had led to a hypothesis of a late introduction of fishing to this coast.

Analysis of midden materials also gives rise to problems concerning how to quantify the materials recovered. Faunal analysts now tend to provide both number of identified specimens (NISP) and a minimum number of individuals (MNI) data for mammal and fish remains, but the quantification of shellfish remains is a thornier issue. Shells appear in sites in the thousands and are often broken into small fragments. Because shells vary in size and in friability, NISP is not a useful figure. MNI is often used but ignores a great deal of the data since only the hinge areas of bivalves can be counted. Thus archaeologists have attempted to use the weight of shellfish remains as a proxy for the importance of various species in past diets. This is an area of active discussion (cf. Claassen 2000).

Season of Occupation

Coastal settlements can be year-round residences or can represent part of a seasonal round, being occupied for one or two periods during the year while the people live elsewhere, exploiting other resources, during the rest of the year. Determining seasonality from the middens themselves requires some means of discovering when animals were killed, which has proven possible for mammals, fish, and shellfish. Marine mammals are born at particular times during the year, so the age at death of immature animals can indicate when a site was occupied: if all of the young animals found in a site died at the same age, it suggests that the site was occupied for only that season. The analysis of seal bones in South Africa indicated that early in the Holocene people exploited seals periodically throughout the year, while in the late Holocene they restricted this activity to the spring (August–November; Woodborne et al. 1995). A similar pattern of change from year-round to seasonal occupation is seen in New Zealand through the analysis of seasonal and annual growth rings on Cod otoliths (earbones; Higham and Horn 2000).

A major approach developed over the past 15 years or so is sclerochronology, the study of physical and chemical accretionary structures of organisms such as shellfish and the seasonal and annual patterns in which they formed, coupled with the sequential analysis of oxygen isotopes (δ18O) (see Andrus 2011 for an excellent review of the processes of analysis and the benefits and pitfalls for shell midden analysis). Recent studies have looked at the Northwest Coast of North America (Hallmann et al. 2013; Burchell 2013; Burchell et al. 2012), California (Culleton et al. 2009; Jew et al. 2013, 2014; Eerkens et al. 2013), Florida (Andrus and Thompson 2012; Wang et al. 2013), Mesolithic Europe (Dupont 2016), Libya (Prendergast et al. 2016), and Tierra del Fuego (Colonese et al. 2011).

Littoral or Maritime

Prehistoric coastal residents who exploited the sea can be defined as littoral, exploiting fish and shellfish resources

available from the shore, or maritime, using seaworthy vessels to capture animals which do not come close to shore. Most maritime people also exploited the shore. In his 2016 article exploring shellfish use among Middle and Late Stone Age hunter-gatherers in South Africa, Jerardino provides an extensive review of the worldwide literature on littoral adaptations.

Littoral adaptations almost certainly preceded the evolution of *Homo sapiens* (Jerardino 2016; Klein and Bird 2016) in Africa and probably Europe. However, determining exactly when people began exploiting the resources of the sea is difficult due to the rising postglacial seas which have buried many coasts. At the Haua Fteah cave in Libya, shellfishing dates to the terminal Paleolithic Early Epipaleolithic, 17.2–12.5 kya (Prendergast et al. 2016). Along the Pacific Coast of North America, models of coastal immigration from Asia to the Americas suggest that the earliest migrants exploited shellfish and might have had a full maritime adaptation; however, the earliest dates, from San Miguel Island in California, are around 10 kya (Braje et al. 2014). In Florida, the date of the earliest shellfish collectors is ca. 7000 ya (Saunders and Russo 2011).

In trying to understand Mesolithic coastal sites in western Scotland, Bonsall (1996) makes use of ethnographic data which indicate that the Australian Aborigines had three recognized types of shell middens: processing camps, where shellfish were removed from their shells and prepared for transport elsewhere; dinnertime camps, where people involved in other activities stopped, collected shellfish, and enjoyed a meal; and home bases. In Scotland, given the generally inclement weather, remnants of the first two types of camp would probably be found in shelters or other protected locales. Bonsall hypothesizes that the Obanian sites found around the western coast of Scotland are probably the remnants of processing and dinnertime camps occupied by gatherers whose base camps were located inland. The latter had a much richer artifact inventory, including the quintessential Mesolithic microliths, which are not found in the coastal sites.

In the much more complex society of Inka Peru, the fishermen of the Chíncha Valley formed a separate community from the valley farmers and overlords whose protein they supplied. Specialized fishing communities seem to have existed in coastal Peru before the Inka took the area over and the Inka probably incorporated these communities into their empire with little change except for skimming the fish off for their own use rather than allowing the valley paramount lord to do so. The fishing here was primarily, if not entirely, littoral netting of small fish such as anchovies and anchovetas, which were then dried for exchange with other members of the valley and imperial communities (Sandweiss 1992).

The earliest known maritime adaptations in South America, and worldwide, come from sites in north-central Chile, where,

between 7400 and 5900 cal BP, “not only were shallow waters and marine mammals exploited, but . . . a dedicated fishery for large pelagic fish existed; with indications suggesting that large swordfish, weighing up to 300 kg, were being caught and brought back complete to the settlement” as well as striped marlin, tuna, and amberjack (Bearez et al. 2016). Around 5000 BP, people bearing Arctic Small Tool Tradition implements entered Alaska and colonized the coast, developing a dual terrestrial/maritime economy (Tremayne and Winterhalder 2017). Around 3000 years ago, the residents of the Hoko River site in Washington state possessed a full-scale maritime adaptation. Due to half of the site being waterlogged, there is excellent preservation of organic artifacts, including numerous fishhooks, landing skids for canoes, pack baskets for transporting the fish, primarily halibut, and mats to protect the canoes and keep them from drying out between fishing trips. Around 3000 years ago, the site was seasonally occupied for a number of years, “mainly in the spring/summer season, when [the people] focused on offshore hook-and-line fishing for bottom fish, and particularly flatfishes” (Croes 1995, p. 229).

Coastal/Inland Relationships

Many studies look at the relationships between coastal people and their inland neighbors, or between coastal sites and the inland sites which formed other parts of their seasonal round. Kennett and Voorhies used the oxygen isotope information from marsh clams, discussed above, to determine changes in the relationship between people and their environments. During the early late Archaic period, people collected clams year round, though they focused on the dry season.

“Through the late Archaic period, a general trend occurred toward wet season use of these locations. This culminated at the end of the late Archaic period with the exclusive use of the littoral zone during wet season months. These data indicate a fundamental shift in the way these estuarine locations were being used. We argue that people living in this region altered their overall subsistence strategy during the late Archaic period due to scheduling conflicts that occurred with the adoption of maize agriculture” (Kennett and Voorhies 1996, p. 689).

A similar process appears to have taken place in the Southeastern United States. Here ranked societies developed which took advantage of both coastal and interior resources. The origins of these societies are not well known, but evidence from Chesapeake Bay suggests that a pattern of springtime collection of oysters for local consumption was replaced, as maize arrived, with intensive spring harvesting and preserving of oysters. Anadromous fish and shellfish were major sources of protein along the coastal plain, where interior mammals such as white-tailed deer were rare. Farther south, in Florida, the development of the Calusa cultural pattern is very imperfectly known, but maritime and littoral resources were clearly of major importance (Waselkov 1997).

On the California Coast trade between the island and mainland sites intensified during the drought-stressed Transitional period (A.D. 1150–1300), when “islanders shipped millions of beads to mainland Chumash communities in exchange for many kinds of tools, services, and foods” (Arnold and Martin 2014).

On Kodiak Island, Alaska during the Kachemak period, the permanent settlements are on the shore and sites in the interior tend to be summer salmon fishing camps. The interior Outlet Site, while it was primarily a fishing camp, also showed evidence of fall/winter bird hunting and boasted substantial houses. Steffian et al. (2016) hypothesize that, as the population on Kodiak grew, the period of use of interior resources was extended and the coastal people, needing to lay stronger claim to their summer settlements, built substantial houses to serve as signs of ownership when they were absent.

Finally, returning to the Saloum Delta in Senegal (Hardy et al. 2016), it appears that the meat from many of the large shell middens was traded, processed, and dried, far into the interior.

Change in Coastal Adaptation

The majority of archaeological studies of coastal societies focus on change through time as people adjust to the constantly changing physical and cultural environment. Studies of the changing relationships between people and coasts have taken place in all parts of the world and only a few can be highlighted to give the flavor of current research on this topic.

The classic, and in many ways still unmatched study of human reaction to changing coastlines, is Shackleton’s study of Franchthi Cave in Greece (cf. 1988). During the 20,000 years the cave was occupied, sea level rose from –115 m to its present level. At varying times in the past, the shore in the vicinity of the cave supported different shellfish communities. By comparing the proposed communities to the shellfish remains in the cave, Shackleton was able to hypothesize that the people selectively gathered these resources rather than merely collecting the closest available species.

A short distance to the north of Franchthi Cave, near Volos, Greece, Zangger (1991) has studied changes in settlement along a prograding rather than submerging coastline. At ca. 6000 BP, Holocene sea levels reached their maximum height, and the shore in the vicinity of Volos was 3 km farther inland than it is today. Combined archaeological and geological investigation has shown that during the following 6000 years coastal cities followed the coastline as it prograded to the east.

In many urban coastal areas, engineers have captured land from the sea, using dikes and landfills to increase human living space. Nowhere did this process begin earlier or proceed further than in Holland. By 1984, Dutch archaeologists

were already involved in detailed geoarchaeological studies looking at the changing relationships between people and the land. Brandt et al. (1984) showed how rising sea level changed the configuration of the land west of Amsterdam and, thus, its use by early settlers. By the early Iron Age, at least, human livestock was also changing the land, simplifying its ecology and creating a human-dominated landscape, a process that continued as people began to enclose the land with dikes.

On the Texas Gulf Coast, Archaic Period habitation closely followed relative sea-level rise. When sea-level rise was rapid, estuarine resources were impoverished, and humans abandoned the region. When sea-level rise slowed, “the emergence of highly photosynthetic bay/lagoon shallows support[ed] nutrient-rich shoreline vegetation [which] provided the basis for a rich, exploitable aquatic biomass” (Ricklis and Blum 1997, p. 287), which the people proceeded to exploit. In the past 10,000 years, this cycle repeated itself three times, with settlements between 7500 and 6800 BP, between 5900 and 4200 BP, and between 3000 BP and the late Prehistoric Period, and abandonments between 6800–5900 BP and 4200–5900 BP. During the first two occupations, the people focused on shellfish; during the course of the last ongoing sedimentation under conditions of stable sea level led to formation of the continuous modern barrier island chain by ca. 2500–2000 BP. As the disconnected barriers coalesced, backbarrier lagoons became extensive vegetated shallows, providing protective and nutrient-rich habitats for important fish species, with a resultant increase in estuarine fish biomass. A marked increase in fish remains in archaeological contexts, beginning at the time, indicates the emergence of intensive fishing as a major subsistence focus (Ricklis and Blum 1997, p. 306).

In Georgia, Turck and Thompson (2016) used a framework derived from Resilience theory to look at changes in adaptation as sea level dropped ca 5000 cal. BP, noting that while there were differences in adaptation between deltaic and nondeltaic environments, communities in both areas were resilient to change, partially through intervillage exchange between regions.

Short- and Long-Term Responses to Environmental Disaster

Many coastal changes are gradual and may take generations to make themselves felt. As Brandt et al. state, people “perceive slow, continuous changes as unchanged, as similar until so much change has occurred that it is clearly seen. Then, they . . . change . . . much more drastically than the circumstance requires” (1984, p. 14). Other changes, on the other hand, such as hurricanes, tornadoes, earthquakes, and tsunamis, are sudden, dramatic, and disastrous. Human response to sudden disaster varies according to the circumstances and can be looked at on various scales (Johnson 2002).

Two recent studies of North Pacific coastal peoples come to quite different conclusions concerning human response to catastrophic change. Jordan and Maschner (2000) have studied the western end of the Alaska Peninsula, where isostasy, eustasy, tectonics, and volcanoes have served as a dynamic stage for human settlement. The cultural phases they have identified mark changes in village organization and subsistence orientation in the area and reflect some combination of environmental, social, economic, and political change. The dominant factor in the environmental history of the area has been isostatic recovery, but Jordan and Maschner argue that two large or great earthquakes have had major effects on regional prehistory. The first, in ca. 2200 BP, led to coseismic subsidence which increased site erosion and submergence. No sites are dated in the area between 2400 and 2100 cal year BP suggesting that sites occupied shortly before the earthquake were destroyed and the area was abandoned for a century following the earthquake. In the following cultural period, there is a major change in village organization and subsistence. The most radical change in village organization took place about 850–950 cal year BP in concert with changes in almost all the eastern Aleutian arc. There does not seem to be any local environmental instability correlated with these changes. However, there was a great earthquake in the Gulf of Alaska between 700 and 900 BP. Environmental recovery probably took a relatively short period, but cultural disruption may have lasted considerably longer. Jordan and Maschner argue that the displacement of populations from the region of the epicenter may have triggered the cultural changes in the wider area.

On the south side of the Alaska Peninsula, Johnson does not see as strong a reaction to catastrophe. The initial occupation of the islands appears to have been rapid, following a major uplift episode, which substantially increased terrace living space in the islands. The immigrants may have been fleeing from areas where the effect of the earthquakes was subsidence rather than uplift. Johnson concludes that “for maritime hunter-gatherers like the Shumagin Islanders, earthquakes are terrifying occurrences which may slow cultural development and force populations to move their settlements, but they do not seem to have a major effect on cultural development. Shumagin history, like ours, is more affected by human events than by natural events, however catastrophic” (Johnson 2002).

Archaeology in Service to Geology

While archaeologists most often use the results of other scientists to help interpret their sites, occasionally archaeological data are used to help interpret geological data. The majority of these studies concern sea-level change. Since people do not live underwater, sites found beneath the sea are clear evidence of rising sea level or sinking land; sites with a clear maritime focus located well away from the shore, on the other hand,

indicate lowered sea level or rising land. These sites can often be dated, using either cultural or geophysical means, with more precision and certainty than natural deposits, allowing geologists to pinpoint the changes of interest to them more precisely.

In a major review, Owen Mason (1993) looked at the geoarchaeology of *beach ridges* (*q.v.*) and cheniers on prograding coastlines worldwide. He looked at studies in which archaeologists and geologists working cooperatively achieved a thorough understanding of the chronology of these features as well as those in which a lack of cooperation between the two has led to less satisfactory results for both. In using archaeological sites to date beach ridge formation, one must first carefully evaluate cultural adaptation of the past people to be certain that site location is closely tied to the position of the sea relative to the inhabited ridge.

An example of the study of rising sea levels is that of Morhange et al. (1996), which used archaeological, biological, and sedimentary data to model the sea level at Marsielles over the past 4000 years. In this area, the archaeological indicators of sea level include objects such as wrecks, which were originally deposited below sea level, the walls of quays and wharf pilings found at sea level, and roads and room floors which were originally above sea level. Using these urban features in combination with barnacles and beach indicators led to a detailed and well supported sea-level record. On a much broader scale, Flemming (1998) reviews drowned sites worldwide that indicate Paleolithic, Neolithic, and Bronze Age sea levels.

Archaeological sites can also provide evidence of falling sea level. In a noncivilized, shell midden context on the Northwest Coast, Cannon (2000) used coring to recover samples from the surface to the base of the cultural deposits in a series of sites. By dating the basal cultural deposits and establishing their relationship to present sea level, he was able to identify steady sea level decline over the past 10,000 years in this region.

Relative sea levels may also change in a complex and irregular manner. In the Shumagin Islands, Alaska, where episodic tectonic uplift is coupled with interseismic downwarping and postglacial sea-level rise over the past 10,000 years, archaeological site locations served to constrain geological modeling, since occupied sites could not be located underwater (Winslow and Johnson 1989).

Sclerochronology and oxygen isotope studies have provided major insights into climate change through time (Andrus 2011). Among recent studies, four looked at areas of the Atlantic. Helama and Hood (2011) used sclerochronology in combination with older methods to study a midden in north Norway, attributing the observed changes to the North Atlantic Oscillation. Also in 2011, Wannemaker and colleagues used oxygen isotope values of shells to examine changes in seawater temperatures during the Medieval Climate Anomaly and the Little Ice Age. In 2012,

two articles came out using oxygen isotope ratios to examine climate change in Scotland. Wang et al. looked at climate changes between the Neoglacial and the Roman Warm Period in northwest Scotland, while Surge and Barrett discovered higher seasonality and cooler winters during early medieval times on Orkney.

In the eastern Pacific Ocean, Carré et al. (2014) used oxygen isotope variation to reconstruct the El Niño–Southern Oscillation (ENSO) for the past 10,000 years using mollusk shells from middens in Peru.

Intensive Studies

Book-length compilations in English have recently appeared on the Indo-Pacific (Ono et al. 2014), Europe (Milner et al. 2007), Northwest Coast of North America (Fedje and Mathewes 2005; Moss and Cannon 2011), California (Arnold 2004; Raab et al. 2009; Jones and Waugh 1995), Southeastern US (Dunbar 2016; Marquardt and Walker 2013), Mexico (Foster 2017), Shetland Island (Moore and Wilson 2014), Ireland (O’Sullivan and Breen 2017), and Maine (Bourque 2012). In addition to these areal studies, there have also been a number of topical books: *Human Impacts* (Rick and Erlandson 2008), *Cultural Dynamics of Shell Midden Sites* (Roksandic et al. 2014), *Shells as Coastal Resources* (Bailey et al. 2013), *Paleolandscapes of the Continental Shelf* (Bailey et al. 2017; Evans et al. 2014), *Building Resilience* (Van de Noort 2013), *Archaeology of Maritime Landscapes* (Ford 2012), and *Changing Coastlines* (Bicho et al. 2011).

Finally, *Archaeolomalacology* (2005), edited by Daniella Bar-Yosef Mayer, presents articles about sites from the Americas, Europe, and Asia which discuss various uses of shellfish and their shells.

The North Pacific coast of North America and the Channel Islands and adjacent coast of California are areas that have seen intensive research on coastal adaptations.

North Pacific

Integrated studies of North Pacific prehistory are descendants of the nineteenth-century studies of Arctic cultures crystallized in Gjessing’s formulation of the Circumpolar Stone Age concept. Early studies of northern maritime adaptations were much too broad and general and have been replaced by more focused studies of particular regions. In the north Pacific, the opening of the “Ice Curtain” led to increased contact between New World and Russian archaeologists interested in the North Pacific. As these studies progressed, links to more southern Pacific cultures became obvious, drawing in Japanese scholars, who have studied both their own and New World cultures. A major focus of these studies has been the inhabitation of the Americas in the terminal *Pleistocene Epoch* (q.v.) (Erlandson et al. 2007; Erlandson and Braje 2011). Additional

studies have been concerned with the issues of culture contact and influence around the North Pacific Basin, with initial adaptations to the coastline, both littoral and maritime, and with interrelationships between Pacific coastal peoples and their interior neighbors (Campbell and Butler 2010a). The question of the degree to which Northwest Coast people managed their resources has come to prominence in the last few years (Jackley et al. 2016), as has the resilience of Northwest Coast adaptations to climate change (Campbell and Butler 2010b). See also the books mentioned above.

Channel Islands

A nearby area in which there has been an explosion of archaeological research in the last few decades is the Channel Islands area in southern California. In this region, prehistorically, there was a mainland adaptation, which often, if not always, included both coastal and inland facies, and an island adaptation, which, perforce, was always primarily littoral or maritime, though sometimes provided with inland resources through trade with the mainland people. Archaeologists and geologists working in this area have been trying to determine environmental changes in the area, how these changes affected the marine and shore-based resources, and what effect these changes had on cultural development and change. Also of concern is whether the whole area can be treated as a unit for paleoecological and paleocultural studies or whether subdivisions, between the southern and northern islands or between the islands and the mainland, are significant enough to make conclusions derived from one zone inapplicable in the other. In the process of arguing these various positions, local scholars have significantly advanced our understanding of the factors involved in both ecological and cultural development in highly variable coastal environments (cf. Kennett and Kennett 2000).

These studies have continued apace. In addition to the ones noted in various sections above, there are Reeder-Myers’s (2014) study asking why variation exists in the islands, Glassow’s (2015) study of red abalone middens, Jazwa et al.’s (2015) study of resource and settlement variability on the northern islands, and Jones et al.’s study of various marine resources (2016), and later (2017) study of Morrow Bay on the mainland as a refugium during the Medieval Climatic Anomaly.

Future Work

“Interdisciplinary efforts to reconstruct shoreline history and associated environmental parameters could significantly raise the potential of locating early postglacial archaeological sites” (Fedje and Christensen 1999, p. 650). More modeling of past environments, both on large- and small-scales, is necessary,

as well as additional detailed studies of regions and sites. More bathymetric studies are needed in many areas: in coastal Alaska, for example, bathymetric charts are very general and not useful for detailed coastal studies, even where they do not meander off into dotted lines. Given the maritime focus that the majority of coastal peoples developed, sooner or later, and the freedom of movement boats provided, regional studies are essential (Waselkov 1997). As detailed understanding of local sea level variation increases, we will be better able to understand how prehistoric people adapted to and used these changes in cultural development.

Conclusions

Absence of evidence is not necessarily evidence of absence, particularly in coastal environments. A dictum all coastal archaeologists should repeat when going to bed and upon arising in the morning: Consider the absent! In modeling past coastal subsistence/settlement systems, archaeologists must take account of sites that are gone! What we don't know will hurt us! Archaeologists should use all methods available, particularly modern tools like *Geographic Information Systems* (*q. v.*), to predict site loss and to get a handle on how much of the archaeological record is actually left in any particular area. Prehistoric coastal peoples saw their settlements disappear in storms just as present coastal people do, and the effect of these disappearances on the archaeological record is something coastal archaeologists ignore at their peril.

Cross-References

- Beach Erosion
- Changing Sea Levels
- Coastal Changes, Gradual
- Coastal Changes, Rapid
- Coastal Subsidence
- Coastline Changes
- Conservation of Coastal Sites
- Continental Shelves
- Demography of Coastal Populations
- Dune Ridges
- Geochronology
- Holocene Epoch
- Human Impact on Coasts
- Late Quaternary Marine Transgression
- Sea-Level and Climate Change
- Seismic Displacement
- Submerged Coasts
- Uplift Coasts

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Archaeology and Sea-Level Change

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Definition

The archaeological remains that can be used, better together with biological, geomorphological, or other sea-level indicators, in order to obtain accurate information regarding their relationship with sea level during the period they were functional may be called archaeological sea-level indicators.

Introduction

The primary publications regarding the assessment of sea-level changes through the interpretation of archaeological indicators came from Flemming (1969), Caputo and Pieri (1976), and Pirazzoli (1976). Despite the large number of important archaeological remains, only a small percentage can be used in order to obtain accurate information regarding their relationship with sea level during the period they were functional. The constraints arising result from their uncertain use and lack of preservation. Two general types may provide evidence for past sea levels. The first refers to those constructions that were probably terrestrial or unrelated to sea level, but today are submerged, and the second to coastal remains that were constructed taking into account the sea-level of that specific time period. The sunken paintings of Cave Cosquer (France) belong to the first category and provide the earliest known archaeological remains, which are associated with the changing sea level in the Mediterranean (Lambeck and Bard 2000). Such findings can only provide a constraint on sea level and they are generally considered as indicators of low accuracy. The second category is especially valuable and

includes shipbuilding berths, fish tanks, jetties, and harbors, which were mainly built after 2500 BC (Flemming 1978). These types of remains are usually directly related to sea level and they are able to provide accurate data for relative sea level (RSL) reconstructions. Some studies have used more indirect indicators, such as wells (e.g., Sivan et al. 2004, Caesarea, Israel), and churches' floors, in order to reconstruct the changes of the water table in coastal areas.

The ancient habitation in the Mediterranean has left a rich archive along its coasts, and it is clear that some archaeological structures provide interesting data regarding the size of relative sea-level changes since antiquity. During the last century, a large number of authors have explored the use of archaeological indicators in the study of relative sea-level changes (e.g., Flemming 1969; Pirazzoli 1976; Blackman 1973; Galili et al. 1988, Morhange et al. 2013; Sivan et al. 2004; Lambeck et al. 2004; Auriemma and Solinas 2009; Evelpidou et al. 2012; Kazmer et al. 2016, among others). The archaeological evidence associated with sea-level changes is rich globally, including the Atlantic Europe, the Mediterranean and the Near East, India, and China.

According to Morhange and Marriner (2015), archaeological indicators are usually inadequate in determining relative sea-level changes; however, when they are combined with fixed bio-indicators, their value can be significantly increased (Fig. 1).

The analysis of both port facilities and biological organisms, found on coastal constructions, has been long recognized as a possible source of information on sea-level data (Laborel and Laborel-Deguen 1994), such as the Roman

columns in the market of Pozzuoli in southern Italy. Where it was possible to establish accurate vertical correlation between the archaeological construction and the biological sea level, the relative sea-level trends were accurately reconstructed since antiquity, in some Mediterranean locations (Morhange and Marriner 2015). By transferring the techniques developed on rocky shores (Laborel and Laborel-Deguen 1994) in ancient harbors, it is possible to accurately calculate the paleo-sea level (Fig. 1). Biological remains have enabled the use of radiocarbon dating and, also, the biological zonation of specific species (e.g., the upper limit of the populations of *Balanus* spp., *L. lithophaga*, *Vermetus triqueter*, *Chama griphoides*) is empirical and associated with the biological mean sea level (Laborel and Laborel-Deguen 1994; Morhange and Marriner 2015). Measuring the highest altitude variation between fossilized and modern populations, vertical error limits of ± 5 cm can be taken into account.

Tectonic uplift (e.g., the port of Falasarna in Western Crete, around 365 BC) and silting of sedimentary basins, as in ancient harbors, contribute to the preservation of marine organisms in archaeological remains and in their use, afterward, as sea-level index points. The accuracy of the measurements depends on the determination of a reliable reference point (e.g., current biological sea level).

Recent archaeological studies, which combine bio-indicators and archaeological findings, allowed the progress in the measurements of sea-level changes in various archaeological sites, such as Marseille and Frejus in France; in

Archaeology and Sea-Level

Change, Fig. 1 The combination of archaeological and biological indicators has raised light to the palaeogeography of Lechaion area (Morhange et al. 2012)



Pozzuoli (Morhange et al. 2006), Italy; and in Vis, Croatia. Multiproxy approaches combining archaeological, biological, geomorphological, and geological sea-level indicators allow for more precise RSL reconstructions.

Functional Heights and Other Considerations

Auriemma and Solinas (2009) presented a very interesting review regarding archaeological sea-level indicators. Many different archaeological constructions, which were initially above sea level or in contact with sea water, are today submerged, thus revealing a relative change between the sea level and the position of the construction.

In the Eastern Mediterranean, coastal and submerged archaeological remains are widely used as indicators of Holocene relative sea level (RSL). Sea-level indicators include fishponds (e.g., Auriemma and Solinas 2009; Evelpidou et al. 2012), harbor infrastructure such as quays, submerged settlements, coastal wells (e.g., Sivan et al. 2004), or any other datable remains, where a functional height can be determined. According to Morhange and Marriner (2015), the functional height is defined as the “elevation of specific architectural parts with respect to an averaged sea-level position at the time of their construction.” The functional use of an archaeological construction is determined by specific parts of it, taking into account the local mean sea level and depends on the type of the construction, its use, and the local tidal amplitudes. Jetty surfaces, for example, can be estimated at an elevation of at least 1 m above sea level, the bollards of the ships at 0.7–1 m and up to 2 m for large harbors, depending on the size of the ship. Slipways, which exist in many areas of the Mediterranean, such as Sicily, Cyprus, and Turkey, need to have enough length, so that the ship can be plucked out of the water and, also, enough water depth at their base, for the ship to get into the water (Blackman 1973).

The concept of functional height is analogous to the indicative meaning (Shennan et al. 2015). If the relationship between archaeological remains and sea level is clear, a sea-level index point can be produced, using the functional height to reconstruct the former sea level. In the case of an unclear relationship between the sea level and the archaeological remain, only an upper or lower constraint on sea level may be achieved.

Also, a large number of archaeological sea-level data derive from older publications, and the assumptions used to determine functional elevations, age, and uncertainties are not always consistent.

The functional heights set the minimum height of the construction over the highest local tide (Lambeck et al. 2004; Evelpidou et al. 2012). However, in practice, this is very difficult to achieve, as their functional heights and their corresponding errors are estimated on the basis of the current

analogues, which are not always related to ancient constructions. Furthermore, there is a great variety of ancient coastal remains that have undergone significant erosion due to wave action. Errors of the order of tens of centimeters are often larger than the absolute measurement (Morhange and Marriner 2015). Several archaeological findings can be used to reconstruct sea-level changes, but with variable accuracy.

In order for these indicators to be used, the archaeological interpretation must ensure the coastal function of the construction and clarify its typology. Further estimates include building techniques, which are important elements for age estimations and functional elevations (Auriemma and Solinas 2009). Therefore, it is very important to determine the age of the construction, the period of use, and the reason of abandonment or destruction (Marriner and Morhange 2006). This evidence can be determined through archaeological excavations in the study area and geoarchaeological investigations. Emphasis should be given on multidisciplinary work, in order to determine the accurate functional height, related to the present sea level and a strict chronological framework, using ages based both on pottery and archaeological constructions (Morhange and Marriner 2015).

Ponds, Fish Tanks

Artificial Roman fish tanks were structures that provided a suitable environment for fishes. There are numerous references to Roman fish tanks in Italy, where the largest number of the best preserved remains exists until nowadays. Along the Mediterranean coast at least 80 fish tanks have been recorded. Fish tanks have been widely used for the reconstruction of sea-level changes (e.g., Pirazzoli 1976; Lambeck et al. 2004; Evelpidou et al. 2012, among many others). They are primarily found along the western peninsula of Italy, while fewer examples may be found in the Adriatic Sea, in North Africa, Spain, Greece, and Israel.

The Roman fish tanks have structural components that are directly connected with sea level at the time of their construction. Their construction and use are well documented by authors such as Pliny, Marcus Terentius Varro, and Columella. The latter one described the use of the tank as a means of storing the fish, intended for cultivation.

Fish tanks may provide information on past sea levels, but it is important to clarify the function of the individual characteristics and estimate the functional height in relation to sea level. Latin writer Columella, in the first century AD, distinguished three types of fish tanks, with variable depth. Depending on the type, the functional heights of their morphological features can be calculated, according to the sea level. Specifically, the fish tanks that had been opened on rocky platforms had a depth of about 9 ft (2.745 m) (Columella, *De Re Rustica*, XVII). These structures,

Archaeology and Sea-Level Change, Fig. 2 Some archaeological remains, such as Roman fish tanks, have structural components that are directly connected with sea level at the time of their construction



often consisting of a few built walls and including channels, started from the outer limit of the platform and had a depth of 2 ft (0.61 m) near the fish tank. Some examples of this type are the fish tanks of Torre Valdaliga and Mattonara in Italy (Fig. 2).

The fish tanks that were built entirely on the coast had a 7-ft depth (2.135 m). Examples of this type are Piscina di Lucullo, Punta della Vipera, Formia, T. Astura, La Banca, and Fosso Guardiole Saracca (A), also in Italy. Finally, the third type was similar to the previous one, but it had a depth of only 2 ft (0.61 m), because it was intended for flat fish, such as soles. An example of this type is a small arched fish tank in Fosso Guardiole (B) in Italy.

Architectural Characteristics

The foot walks are narrow paths along the inner tanks. Initially, they were used for maintenance purposes and therefore they are considered to have been above mean sea level. In some cases, such as the fish tank of Lucullus (Circeo National Park, Italy), lower foot walks were built beneath the openings for the water's entrance (Pirazzoli 1976). The measurement of the foot walk position in relation to sea level provides some evidence for the range of RSL variation.

The channels were used in order to fill and empty water from the basins using the local tide. They should, therefore, be studied and used carefully.

The in situ mid-tidal gates (closing gates), which are precise indicators, are extremely rare, due to their original location in the surf zone. The metal grids (*cataractae*) along the channels or between the tanks were operating within opened grooves on the rocks. *Cataractae* were placed within the tidal range levels and consequently, their marks can be particularly

useful, if interpreted correctly, for the estimation of the paleo-sea level.

This type of indicator is quite complex, and different researchers (Lambeck et al. 2004; Evelpidou et al. 2012) have calculated different sea levels for the same fish tanks in Italy. When dating relies on indirect means such as pottery typology, outlier data points result.

Channels and Quarries

Rock-carved structures are defective. Quarries exist along many coastal areas in the Mediterranean and elsewhere, but in most cases direct dating is not possible (Fig. 3). According to Auriemma and Solinas (2009), the height of the mining surface of the blocks in coastal quarries brings a higher margin of error. For coastal quarries, which are now found at least partially submerged, the assumption that they were completely emerged in order to enable mining is considered (Auriemma and Solinas 2009). Morhange and Marriner (2015) suggest that this type of indicator can provide information only on the direction of sea-level change.

Harbor Structures and Shipheds

Harbor structures refer to moles, quays, harbor towers, docks, shipheds, etc. These structures are frequently problematic. It is difficult to decipher the function for many of these features since they frequently consist of rock mounds in the water; their state of preservation adds further uncertainties. If the preservation state is poor and eroded,

Archaeology and Sea-Level Change, Fig. 3 Direct dating of quarries is not possible; other nearby indicators may be used in combination



Archaeology and Sea-Level Change, Fig. 4 Moles are frequently problematic since they consist of mounds of rocks in the water that raise many uncertainties related to their preservation



reconstruction of past sea level is not possible. Even in the case of well-preserved and well-dated quay remains, a vertical uncertainty of ± 0.5 m is very common (Auriemma and Solinas 2009), since interpretations rely on assumptions

regarding the draughts of ancient ships and unloading techniques. In such types of indicators only moderate accuracy may be achieved and only few provide upper constraints on sea level.

Archaeology and Sea-Level Change, Fig. 5 Submerged buildings, roads, and similar structures may only provide a constraint on sea level and thus are considered as low accuracy indicators



Within harbors, rock-carved shipsheds, cannot be directly dated, so they can only be correlated with available nearby sea-level indicators (Fig. 4). However, this is also a difficult task, considering that such structures are sometimes incomplete, and their relationship to sea level depends on assumptions regarding ancient maritime technology.

Building Foundations and Floor Surfaces

The base level of many structures, including foundations and floor surfaces from roads, buildings, walls, etc. (e.g., Auriemma and Solinas 2009) is a common sea-level indicator. However, such archaeological remains can only give a maximum constraint for sea level (Fig. 5). Furthermore, their interpretation often relies on various speculations, e.g., how close they could be built above water (e.g., Morhange and Marriner 2015).

Coastal Wells

Coastal water wells have been very useful indicators of RSL in Israel due to a large, continuous dataset (e.g., Sivan et al. 2004). According to Sivan et al. (2004) and references within, the reconstruction of sea level is accomplished from the base level elevation of the well, plus 35 cm for a water drawing jar, minus the thickness of the freshwater table resting on the

saltwater intrusion. Uncertainties in age estimation of wells may also arise; however, in the case of well-studied archaeological sites, such as Caesarea, domestic wells are part of well-understood and accurately dated context (Sivan et al. 2004).

Summary

Archaeological remains may help in studying sea-level fluctuations since the prehistoric times. There are many different types of archaeological sea-level indicators with variable accuracy and precision. Among them, one of the most prominent are fish tanks, providing RSL data since the Roman times. Archaeological indicators alone are usually inadequate in determining relative sea-level changes with a high level of precision. However, multiproxy approaches, combining geomorphological, geological, or biological indicators, can allow not only to reliably reconstruct RSL changes but also coastal landscape evolution, providing archaeologists with valuable data related to former human occupation.

Cross-References

- [Changing Sea Levels](#)
- [Eustasy and Sea Level](#)
- [Holocene Coastal Geomorphology](#)

- Paleocoastlines
- Sea-Level Indicators, Biologic
- Sea-Level Indicators, Geomorphic
- Submerged Coasts

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Arctic, Coastal Ecology

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Unlike Antarctica, which is an ice-covered continental plateau surrounded by oceans, the Arctic is made up of a central ocean nearly enclosed by land. This entry will mainly focus on the coastal ecology of the Canadian Arctic (Nunavut and Northwest Territories), northern Alaska (United States), Norway, Greenland, and Iceland, as well as the islands of the Barents, Kara, Laptev, and Siberian seas (Fig. 1). The intertidal zone, the benthic, and pelagic subtidal communities and the ecosystems associated with the ice itself will be discussed, as well as the ecological importance of polynyas, which are characteristic habitats of the Arctic.

Basic Characteristics of the Arctic

The major marine ecosystems of the Arctic include the High Arctic oceanic region, which comprises the deep sea; the High Arctic coastal region between the abyss and the shore; the High Arctic brackish water subregion surrounding the shores of northern Russia, Canada, and Alaska, where the fauna must adjust to seasonal changes in temperature and salinity; and the boreal littoral region of Norway influenced by warmer currents which favor a faunal penetration from the south. Lastly, the Low Arctic shallow region marks the transitional area between the boreal Norwegian fauna and the High Arctic fauna.

The Arctic coastlines of Greenland, northern Iceland, Svalbard, northern Norway, Novaya Zemlya Islands, Murman coast, eastern Siberia, and Arctic archipelago of Canada are generally steep and deeply cut by fjords. However, the northern coast of Alaska, the Canadian Northwest Territories mainland, and most of the Arctic Russian coast are low relief coastal plains. The southern part of the Arctic coasts is characterized by the tundra biome where only grasses, mosses, lichens and about 400 flowering plants grow on a permanently frozen soil during short summers. Most of the biomass is in the root systems. The northernmost portions of Arctic landmasses, consisting mainly of the northeastern islands of Nunavut (Canada) and northern and central Greenland, lie under a permanent ice and snow cover.

Arctic, Coastal Ecology,

Fig. 1 Continental boundaries, major islands and seas of the Arctic. The approximate limits of the Arctic circle ($66^{\circ} 30'N$) are shown in white

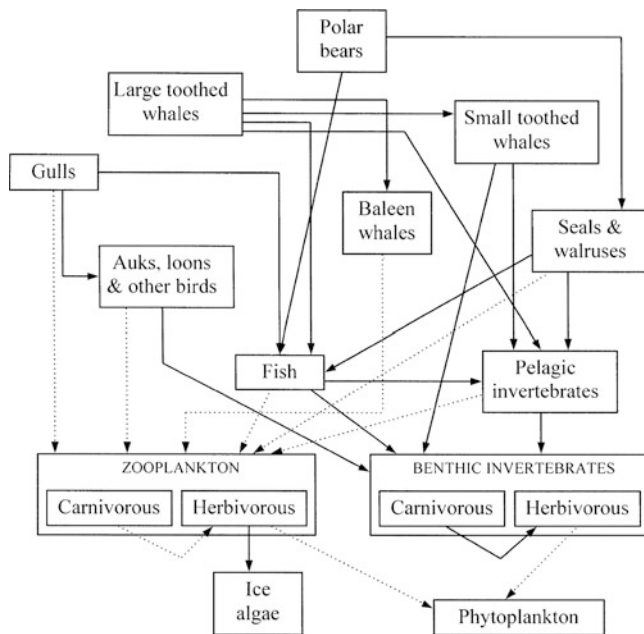


Overall, much of the Arctic coastline is ice-covered for about 7 months a year and the water temperature generally remains below $0^{\circ}C$. Air temperatures as low as -65 to $-68^{\circ}C$ have been recorded in Arctic Russia and Canada, and precipitation in the Arctic generally averages less than 250 mm per year. Melting and refreezing processes, wind-chill, runoff from the land and ice movement generate temperature and salinity variations in the upper water layer. The extreme seasonal oscillation in light is an additional factor affecting the biotic components of the Arctic ecosystem.

Due to its particular structure as a semi-enclosed sea (Fig. 1), the Arctic Ocean is fairly isolated from the other major oceanic masses, except through the Bering Strait, the Denmark Strait, and the Barents and Norwegian Seas. However, no invertebrate genus is endemic either to the Low or High Arctic. This suggests that most Arctic genera are distributed worldwide or at least found in the Atlantic, the Pacific, or both. Moreover, 77% of the Arctic genera are also common to the Antarctic.

The marine Arctic environment is generally perceived as stressful to the fauna and flora that live there (Menzies 1975). Low temperature is apparently not the primary factor that has shaped the Arctic marine ecosystem, but rather the marked seasonal oscillation of other physical components. Among the most important characteristics of the Arctic are reduced light, seasonal darkness, prolonged low temperature and icy period, which result in impoverished fauna and flora and low seasonal productivity. Contrary to habitats of lower latitudes, including coral reefs, where biodiversity is, generally, inversely proportional to the respective biomass of each species present, the Arctic is characterized by the dominance of a few species both in abundance and biomass.

From mid-November to late January, the sun is not seen above the horizon in the far North, whereas from mid-May to late July, the sun does not set. However, owing to its geographical position, the Arctic receives an overall low input from the slant illumination of the sun in summer. Moreover, on the polar sea itself, the permanent pack ice reflects 60–90% of the meager



Arctic, Coastal Ecology, Fig. 2 Schematic Arctic food chain

incoming radiation. When there is constant heat during the summer, widespread melting of snow and ice occurs, uncovering the seas and land surfaces. The weak organic production is in particular contrast to the high productivity of adjacent subarctic and boreal regions. Production in Arctic waters is probably below an average of 50 g of carbon per square meter of sea surface a year, and mainly concentrated in summer. Phytoplankton peaks in August in northern Canada, while polar zooplankton counts increase by tenfold in July compared with values in the winter dark. Because of this seasonal production, fish feeding on the lower elements of the food chain must glean most of their food in a short period and be able to subsist on small amounts of ingested food and stored energy for long periods. The phases of high productivity and feeding coincide with higher temperatures, which allow higher metabolic rates and levels of activity. Alternatively, when productivity is low, colder temperatures occur and hence, metabolism and energy consumption are low. When and where resource turnover and availability are sufficiently high, the systems may be relatively productive and sustain complex trophic levels, composed of numerous production and consumer organisms (Fig. 2).

Supralittoral, Littoral, and Sublittoral Zones

A bar of coarse sediment can form by waves or ice scouring on certain Arctic shores of Alaska and Canada. Pools of brackish water and typical marsh vegetation, including *Carex* spp., *Puccinellia*, mats of mosses, blue-green algae and diatoms, often occur behind these beaches. These supralittoral Arctic

marshes can sustain populations of cladocerans, copepods, branchio-pods, dipteran larvae and oligochaete worms, numerous insects, shorebirds, and caribou herds (Broad 1982).

The intertidal and shallow subtidal areas are relatively barren, at least on the surface. For instance, the Arctic coasts of Alaska and western Canada have no littoral flora and only sparse vegetation in the supralittoral (Broad 1982). The virtual absence of intertidal fauna, which results in the absence of a fauna in the upper littoral zone, may be attributed to the freezing air and low humidity. However, the barrenness is usually explained, according to many authors, by the mechanical action of ice on the shores and/or to the weak tidal amplitude itself.

When wind, current, and tide drive ice floes against the shore, bottom sediments, rocks, and benthic fauna and flora are mixed, overturned, and ground together. The resulting clouds of mud may extinguish the life of nearby organisms under a blanket of mud and smother filter-feeders hundreds of meters away. Consequently, when they are present in the Arctic, the intertidal species found in the Atlantic or Pacific are usually located in the subtidal area.

While sparse, the littoral fauna of the High Arctic is characteristic. In Alaska, the shore to a depth of about 2 m, the approximate depth of seasonal ice, is probably colonized by less than 50 species of macrobenthos, mainly oligochaete worms, midge larvae, amphipods (*Gammarus setosus* and *Onisimus littoralis*) and the isopod *Saduria entomon*, which form nearly the entire biomass ($3 \pm 5 \text{ g m}^{-2}$; Broad 1982). The sublittoral fauna beyond the depth reached by ice is more diverse, including species such as mysid shrimps (*Mysis relicta* and *Neomysis rayii*), polychaete worms (*Scolecopides arcticus*, *Ampharete vega*, *Prionospio cirrifer*, *Terebellides stroemi* and others), bivalve mollusks (*Cyrtodaria kurriana* and *Lyocyma fluctuosa*), the priapulid *Halicryptus spinulosus*, amphipods (*Pontoporeia affinis* and *Calliopius laevisculus*), and the four-horned sculpin *Myoxocephalus quadricornis* (Broad 1982). The boundaries between Arctic and Subarctic shores of northern Canada and Greenland can be determined by the sudden disappearance in the northern parts of three common invertebrates, *Mytilus edulis*, *Littorina saxatilis* var. *groenlandica*, and *Balanus balanoides*. The last two are the most decisive indicators as their disappearance is more abrupt. In northern Iceland, the littoral zone is characterized by the succession of different species of *Fucus*, from *Fucus spiralis* in the upper section to *Fucus edentatus* in the lower littoral (Munda 1991).

The sublittoral of Alaskan shores harbors benthic macroalgae, including three species of Chlorophyta, which infrequently grow in the littoral zone, 14 species of Phaeophyta and 11 species of Rhodophyta. The intertidal and subtidal zones are feeding grounds for about 30 species of fish, including anadromous salmonids, typically marine

gadids, pleuronectids, cottids and a few freshwater species (Broad 1982).

The Arctic shores of eastern Greenland are similar to those of Alaska. A few algae occur on rocks with *Fucus inflata* in the lower littoral. The fauna of the littoral consists of mostly oligochaete worms (*Enchytreus albidus* and *Lumbricetta lineatus*), mites (*Molgus littoralis* and *Ameronothrus lineatus*) and sublittoral crustaceans (*Gammarus wilkitzkii* and *Mysis oculata*) (Broad 1982). Warm currents along some coasts of Norway, Iceland, Greenland, and other regions of the Barents and Norwegian Seas promote the occurrence of a boreal biota along Arctic shores. In such environments, the most common species of the mid-littoral and continuing subtidal are the blue mussel *M. edulis*, the gastropods *L. saxatilis*, with some *Buccinum groenlandicum*, *Margarita helcina* and crustaceans of the genera *Balanus*, *Gammarus*, *Caprella* and *Ischyrocerus*. Some species, such as the polychaete worm *Spirorbis spirium* and certain gastropods also colonize floating thallus of *Fucus* and other macroalgae. Jorgensen et al. (1999) indicated that 387 species were observed in the Kara Sea, Arctic Russia, with predominance for polychaetes, crustaceans and mollusks. Boreal-Arctic species clearly dominate this area. The sedimentation rate, as well as depth, sediment structure, and salinity apparently influence the faunal distribution.

Studies performed in southern Baffin Island indicated that numerous pelagic and benthic invertebrates were found below ice-covered sea. The fauna and flora in the long lasting ice cover, the small tides and the reduced wave action allow fine sediments to settle and produce low oxygen concentrations near the bottom. The brittle star *Ophiura sarsi*, shrimp *Sclerocrangon boreas*, sea cucumber *Chiridota laevis*, starfish *Crossaster papposus* and *Leptasterias groenlandicus* are observed commonly. Rocks deposited by ice and human artifacts provide a scattered hard substrate, which supports a kelp community of *Laminaria* and *Agarum*. Barnacles, ascidians, sea cucumbers, sea anemones, gastropods are also observed abundantly on rocky bottoms.

Bluhm et al. (1998) indicated that the sea urchin *Strongylocentrotus pallidus* was considered an important part of the benthic standing stock and carbon flux in the northern Barents Sea. Holte and Gulliksen (1998) noted that macrofaunal dominant species in the sediment of the north Norwegian coast were detritivorous and carnivorous polychaetes, and detritivorous bivalves. Sponges, bivalves, polychaetes, and nematodes are common in the Svalbard waters (Strömberg 1989).

Primary Production (Phytoplankton)

The algae that grow on the under-surface of the ice, the phytoplankton and the benthic macroalgae, produce biomass

that feeds numerous animals. Together, these plants are called the primary producers. Among them, phytoplankton produces the majority of the edible material, although ice-associated algae are of equal importance in some areas (see “Sub-Ice Flora and Fauna” section below).

The onset of ice cover and the loss of winter solar radiation have the clearest impact on the primary production in the Arctic. Although, the level of incident radiation in the polar summer is high, much of it is lost by the high reflectivity from the snow and ice. This restricts the production and leads to extreme seasonal oscillation in both phytoplankton and herbivorous zooplankton. The onset of seasonal production is clearly linked to the melting of snow from the ice surface and the subsequent melting of the ice. The annual primary production in polar regions varies from one area to another and ranges from 5 to 290 g C m⁻² (Sakshaug 1989). Blooms are formed mainly in spring or summer when nutrient supply is adequate and vertical mixing is restricted to the upper 30–50 m. They are particularly frequent at the ice edge because melting generates water column stability (Alexander 1980). In the northern waters of Iceland, spring growth of phytoplankton begins in late March and culminates during mid-April (Gislason and Astthorsson 1998).

According to Zenkevitch (1963), more than 200 species of phytoplankton are present in the Barents Sea. About 80% are diatoms, 20% are dinoflagellates and flagellates. During winter and early spring, phytoplankton biomass is small (<0.5 mg chlorophyll *a* m⁻³) and mainly composed of small flagellates. The spring bloom gives place to the growth of larger groups, such as diatoms and dinoflagellates. During summer, two main composition patterns are found. One is dominated by microflagellates in the upper oligotrophic layers and diatoms and other large groups dominate the other, in the sub-surface chlorophyll maximum. Toward autumn, the chlorophyll sub-surface maximum tends to disappear and most of the phytoplankton comprises microflagellates (Loeng 1989).

Small flagellates dominate the first phase of the phytoplankton bloom in ice-covered water. The main spring bloom occurs in the surface layer close to the ice edge and is usually dominated by diatoms, although flagellates can also be abundant. Toward the end of the diatom bloom, the maximum phytoplankton biomass is found at the nutricline, located below the pycnocline. After the bloom, the major part of the diatoms sink out of the euphotic zone (Loeng 1989).

Strömberg (1989) indicated that as the ice melts in the Svalbard waters, a very active pelagic primary and secondary production begins, which follows the ice-edge as it retreats. Horsted (1989) noted that there are generally two maxima of primary production from April and from August to September in Greenland waters.

Sub-Ice Flora and Fauna

Wherever a sufficient amount of light penetrates the ice and enough nutrients are present, ice algae develop. These algae make ice-covered seas unique. The ice-edge phenomena are closely linked to the cryo-opelagic organisms. Thus, one of the short food chains is: primary production (ice-algae, phytoplankton), sea ice amphipods (*G. wilkitzkii*, *Apherusa glacialis*), polar cod (*Boreogadus saida*), sea birds, or mammals. The biomass of the cryopelagic fauna is high close to the ice-edge and decreases with increasing distance from it.

Daily and seasonal melting and formation of the ice has a thermostatic effect, so that the temperature of the water immediately below the ice is almost constant. It is also often of low salinity, hence the ice biota have evolved as euryhaline and stenothermal organisms. Mel'nikov (1980) distinguishes an endemic or autochthonous flora, consisting of Chlorophyta and Cyanophyta, present year-round and developing each spring in the snow above the ice and gradually leaching downward through the ice; and a non-endemic flora dominated by diatoms. This latter flora develops at the bottom of the ice, is planktonic in origin, and is carried upward as the ice thickens in the fall. He indicated that when diatoms die, they become a source of food for the endemic species of flora. The associated fauna is endemic, *G. wilkitzkii*, *Mysis polaris*, *Derjugina tolli*, *Tisbe furcata* among others, or non-endemic.

The brown zone forming the bottom layer of sea ice is dominated by pennate diatoms, but contains other algae and several species of small invertebrates. Chlorophyll measurements in Arctic Canada show that production of the ice diatoms peaks in mid-June before the phytoplankton blooms in the water below the ice, at a time when the sea ice is still snow-covered (Mansfield 1975). As soon as the snow melts this ice flora disappears, suggesting that it is adapted to low light intensities. It extends the season of production by preceding the normal phytoplanktonic bloom. These algae also provide a source of food for benthos and fish in coastal waters, especially with the cods *B. saida* and *Arctogadus glacialis* (Mansfield 1975).

Werner and Arbizu (1999) noted that the lower part of the ice in the Laptev Sea, northern Russia, showed an accumulation of organic matter, mainly composed of organic carbon and chlorophyll, that could provide a food source to the fauna living below the ice in the pelagic habitat, and to the underlying benthos. In fact, they found that large quantities of nauplii, copepods, foraminifers and pteropods were common. Poltermann (1998) found that the Franz Josef Land islands sea ice was inhabited by several amphipods species. Abundance, biomass and small-scale distribution of these cryopelagic amphipods reached 420 ind m⁻². Amphipods were concentrated at the edges of the ice floes and were less frequent in areas further away under the ice. Species

such as *G. wilkitzkii* were dominant with the less abundant *Apherusa glacialis*, *Onisimus nansenii*, and *Onisimus glacialis*.

Secondary Production (Zooplankton)

Water near the surface of the Arctic Ocean supports the greatest summer biomass of zooplankton, the abundance generally decreasing with increasing depth. At least 158 species of zooplankton are known to be found in the central Arctic and adjacent seas including Kara, Laptev, East Siberian, and Beaufort. Crustacea, Ctenophora, Mollusca, Annelida, Ostracoda, and Chaetognatha are the most common (Grainger 1989).

Like phytoplankton, zooplankton shows an extreme annual oscillation in Arctic Canada. This is mainly due to the increase in populations of the herbivorous species, particularly copepods, which are the dominant species. As phytoplankton grows, copepods grow and reproduce, and then as their food supply comes to an end, they gradually die off. In contrast, the primarily carnivorous zooplankton species show relatively little change in numbers throughout the year and no particular period of reproduction (Mansfield 1975).

Gislason and Astthorsson (1998) indicated that the seasonal variation in biomass and abundance of zooplankton in the waters north of Iceland were low during the winter and peaked once during spring in late May. The principal constituents were *Calanus finmarchicus*, *Pseudocalanus* spp., *Metridia longa*, *C. hyperboreus*, chaetognaths and euphausiid larvae. Kosobokova et al. (1997) studied the composition and distribution of zooplankton in the Laptev Sea where total biomass ranged between 0.1 and 7.9 g m⁻². The dominant species were *Calanus glacialis*, *Calanus hyperboreus*, and *M. longa*. Copepods are also the most important grazers in the Barents Sea with ca. 38 species, *C. glacialis*, *C. finmarchicus*, and *C. hyperboreus* being among the most common (Loeng 1989). Euphausiids as *Thysanoessa* spp. are the most important grazers. Among carnivorous zooplankton species, the chaetognaths *Sagitta elegans* and *Eukrohnia hamata*, together with ctenophores are considered to be important (Loeng 1989).

Dunbar (1989) and Grainger (1989) pointed out that diel vertical migration of zooplankton in the Arctic Ocean is rare, but seasonal vertical migration is fairly common. In the Barents Sea, larger zooplankton species, such as *C. glacialis*, are thought to overwinter in deeper layers, returning to the surface layer in early spring to mature and spawn. Spawning takes place during the diatom bloom and the new generation feeds on the remnants of this bloom and subsequent production of phytoplankton. Zooplankton in turn is preyed on by capelin, which during summer has a northward feeding migration (Loeng 1989).

Zenkevitch (1963) characterized the Svalbard water area in the same way as the Barents Sea, with a dominance of copepods *C. finmarchicus* making up 90% of the zooplankton biomass. The plankton biomass was reported to decrease in a northerly and easterly direction. In the area north of Svalbard, chaetognaths, ctenophores, and amphipods comprise most of the remaining biomass (Strömberg 1989).

Horsted (1989) observed that zooplankton biomass in Greenland waters had its maximum in July at the same time as the second maximum of primary production, while macroplankton had a somewhat longer lasting maximum from June–July onward. Thus, the newly hatched cod larvae seem to depend heavily upon the availability of naupliar and copepodid stages of *C. finmarchicus*.

Polynyas

In the Arctic, there are ice-free areas in the proximity of completely ice-covered regions, so called polynyas. The correlation between areas of open water in ice-covered Arctic and increased biological productivity is clear (Bazely and Jefferies 1997; Stirling 1997).

Polynyas form where warm, upwelling sea currents prevent the water surface from freezing. They are either small and temporary or extensive and free of ice all winter. The largest ones appear at the same place every year. The biggest of all polynyas is located at the head of Baffin Bay, Canada, and can reach a surface area as large as Lake Superior. In the desert of floating ice, polynyas represent a rich feeding ground and ideal habitat for many tiny plants and animals at the base of the food web. They also represent a temperate environment that enhances survival of Arctic animals. For instance, the ice around polynyas is ideal for the growth of ice algae. Owing to this large amount of food, countless amphipods and larger crustaceans, squids, fishes and mammals can be observed. Ice algae develop soon in spring, as soon as the sun rises high enough to promote minimum photosynthesis, long before the weather begins to warm up. This early start of primary production and its effect on other levels of the food chain is important for seabirds that need time to build up their strength before they can lay their eggs. Moreover, since seabirds can only hunt for food in open water, they are largely dependent upon polynyas. Among them are fulmars, murre, guillemots, and gulls. Marine mammals such as whales, seals, polar bears, and walrus also converge at polynyas to take opportunity of the abundance of food. If no patches of open water existed in the winter pack ice, many animals could not survive.

Fish

There are no schooling pelagic fish, though the cryopelagic Atlantic cod (*Gadus morhua*) is known to occur in large

numbers during the spring period in the Barents Sea, and the polar cod (*B. saida*) appears to form scattering layers. The giant among benthic fish is the Greenland shark (*Somniosus microcephalus*) long known for its scavenging behavior largely oriented on whale carcasses. Where the Bering Sea intrudes into the Beaufort Sea, abundant pelagic herrings (*Clupea harengus*) and capelins (*Mallotus villosus*) occur in the region of the Mackenzie delta. Demersal species such as the Arctic and starry flounders (*Liopsetta glacialis* and *Platichthys stellatus*) and the Greenland and saffron cods (*Gadus ogac* and *Eleginus gracialis*) are also common. Similarities exist in the southern part of Baffin Island where the Atlantic cod and several deepwater forms like the Greenland halibut (*Reinhardtius hippoglossoides*) and the rock and roughhead grenadiers (*Coryphaenoides rupestris* and *Macrourus berglax*) are observed. The Barents Sea is a feeding ground and nursery for large commercially important fish stocks such as Arctic cods, herrings, Greenland halibuts, haddock, capelins, redfish, and saithe (Loeng 1989).

Arctic fish must be able to live under conditions of reduced light and in total darkness. Air breathing predators, except for certain seal species, are displaced to gaps, leads and polynyas through interference with respiration. Ice also reduces avian predation to practically zero. In these conditions, ice may be considered advantageous for fish. Moreover, a new biological habitat that might be described as an inverted benthos is provided by ice for fish. A distinct phytoplankton flora occurs immediately below the ice and an abundant crustacean fauna may be tributary to this food source. This diatom-crustaceans community is an important one that supports the Arctic cod, which helps to sustain Arctic char, birds, seals, and belugas.

Birds

About 12 species of birds are known to live all year round in the Arctic, including the raven, gyrfalcon, willow and rock ptarmigans, snowy owl, redpoll, Ross's and ivory gulls, thick-billed murre, dovekie and the black guillemot. Most of them shift south slightly during the Arctic winter. All other birds, about 90 species, are migrants. Numerous species of birds nest in the Arctic, such as the fulmar *Fulmarus glacialis* and the kittiwake *Rissa tridactyla*. In the Greenland Sea, the most abundant species are the fulmar, little auk *Alle alle*, guillemot *Uria lomvia*, and kittiwake (Mehlum 1997). The abundance of organisms that bloom during summer, including insect larvae, mollusks and crustaceans, are a substantial source of food. Lack of food rather than the cold itself is what makes the Arctic so inhospitable for birds in winter. Nevertheless, considering that the cold is fierce, some of the permanent residents have evolved special adaptation to cope with it.

Mammals

The generally scattered distribution of marine fish in Arctic waters has not prevented mammals from attaining a dominant position among the marine vertebrates, for they depend mainly on the larger invertebrates, both planktonic and benthic, for their food. Thus, most common mammals of the Arctic are either part of or dependent upon coastal ecosystems. They include polar bear, arctic foxes, wolverines, reindeer, musk oxen, lemmings, and many species of marine mammals.

Five species of seals spend at least part of the year in the Arctic. The bearded seal (*Erignathus barbatus*) and ringed seal (*Pusa hispida*) remain all year round. The others are the harp seal, hooded seal, and harbor seal. Except for the harbor seal, all species give birth to their pups on the ice. Most seals are important predators of benthic invertebrates and fish. The most thoroughly investigated species has been the ringed seal, the most truly Arctic of all the mammals. This species is able to survive under the winter fast ice by keeping open breathing holes. In spring, the pup is born in a cavern among rafted ice blocks or on the surface of the ice in a lair excavated in the overlying snow. Predation by the polar bear (*Ursus maritimus*) and Arctic fox (*Alopex lagopus*) can cause very high mortality of young in certain areas. In fact, the polar bears feed almost exclusively on seals. The ability of ringed seals to live under fast ice is not shared by any other Arctic pinnipeds, though the bearded seal is known to keep breathing holes in some areas.

Walrus (*Odobenus rosmarus*) live in the Arctic waters east of Somerset Island, notably in Baffin Bay, Davis Strait and Foxe Basin. Pacific walrus are found along the coast of Alaska and westward along the Arctic shores of Siberia. Walrus give birth on ice floes and, when not feeding, spend most of their time lolling on the ice floes in great numbers. Most of them are bottom feeders grazing on clams and others bivalves in shallow waters. They also eat whelks, sea cucumbers, and other benthic invertebrates.

Four species of whales may be encountered in Arctic waters. Three of them, the bowhead whale (*Balaena mysticetus*), beluga (*Delphinapterus leucas*), and narwhal (*Monodon monoceros*) are permanent residents. Like the walrus, narwhals and belugas are gregarious. Narwhals are found in large herds in the fjords of eastern and northern Baffin Island, Devon Island and Ellesmere Island. These whales occur in the High Arctic waters of the northern hemisphere, where they occupy the most northerly habitat of any cetacean species. The fourth whale species, encountered in summer, is the killer whale (*Orcinus orca*). Other species, such as right, gray, blue, fin, minke, and humpback whales move to the Arctic during their normal annual migration cycles (Gambell 1989). Toothed whales like sperm, Baird's beaked, northern bottlenose, long-finned pilot, and beaked whales, among others, can also be observed in the Arctic.

The bowhead whale is the only plankton eater, its diet is focused on euphausiids. The beluga and narwhal mostly prey on fish and squids as well as pelagic and benthic invertebrates. Killer whales largely prey on other marine mammals.

Human Impact

The native people of the Arctic are at the top of the food chain and harvest fish, birds, and marine and terrestrial mammals. Inuits have their homeland stretched from the northern tip of Russia across Alaska and northern Canada to parts of Greenland, but their populations remain sparse.

In spite of its remoteness, the Arctic is threatened by airborne pollutants spreading from industrial areas in the lower latitudes, by growing mineral exploitations (coal, nickel, uranium, tin, gold) and also by natural gas and petroleum industries. As a result, Arctic ecosystems and Inuit populations are seriously endangered.

Moreover, during the course of the past century, there has been an overall increase in global temperature of some 0.5 °C. The past decade has been the warmest of the past 100 years with the five warmest individual years having been recorded during it. The recent trends observed are generally consistent with those projected by the global circulation models under increased atmospheric CO₂ concentrations (Maxwell 1997). Higher temperature is generally associated with less sea ice and snow cover extent. Thus, global warming could be one of the most important factors affecting the biodiversity, distribution, abundance, and seasonal cycles of Arctic species in the future.

Cross-References

- ▶ [Antarctica, Coastal Ecology and Geomorphology](#)
- ▶ [Arctic, Coastal Geomorphology](#)
- ▶ [Ice-Bordered Coasts](#)
- ▶ [Littoral](#)

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Arctic, Coastal Geomorphology

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Introduction

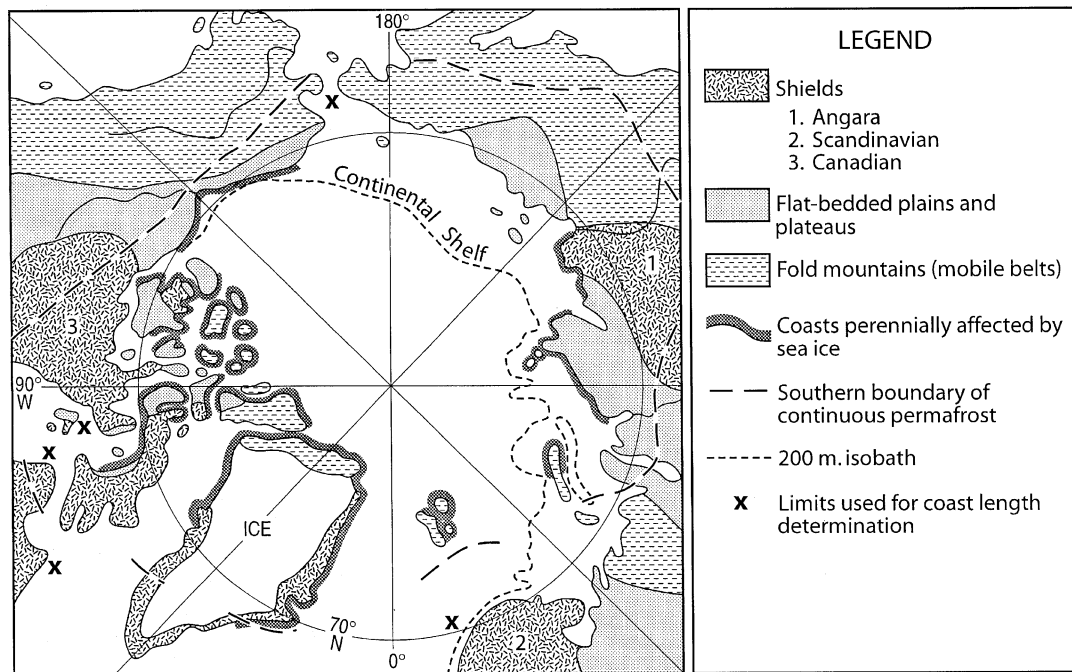
The Arctic, long considered a “... region of darkness and mists, where sea, land and sky were merged into a congealed mass” (Nansen 1911, p. 1), has subsequently been defined according to its astronomic, biotic, climatic, cryologic, geomorphic, and hydrologic characteristics (Walker 1983). From the standpoint of coastal morphology, it is the juncture of whichever category is used with the coastline that is important. The determinant that provides the greatest extent to the coastline is sea ice. In the Northern Hemisphere using sea ice as the limiting boundary results in the inclusion of the coastlines of Hudson Bay, Labrador, and Newfoundland, the Sea of Okhotsk and most of the Baltic and Bering Seas (See Ice-Bordered Coasts). By many other criteria, most of these coastlines do not qualify as Arctic.

In this entry, the discussion centers on those coasts that border the Arctic Ocean and surround Greenland and the islands of the Canadian Archipelago. When the lengths of these coastlines are measured on the American Geographical Society 1:5,000,000 *Map of the Arctic Region*, they total 82,000 km, Greenland (16,000 km), the Canadian Archipelago (26,000 km) and the other arctic islands (9000 km) (Fig. 1). Other calculations, depending on the detail used, provide larger numbers. For example, Bird (1985) states that the coastline of the islands of the Canadian Archipelago is at least 90,000 km long and Nielsen (1985) reports that that of Greenland is some 40,000 km long.

The Geologic Base

Three large, stable shields composed of Pre-Cambrian rocks form the cornerstones of the Arctic, one each in Canada, Scandinavia, and central Siberia (Fig. 1). The actual exposed lengths of these Pre-Cambrian rocks along the coast is less than what the size of the shields would lead one to believe because their margins have been buried under eroded shield materials. Between and adjacent to the shields, including their buried margins, are folded mountains (or portions of the so-called “mobile belts” of the Northern Hemisphere) some of which extend to the coast (Fig. 1).

H. Jesse Walker: deceased.



Arctic, Coastal Geomorphology, Fig. 1 Geologic base, continental shelf and sea ice distribution in the Arctic. (Compiled from numerous sources including Sater et al. (1971), Péwé (1983), and Walker (1998))

Like the embedded coasts surrounding the Atlantic Ocean, Arctic coastal margins are affected by minimal tectonic activity. Earthquakes do occur, as on the north coast of Baffin Island, north of the Mackenzie delta and in the northwest Canadian Archipelago, but they are of low magnitude.

The most recent major event to impact most of the arctic coastal zone is glaciation. The main coastal areas not directly affected by glacial ice during the Pleistocene include the east-central Siberian lowland, a small part of the western Canadian Archipelago and the north coast of Alaska. However, even those coastal zones, though never actually covered by ice sheets, were affected by drainage across them from glacial fronts and by the changes in location of the interface between land and sea due to changes in sea level as the continental glaciers waxed and waned. Today, only a small percentage of the arctic coastline is being modified by glaciers. Major segments are found on Greenland, Ellesmere Island, Novaya Zembya, and Spitzbergen.

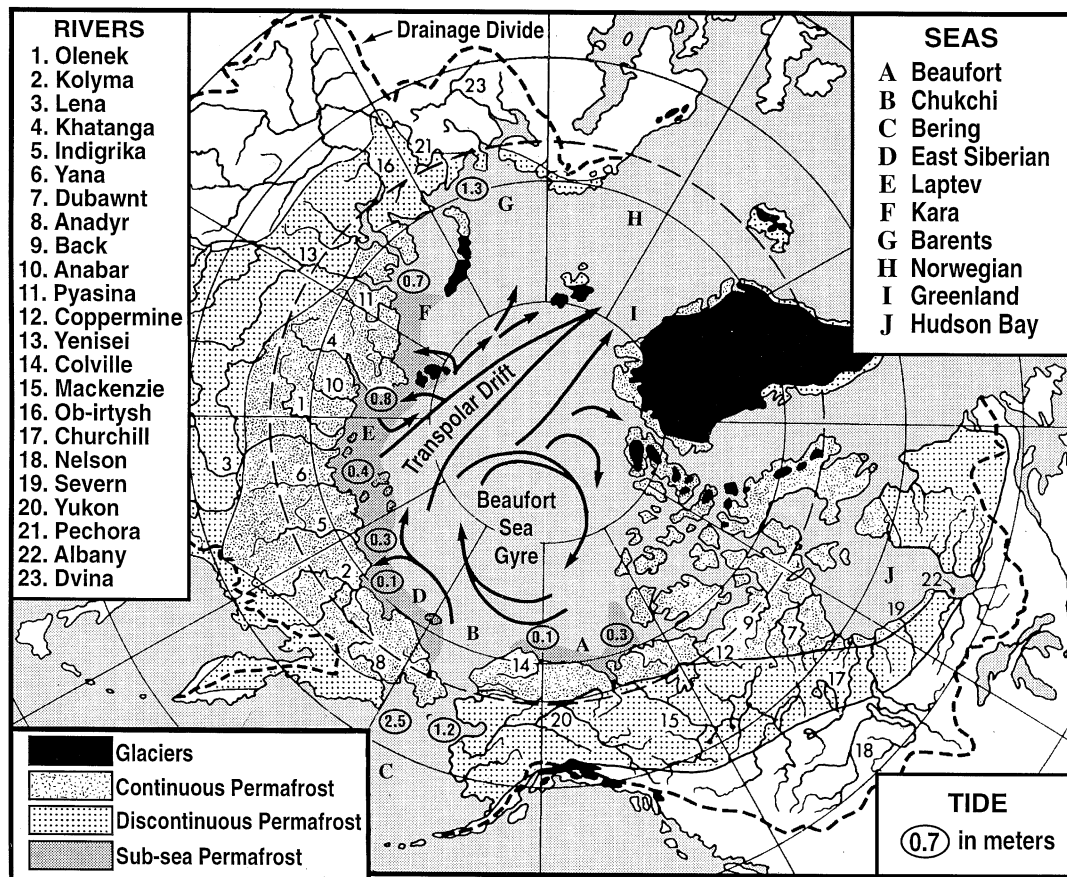
In addition to the direct modifications of the coastlines by glaciers in the form of moraines, drumlins, fiords and strandflats, the rebound that followed deglaciation has (and is) converted formerly submerged coastal belts into subaerial coastal plains. Such occurrences are especially common in northern Canada, the Canadian Archipelago, Scandinavia and the islands north of Siberia. In many locations former coastal features are found today at elevations of as much as 250 m.

In the Canadian Archipelago and the Hudson Bay area rebound is continuing at rates of as much as 1 m/century (Andrews 1970).

The Continental Shelf and the Coastal Plain

The two basic contrasting zones of coastal significance in the Arctic are its subaerial and subaqueous portions. These two zones are highly variable in width and in the forms and processes they possess. The continental shelf varies in width from a few kilometers, as off parts of Greenland, to more than 800 km in the East Siberian Sea. With sea-level rise and with coastal erosion, the width of the subaqueous portions of the Arctic Ocean has been increasing. Weber (1989, p. 815), for example, writes that "The Chukchi Shelf was eroded far into the continent and transects the principal mountain ranges of Northern Alaska." Much of the continental shelf of the Arctic Ocean is flat and shallow, cut only by a few submarine valleys such as those off rivers like the Kolyma and Indigirka and off Barrow, Alaska. Most of the smaller features on the shelves are the result of ice gouging and deltaic deposition or are remnants of subaerial erosion that occurred during lower sea levels (Reimnitz et al. 1988; Weber 1989).

The coastal plain, although not as extensive as the continental shelf, possesses a greater variety of forms and processes than the shelf it borders. Although, the coastal plain of



Arctic, Coastal Geomorphology, Fig. 2 The Arctic illustrating the numerous factors that affect the coast. (Compiled from numerous sources including NOAA (1981), Lewis (1982), Péwé (1983), and Walker (1998))

today is subaerial much of it, as in Alaska and northwestern Canada, is the "... landward edge of a continental shelf that has experienced repeated transgressions and withdrawals of the sea ..." (Bird 1985, p. 243). It is comprised mainly of gravels, sands, and silts that presently are ice-bonded in the permafrost. In places it is low-lying and level. Near the Indigirka River mouth, for example, storm surges have reached as far as 30 km inland (Zenkovich 1985).

The major forms of the coastal plain that border the ocean include barrier islands and lagoons, sand and gravel beaches, mudflats and marshes, sand dunes, low coastal bluffs, deltas and lengthy rias (*gubas* in Russian).

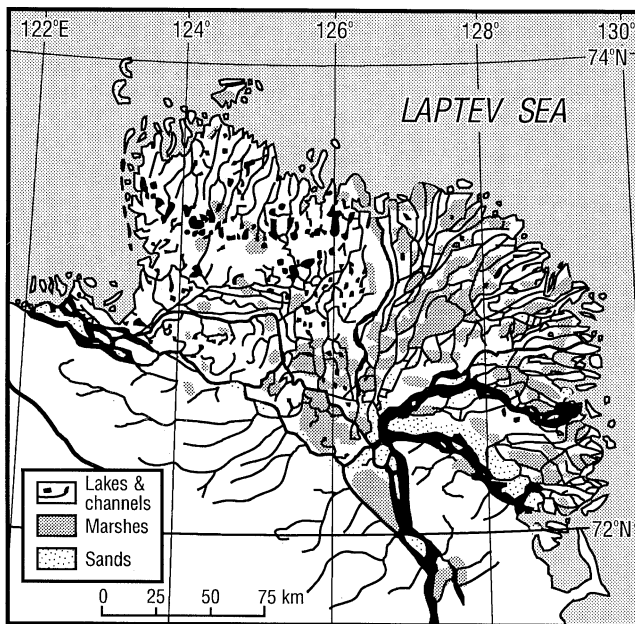
Among the arctic coast's most conspicuous and extensive features are its barrier islands and sandy spits. They are present along much of the coast of arctic Alaska and north-west Canada and along various parts of the Siberian coastline. Some barrier islands are remnants of the coastal plain whereas most are composed of gravel and sand which originated offshore.

Although, most of the deltas in the Arctic are small, they are sufficiently numerous to occupy a sizeable proportion of

the coastline. For example, 135 km (or 9%) of the coastline of Alaska between Cape Lisburne and the Canadian border is deltaic (Wiseman et al. 1973).

Some arctic deltas, like the Mackenzie, Lena, Indigirka, Kolyma, and Colville, face the open ocean, whereas others, such as the Yenisey and Ob, are located at the head of lengthy and narrow gubas (Fig. 2). These two deltas, even though confined to the head of their estuaries, are the fourth and fifth largest deltas in Siberia. Among arctic deltas, that of the Lena River (Fig. 3) is unique because its structure, shape, and relief appear to be the result of tectonic activity as well as modern hydrological processes (Are and Reimnitz 2000).

The deltas of the Arctic, like those elsewhere, are relatively young in that their present-day expression stems only from the time sea-level rise reached a nearly stillstand position about 5000 BP. All of the older deltas in the Arctic contain most of the features, such as distributaries, abandoned channels, lakes, sand bars, mudflats, and sand dunes, that are typical of deltas elsewhere. However, they also possess such cryospheric forms as ice-wedges, ice-wedge polygons, pingos, and thermokarst lakes.



Arctic, Coastal Geomorphology, Fig. 3 The Lena River delta, the largest at 32,000 km² in the Arctic



Arctic, Coastal Geomorphology, Fig. 4 Ice complex coast on Muostakh Island facing the Laptev Sea. The cliff is about 20 m high and is retreating at a rate of about 11 m/a. (Photo courtesy of Feliks Aré)

Conditions, Forms and Processes

The present-day appearance of the arctic's coastline, like that of coastlines elsewhere, depends on modifications that have occurred to the geologic base it inherited. Along many coastlines, these modifications include those engineered by humans. In the case of the Arctic, however, human modifications are still minimal, although they do occur at the mouth of some rivers, adjacent to some coastal villages, and where mining operations, including petroleum exploration and production enterprises, have been developed.

Thus, most of the coastal forms in the Arctic, as is true also of the Antarctic, are the result of natural conditions and processes. Included are those conditions and processes associated with cold climates such as low temperature, snow, ice (river and sea) and permafrost with its many forms of ground ice as well as those of more universal occurrence like relief, structure, sediment type, river discharge, offshore gradient, wave action, currents, storm surges, and tides (Fig. 2).

Cold climate processes are frequently divided into two types: glacial and periglacial (French 1989). Although in the arctic of today glacial processes impact only short lengths of shore, periglacial processes are of major importance along the entire coastline (Harper 1978) including those sections most recently deglaciated.

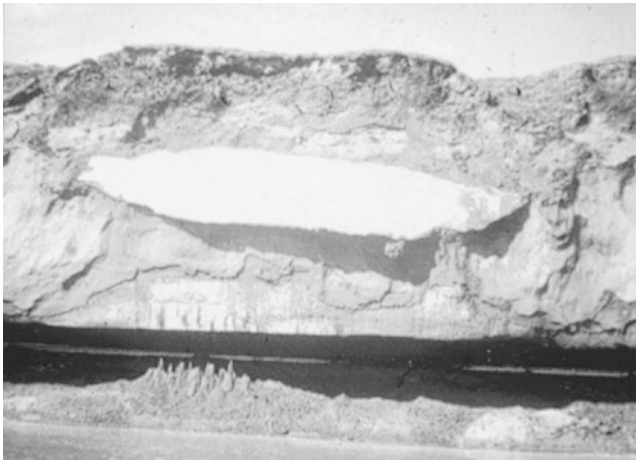
The sine qua non of periglacial processes is the freezing and thawing of the ground, which is mainly temperature

dependent. The change from one state to the other may be daily, seasonal or over longer (thousands of years in some cases) periods of time.

Permafrost

Permafrost (a condition in which ground temperature remains below 0 °C for more than 2 years) is present along most of the coastline of the Arctic from which it extends both landward and seaward (Fig. 2). Although permafrost is defined by temperature, it is water in the form of ice that is especially critical in coastal modification (Fig. 4). The included ice, which serves to bond unconsolidated sediments as long as they remain in a frozen state, makes the mass especially unstable when thawed. Ground ice is highly variable in type and quantity. In the upper few meters of permafrost it may account for more than half of the volume whereas its presence in bedrock is minimal. It can range in form from microscopic pore ice to large segregated masses (Mackay 1972).

Permafrost cliffs, which border much of the coastal plain of the Arctic, are relatively stable for 8–10 months of the year because they are protected from erosion by being in a frozen state, being covered by snowdrifts, and by facing a frozen, immobile sea. With snowmelt, an active layer (the thin layer that alternately freezes and thaws on the top of permafrost) begins to form and some surface flow is initiated. However, it is after the shore has become sufficiently free of ice that wave action can impact the base of the cliff and major erosion



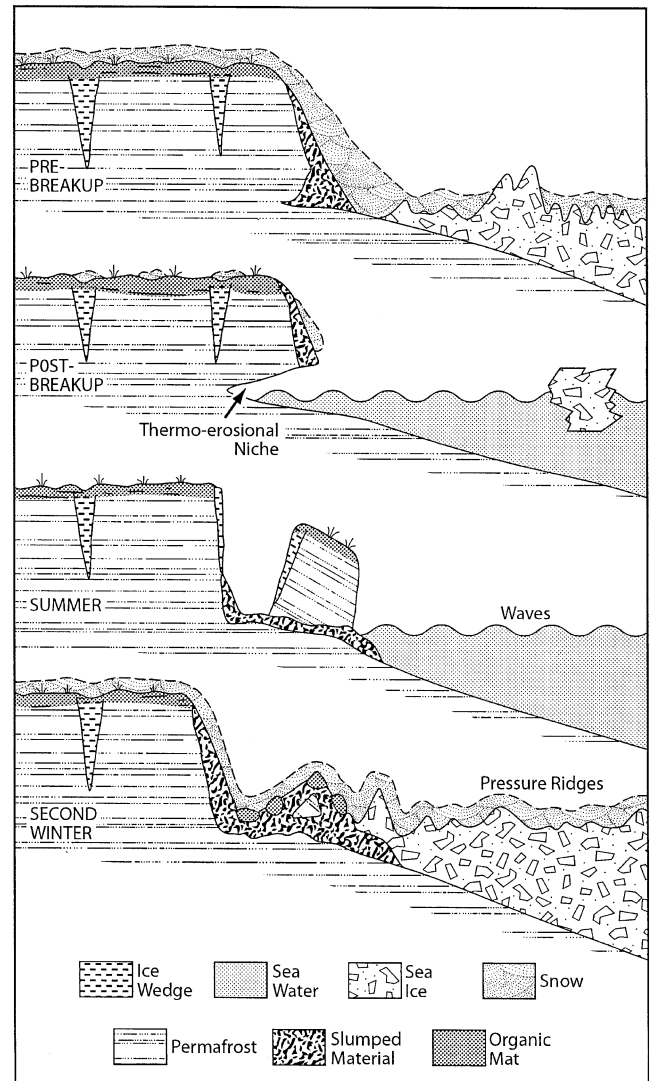
Arctic, Coastal Geomorphology, Fig. 5 Thermoerosional niche in deltaic permafrost



Arctic, Coastal Geomorphology, Fig. 6 Example of block collapse and ice wedge from E. de K. Leffingwell's classic research monograph on permafrost and coastal morphology (de Leffingwell 1919)

can occur. With wave impact, thawing at the base of the ice-bonded cliff occurs and with sediment removal a thermo-erosional niche (*termoerozionaja niza* in Russian) is created (Gusev 1952) (Fig. 5). The height of thermoerosional niches depends largely on the change in water level during wave attack and the depth of penetration varies with the stability of the sea cliffs.

The permafrost of the Arctic frequently supports other features such as pingos (Mackay 1972) and ice-wedges which often express themselves at the surface in the form of ice-wedge polygons (Lachenbruch 1962). Although their size varies, polygons are usually on the order of 10–30 m in diameter. Along the coastline, paralleling wedges serve as lines of weakness and often lead to block collapse (Figs. 6 and 7) as thermoerosional niches develop. Those wedges that are perpendicular to the shore can serve as points of more



Arctic, Coastal Geomorphology, Fig. 7 Block collapse and erosion along permafrost coastlines

rapid retreat, especially if the matrix is dominated by peat. Such coastlines are highly irregular in detail. In addition to the large block falls that result from thermoerosional niche development, many coastlines exhibit thermokarst forms such as ground-ice slumps (creating what some have called thermocirques) and breached thermokarst lakes. Along hard rock coasts (which are in the minority in the Arctic) erosion of a cliff face appears to be mainly by frost wedging (Fig. 8).

Sub-sea permafrost (offshore permafrost, submarine permafrost) occurs beneath the seabed of the continental shelves of northwest Canada, north Alaska, and Siberia (Fig. 2). It is relic in that it formed when sea level was low and much of the present-day continental shelf was subaerial. It is of major interest today because of its importance to the petroleum industry and as a store of carbon dioxide and methane (Osterkamp 2001).



Arctic, Coastal Geomorphology, Fig. 8 Bedrock shore on the Taymyr Peninsula, Siberia. (Photo courtesy of Feliks Are)

Sea Ice

Sea ice which borders virtually all of the arctic coast during winter, is present along some sections even in summer (Fig. 1). During the time it is shorefast and bottomfast marine action along the coastline is nil. Also, during summer, much of the coastline is protected from wave attack because the presence of sea ice offshore reduces the fetch available to wind and dampens wave action. However, during these periods, often in the Fall of the year when the ice pack has been removed some distance offshore, storm waves may become powerful eroding agents. In 1986, strong winds over the Chukchi and Beaufort Seas resulted in severe erosion of low coastal cliffs impacting heavily the villages of Wainwright and Barrow, Alaska. Other erosive storms also occurred along these coasts in 1954, 1956, and 1963.

Some coastal specialists consider that the passive nature of sea ice is its most important role *vis-à-vis* the coastline. Nonetheless, sea ice is also an active agent in coastline and offshore modification. Ice push during the periods of freeze up and to a lesser extent during breakup create highly irregular surfaces on arctic beaches. Large pressure ridges formed during freeze up help protect the coastline during breakup because they may last well into the Fall. Although some of the reworked beach forms, including mounds, ridges and kettles, may last for years, depending mainly on how high up on the beach they developed, most tend to be ephemeral because they are removed by summer wave action. Large amounts of beach material (especially sand and gravel but also, in some locations, boulders) can be bulldozed, scoured, rafted, and



Arctic, Coastal Geomorphology, Fig. 9 Remnant snowdrift along Beaufort Sea in western Canada in August. Note the coarse textured beach and driftwood typical of many arctic beaches

resuspended by sea ice. In some instances, sea ice rides up on the shore and even overtops low cliffs. If rideup is over a snow bank or an ice foot it will cause minimal modification to the frozen ground beneath.

In addition to the shoreface itself, sea ice causes ice scour out to depths of more than 20 m often several kilometers from the coastline. Deep (2–3 m) gouges or trenches are created by the keels of drifting floes.

Most floes eventually become grounded in which case current movement around them can resuspend bottom sediment and transport it elsewhere.

From the standpoint of the coastline, changes occurring to sea ice because of global warming are likely to be very important in the future. Recent data show that, not only is the ice pack thinning, but its seasonal regimen is changing (Parkinson 2000). If such a trend continues, ice-free periods along coastlines will lengthen and the fetch will increase in length. Thus, both the intensity and duration of waves impacting the shore will also increase.

Snow

Although the amount of snow that falls in the Arctic is limited, partly because of low temperatures, it is an important factor in regard to the coastal zone. It accumulates as snow drifts against irregularities such as coastal bluffs and serves as protective ramps often until late summer (Fig. 9). Snow also becomes incorporated in the icefoot in combination with seawater and freshwater ice, and organic and inorganic sediment where it becomes part of the beach (McCann and Taylor 1975). Possibly the most important role of snow insofar as the arctic coastline is concerned is its service as the main source of water for the rivers that flow with their sediment load into the Arctic Ocean.

Impact of Rivers on the Arctic Coast

Four of the ten longest rivers of the world (Ob, Yenisei, Lena and Mackenzie) drain into the Arctic Ocean. In addition, there are numerous smaller rivers, some of which are confined to the zone of continuous permafrost. Many of these smaller rivers cease flow completely during winter. Even the larger rivers are highly seasonal with the major discharge of water and sediment occurring during the snowmelt season (Walker 1998). Thus, the impact of rivers on the coast is confined, like that of most other arctic coastal modifiers, to a few months of the year. Nonetheless, during that restricted season river water impacts the sea ice over and under which it flows.

Floodwater aids in earlier melting of the sea ice at the mouth of the river than otherwise would be the case. Nearshore, much of the flood water flows over the top of the bottom/shore fast ice until it finds holes, as along pressure ridges, down through which it drains. From these discharge holes, which Reimnitz and Bruder (1972) call *strudel*, water continues to flow seaward beneath the ice (Fig. 10). As the water pours through these holes in the ice, it creates depressions in the seafloor deposits below.

As sediment-laden floodwater spreads out over the sea ice, its velocity decreases and deposition occurs. Most of the sediment deposited on this bottomfast ice becomes a part of the subaqueous delta as the sea ice melts from under it (Walker 1974). The sediment that is carried by floodwaters beneath the sea ice may be transported many tens of kilometers seaward and become incorporated in long-shore currents and therefore, lost to the deltas.

Ocean Currents, Tides, and Waves

"The Arctic Ocean, relatively isolated in terms of worldwide wind systems and partially or wholly covered with sea ice, depending on season, is a body of water in which low tides, slow-moving currents, and low-energy waves predominate" (Walker 1982, p. 61).

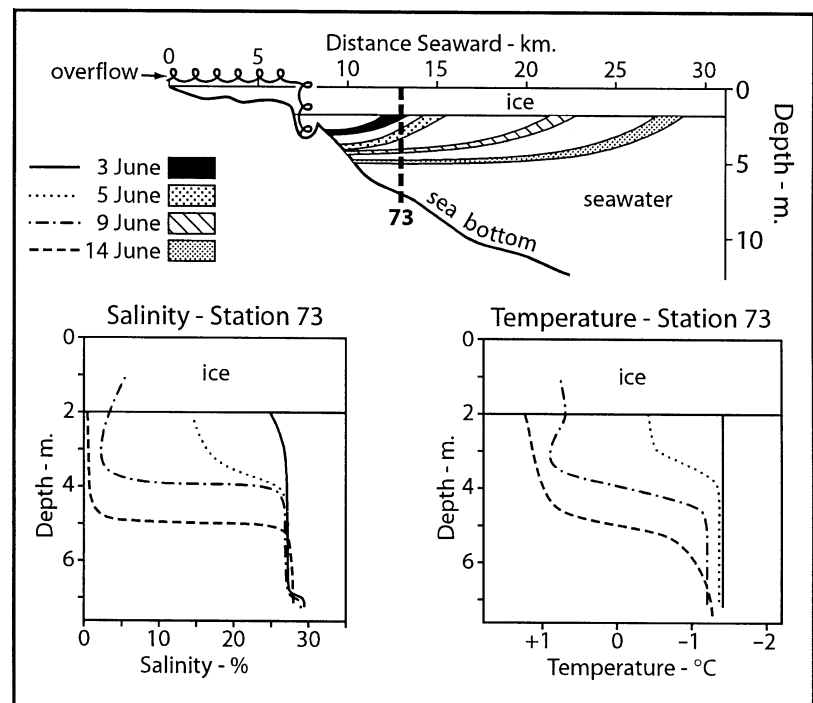
Although there are a few locations in the Arctic where macro-tides occur (e.g., off southeast Baffin Island, Canada, and in the Mezen Gulf, Russia, where ranges are more than 10 m), along most of the arctic coast tidal ranges are less than 2 m (Fig. 2). Marshes are often associated with the high tide range locations. Despite low astronomic tidal ranges, storm "tides" occasionally result in deep intrusions of water over low coastal plain surfaces. The inner edge of surge lines are often delineated by the presence of driftwood which is a common contribution of the rivers that flow through the taiga of Asia and North America.

The dominant currents in the Arctic Ocean are the Beaufort Sea Gyre and the Transpolar Drift (Fig. 2). They are the currents that provide the general direction of flow to pack ice and the occasional ice island that breaks off of Greenland and Ellesmere Island. There are other localized currents that affect the coastal zone by transporting sediments along shore and cause sea ice to impinge the coastline. Such ice movements are affected by wind as well as by ocean currents.

Waves, even though their formation is impossible for most of the year because of the canopy of sea ice that covers the ocean, are, nonetheless, the dominant agent in shore modification along most of the arctic coastline. The length of time, waves are effective, varies greatly ranging from several

Arctic, Coastal Geomorphology,

Fig. 10 Floodwater overflow and the development of a freshwater wedge beneath sea ice during breakup



months along the longest-lasting ice-free zones to only occasionally or even rarely along coasts that are often ice-bound even during summer as exemplified by parts of the northwest Canadian Archipelago.

Coastal Erosion and Climate Change

Many of the coasts of the Arctic are retreating, some by tens of meters per year especially along some Siberian coastlines (Aré 1988). Along the Beaufort Sea coast of Alaska, erosional rates have been such that since the sea reached its present level 4–5000 BP, the coastline has retreated as much as 25 km (Reimnitz et al. 1988). Rates ranging from 1 to 5 m/a are common in the ice-rich permafrost bluffs bordering much of the Arctic Ocean. In areas north of the Siberian coast, erosion has resulted in the disappearance of some offshore islands.

As most of this retreat is the result of the combined effect of normal marine (wave action) erosion and the periglacial process of thermokarst collapse and thermal erosion, it appears that an increase in the rate of coastal retreat with any rise in sea level and sea-ice degradation is likely.

Such a change would increase the amount of sediment contributed to the shore from cliff erosion which by some calculations already exceeds that contributed by the rivers flowing into the Arctic Ocean (Brown and Solomon 1999). However, it is likely that the same climate changes that affect sea level and sea-ice degradation will also increase permafrost thaw and river discharge and, therefore, the sediment loads of arctic rivers. Whether the relative proportions of the two sources will change is uncertain.

Conclusion

Although a number of localized programs have been undertaken in the last four decades, some of them out of research stations such as the Naval Arctic Research Laboratory, Barrow, Alaska; the Inuvik Research Laboratory, Inuvik, Canada; and the Permafrost Research Institute, Yakutsk, Russia, Arctic coastal research has generally been neglected.

In 1998, at the International Permafrost Conference, a Working Group on Coastal and Offshore Permafrost was formed. This led to the development of an Arctic Coastal Dynamics (ACD) project which has among its objective the establishment of rates of erosion and accumulation, the development of a network of monitoring sites along the coast and the initiation of research on critical coastal processes in the Arctic. The ACD is an international undertaking that bodes well for the future of arctic coastal research.

Cross-References

- ▶ [Antarctica, Coastal Ecology and Geomorphology](#)
- ▶ [Arctic, Coastal Ecology](#)
- ▶ [Glaciated Coasts](#)
- ▶ [Ice-Bordered Coasts](#)
- ▶ [Paraglacial Coasts](#)

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Artificial Islands

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Definition

Islands are defined as a relatively small land surface, surrounded by water. The largest island, Greenland, has a surface, which is still four times smaller than that of the smallest continent, Australia. The total surface of the earth's islands approximates 10 million km², which is comparable to Europe's total surface area.

It is common to distinguish between continental and oceanic islands. Continental islands are considered part of a continent when they are located on the associated continental shelf. Oceanic islands, in contrast, do not belong to a continental shelf. The majority of oceanic islands are of volcanic origin. Some oceanic islands (e.g., Madagascar, Greenland, Zealandia) could be considered as subcontinents, which have been separated from larger continents through tectonic processes.

All realizations of artificial or man-made islands as constructed so far are close to shores on relatively shallow water either in intertidal zones, in bays, on the shoreface, or on the nearby shallow continental shelf, and most are located in a sheltered environment. Based on the above definition, artificial islands belong to the class of continental islands. Because of their general close vicinity to the shore, artificial

islands are very similar to land reclamations into the sea, certainly when such reclamation is separated from the existing land by a waterway or channel, which often has a water management purpose.

The history of island and land reclamation into the sea is some 2,000 years old, starting with dwelling mounds in flood-prone areas (cf. Kraus 1996; this reference provides historic context of works in Canada, Germany, Japan, the Netherlands, and Taiwan, described below). In the sixteenth and seventeenth centuries, land reclamations were becoming common practice in Western Europe. In the twentieth century, this practice has evolved worldwide. While this continues to be the case, the construction of islands or reclamations further offshore is only of recent times. Islands in support of a larger infrastructure or temporary islands for petroleum exploitation are of the last decades. Many feasibility studies on artificial islands with an intrinsic purpose (such as industry, urbanization, and airports) have been undertaken all over the world since the 1970s. However, it was not until 1994 that the first offshore-located island was completed for the realization of Kansai International Airport in the Bay of Kobe of Japan. Most recently, strategic tensions in the South China Sea also known as the East Sea has led to nourishments increasing the surface areas of very small islands that are not necessarily shoreface connected.

Concepts and Applications

Dwelling Mounds

Historically, artificial islands and reclamations go back a long way (cf. Van Veen 1962). The first known historic recordings of artificial islands are by the Romans. Plinius in 47 AD describes how the Friesians along the northern boundaries of the Roman Empire lived on artificial mounds (called “terpen” in Dutch) in order to keep “their heads above the water” during spring tides and storm surges. In all they built 1,260 of these mounds in the northeastern part of the Netherlands, which still exist today, although by now situated in reclaimed land. The areas of the mounds vary from 2 to 16 ha, and their surface level may reach as high as 10 m above mean sea level. The volume content of a single mound may be up to one million cubic meters (Mm³).

Polder-type Reclamations

This period of family-based shelter against high waters reverted into a more collective form of shelter by seawalls and associated reclamations. The start of these constructions is expected to have been shortly after the declaration of the Lex Frisonian in 802. In the sixteenth and seventeenth centuries all over Western Europe, including Russia, polder-type reclamations have been realized both on seashores, estuaries and rivers, and lakes. A polder that is used to reclaim part of the sea bottom consists of an endiked continental shelf or

intertidal basin area, with its surface, being the original sea bottom, below mean sea level which is maintained dry by the use of pumping stations (see entry on “► [Polders](#)”).

One of the world's largest polder reclamations (Van de Ven 1993) concerns that of the Zuiderzee in the Netherlands, which started in 1918 (with the main damming realized in 1932 and the principle of constructing small polders first to learn by doing) and was finalized in the early 1970s when it was decided to not reclaim the last planned polder (Markermeer Polder). The closing off and the partial reclamation of the Zuiderzee have resulted in the gain of 166,000 ha of new land distributed over four polders. This new land is used for agriculture, urban development, recreation, and nature conservation.

Although the polders in the Bay of San Francisco are – intriguingly – called “islands,” these are basically real polders with their ground elevation below mean sea level. These solutions are also introduced in Saemangeum, South Korea; in the Po Delta, Italy; and very recently in Jiangsu Province, China.

Petroleum Exploration and Exploitation Islands

In the petroleum exploration and exploitation industry, there exist several examples of island reclamations. One such example concerns Rincon Island off California (U.S. Army Corps of Engineers 1984). Another important example concerns a series of artificial exploration drilling rig islands of temporary nature in the Beaufort Sea – McKenzie Delta region, Canada. Through the construction of a number of such islands in the period 1973–1986, the presence of oil and gas reserves was confirmed, but exploitation in this remote, arctic environment was costly. The decline of the oil prices had hampered exploitation. The islands, with an expected lifetime of 3 years, were constructed in water depths varying between 3 and 21 m, just before the frost and ice formation would set in. A total of eight exploration islands of the sacrificial beach type and nine exploration islands of the sandbag-retained type were constructed. In both cases, a dense sediment core was placed, which provided the exploration space needed (surface areas of the order of 1 ha). The sacrificial beach type implied that the relatively steep slope of the core was protected with a gently sloping lower beach of less dense sediment, which would be able to withstand erosive forces by water (in summer) and by ice (in winter). The sandbag-retained type resolved the issue of required resistance by protecting the slopes of the core through geotextile sandbags (with volumes of 1.5–3 m³). The issue of subsidence in this area is virtually absent because of the permafrost conditions.

Elevated Reclamations

The strong economic development of Asian countries over the last decades where the availability of land is scarce, for example, Japan, Singapore, Taiwan, and South Korea, has

also resulted there in a series of reclamations. With the exception of South Korea, the acceptance of the polder-type reclamation in Asia has shown to be low. Although the execution method initially is similar, viz., endiking the area first and then drying by pumping, there appears to exist a preference in Asia for elevating the reclamation to above mean sea level (see entry on “► [Reclamation](#)”).

Examples of this practice are the recent realizations (mid-1990s) of the industrial reclamation estates Chang Hua and Yun-Lin on the west coast of Taiwan, comprising 3,000 ha and 10,000 ha, respectively. The fill of these reclamations amounted to 800 mm³ of sand, which was dredged by trailing suction hopper dredgers in the nearby offshore area. The reclamation locations were primarily on diluvial substrates formed by pre-Holocene ebb-tidal deltas, which implies that subsidence problems were virtually absent.

A recent (2017) mind changer in Singapore is the Pulau Tekong reclamation with 810 ha which is being executed as a polder, whereas all earlier reclamations were executed with the ground level above mean sea level. The main reason behind this is the difficulty to arrange access to sand resources of the neighboring countries.

Infrastructure Supporting Islands

Over the last decades, a number of islands have been reclaimed to support the construction of a larger infrastructure. An early example is the island Neeltje Jans in the mouth of the Eastern Scheldt estuary. This was created on a subtidal flat, separating the main estuarine channels, with a twofold purpose. The island served both as a working area for construction of the elements that were to form the Eastern Scheldt Storm Surge Barrier (completed in 1988) and to form part of the barrier itself.

Other examples concern the construction of islands in cases where the two banks of a waterway are connected by a combined bridge-tunnel connection. At the transition of the bridge to tunnel and vice versa, the island serves as the connection area. One North-American example is the Chesapeake Bay bridge-tunnel connection; a European example is the Øresund bridge-tunnel connection between Denmark and Sweden.

Airport Islands

It might be stated that the introduction of offshore artificial islands in larger depths with a ground level above mean sea level was benchmarked by the construction of Kansai International Airport, Japan (1994). Chek Lap Kok International Airport, Hong Kong (1998), and Incheon International Airport, South Korea (2001), followed up soon. A common problem to these three airport islands is formed by the foundation. Being international airports demands that important loading forces (400 metric tons for future intercontinental aircraft) need to be sustained. Since all of these islands are

constructed in relatively sheltered areas, the local sea bottom commonly consists of alluvial clay, which calls for special measures.

Kansai International Airport, Japan: The main concern in creating the artificial island has been the foundation of the island. Not only was the local water depth some 18 m but also an alluvial clay layer of more than 20 m covered the geotechnically stable diluvial (Pleistocene) clay substrate. The alluvial clay layer was artificially compacted by about one million piles that were driven into the layer to drain the water out. Subsequently, a sea defense was constructed of some 11 km circumference, within which 178 mm³ of material was dumped. Because of the scarce availability of suitable sand dredge material, the majority of the fill was taken from quarries near Osaka. Thus, an island was created which has an elevation of 33 m above the sea bottom. Since the construction, the subsidence has continued and is expected to continue another 30–50 years. In the next phase, the surface area of the island will be extended.

Chek Lap Kok International Airport, Hong Kong: A quarter of the artificial island consists of the island of Chek Lap Kok (350 ha); three quarters consist of sea reclamation. The average local depth was some 6 m. In contrast with the solutions for Kansai and Inchon, the alluvial clay layer of 10–15 m was removed. The reclamation consisted of a lower layer of rubble mound material quarried from Chek Lap Kok and supplemented with sand dredged in the nearby area. To avoid expensive layer gradation, transitions between the rubble mound material and the sand geotextiles were applied. The fill quantities amounted to 350 Mm³, of which two-third was dredged sand.

Inchon International Airport, South Korea: This airport site covers 5,600 ha in total, of which 4,700 ha consists of reclaimed land between the islands of Yeongjong and Yongyu along with 900 ha of existing land. To endike the area between the islands, three dams of a total length of 17 km were constructed, varying in crest height between 7.5 and 9.4 m. Their widths at their foundations are 90–120 m and 20 m at the crest. The reclamation of the 4,700 ha required approximately 180 Mm³ of fill. Around 80% of the fill was dredged from the sea, and the remaining 20% was quarried from mountains and hills in the area. The top layer of the reclamation site, which is largely an intertidal area, is composed of a soft alluvial layer of 5 m average thickness. A sand drain technique was applied to achieve higher soil-carrying capacity. Vertical sand pipes of 400 mm diameter were driven into the soft soil every 2.8–3.8 m to drain pore water. The expected additional settlement is less than 0.5 m.

Conclusions

Although the history of artificial islands or island reclamations is long, it may be expected that the increasing pressure

of urbanization, industry, and tourism in the densely populated coastal regions is yet to result in a further boom of construction of islands and island reclamations, even though the constructed Palm Islands (Dubai) and the World Islands (Dubai) are still struggling to turn into a success. The three main technical problems that one will generally face in the design and execution are those of geotechnical foundation, of connection to the mainland, and of availability of fill material. Besides this, one faces the important issue of environmental impact, where the aspects of impact minimization, nature substitution, and working in harmony with natural system forces are the keywords.

Cross-References

- [Dredging of Coastal Environments](#)
- [Geotextile Applications](#)
- [Polders](#)
- [Reclamation](#)
- [Small Islands](#)
- [Storm Surge](#)

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Asia, Eastern, Coastal Ecology

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Coastal Geography

Extending from Bangladesh in the west, to the Korean Peninsula and Siberia in the northeast (latitudes 89–129 E), Eastern Asia lies within the Indo-West Pacific Biogeographical region described by Ekman (1967). Eastern Asia contains diverse coastal land formations and habitats, ranging from rocky shores and sandy beaches, to coral reefs, seagrass beds,

salt marshes, mudflats, and mangrove swamps. The main landmass of Eastern Asia includes long, uninterrupted coastlines, for example, the coast of Rhakine (Arakan) in Myanmar, the coastlines of central Vietnam and China, and the peninsulas of Malaysia and Korea. The region also supports immense archipelagoes, most notably those making up Indonesia and the Philippines, which comprise of more than 13,700 and 7000 islands, respectively. Innumerable small islands make up the Mergui Archipelago extending from Myanmar to Thailand, the Andaman and Nicobar Islands, and several minor archipelagoes. Huge deltas have also been formed by some of the world's great rivers (the Ganges–Brahmaputra in Bangladesh, the Irrawaddy in Myanmar, the Red River and Mekong River in Vietnam, the Yangtze River and Yellow River in China).

Climate and Ecology

Climatically, Eastern Asia features a tropical equatorial subregion consisting of Malaysia, Singapore, Sumatra, Java, and Borneo; and northern subtropical to temperate subregions, which include the Bay of Bengal, the South and East China Seas, and the Yellow Sea (Fig. 1). The Malay-Indonesia region has a hot and humid climate and is bathed by shallow, highly productive seas overlying the Sunda Shelf (the submerged peninsula of Pleistocene Sundaland). Extensive coastal areas of the Sunda Shelf are less than 60 m in depth. The Philippines borders the eastern edge of the Sunda Shelf, with much deeper waters of the Pacific Ocean beyond. Although the southern archipelago of Indonesia lies mainly within the humid tropical zone, there are some subhumid to semiarid areas, particularly to the east, including the northeast coast of Java and the islands of Timor, Lombok, and Sumba.

North of the equatorial region, the NE and SW monsoons dominate the climate and surface ocean circulation. Rainfall becomes more and more seasonal eastwards and northwards from the equatorial subregion, with typhoons and cyclones (defined as storms with wind speeds exceeding 73 mph) associated particularly with the warm SW monsoon season. Typhoons originating in the Pacific pass through the Philippines before reaching the coastline of Vietnam and southern China. There is a “typhoon belt” from central Vietnam to Hainan Island where most of these typhoons strike land, causing wave surges of up to 2.5 m above normal sea level. Similar storms originating in the Bay of Bengal move northwards and commonly strike the coast of Bangladesh as cyclones. A key consequence of these monsoonal weather phenomena is that the most vulnerable coastal regions of Vietnam and Bangladesh are protected by man-made earthen sea dikes that have transformed the coastal ecology of the upper intertidal zone; these coastal defenses have also served to reclaim land for agriculture, salt-making, and aquaculture.

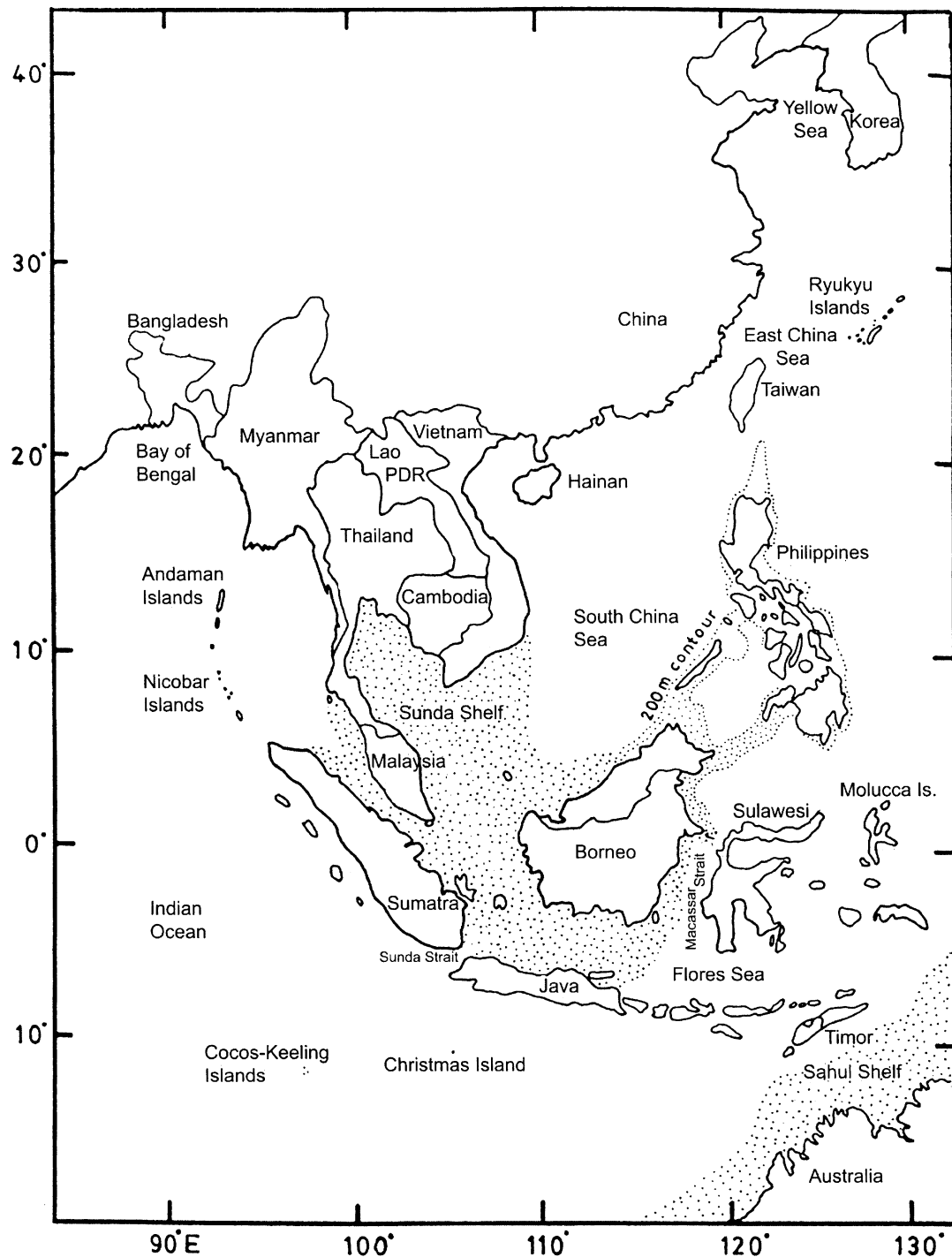
Further north, the coastlines of China and the Korean Peninsula are influenced by the East China Sea and the Yellow Sea. The latter is a productive, semi-enclosed water body with an average depth of only 44 m and a maximum of 100 m. The Yellow Sea is influenced strongly by several major rivers, especially the Huang He (Yellow River) and Yangtze River, which discharge more than 1.6 billion tonnes of sediments annually. Sedimentation and salinity, as well as seasonal temperature changes, control the ecology and productivity of the Yellow Sea ecosystem. The coastline of China consists mainly of mudflats and salt marshes. There are also large areas of intertidal mudflats and salt marshes along the west coast of the Korean Peninsula that provide an important ecological support function to migratory birds (including the very rare black-faced spoonbill *Platylea minor*: world population in 1998 only 613), as well as fisheries stocks (Kellerman and Koh 1999).

Coastal Habitats of Eastern Asia

This section describes the ecology of the main coastal habitats found in Eastern Asia, particularly the mangrove forests, seagrass beds, and coral reefs of Southeast Asia which are the most productive and diverse in the world. However, Eastern Asia is a region where the natural ecology of the coast shows a complete spectrum of impacts due to human activities, from near pristine beaches and coral reefs on remote islands, to massive coastal land conversion for agriculture, aquaculture, and urban/industrial uses. The pressures from human population and development, particularly in heavily populated countries like Bangladesh, China, Indonesia, and Vietnam, have transformed or degraded the coastal zone at an alarming rate over the past 20 years.

Mangroves

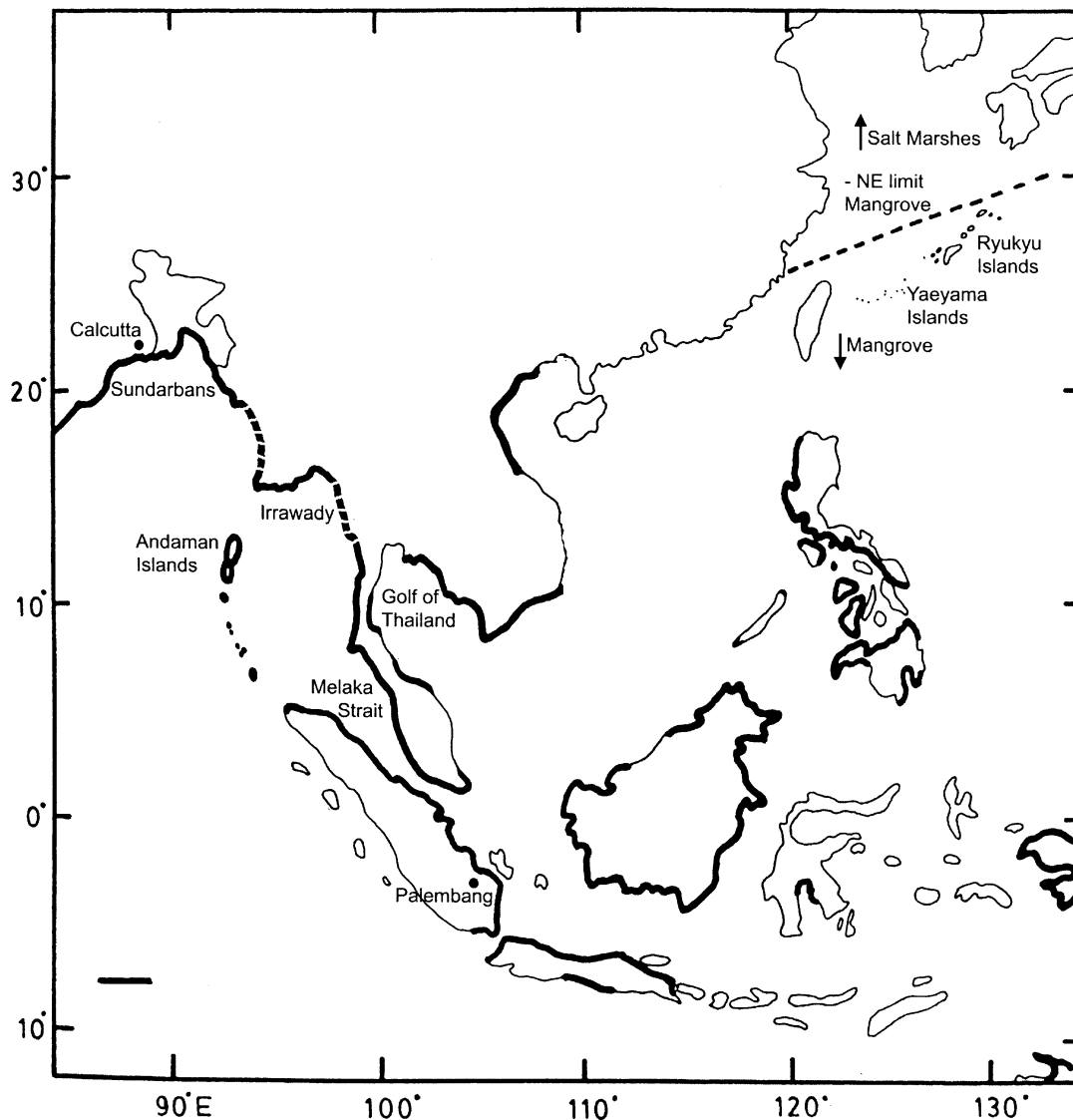
Mangroves, or mangrove swamp forests, consist of trees and shrubs growing in the upper intertidal zone of muddy tropical and subtropical shores, including estuaries and lagoons. Mangroves develop best along sheltered shores where there is high and prolonged rainfall and an average temperature exceeding 20 °C (Macnae 1968). Mangrove forests provide important habitat and other ecological support functions for many terrestrial and aquatic animals, including both resident and visiting species, especially birds, as well as fish and shellfish populations that support substantial coastal fisheries (Macintosh 1982). For this reason, it is usual today to regard mangroves as an ecosystem rather than simply a community of highly adapted plants and animals. The biology and ecology of mangroves has been reviewed in detail, notably by Macnae (1968), Tomlinson (1986), Robertson and Alongi (1992), and Hogarth (1998), while their traditional exploitation for forestry, fisheries, and aquaculture production is also well documented. An early account by Watson (1928) and a



Asia, Eastern, Coastal Ecology, Fig. 1 General map of Eastern Asia; the Sunda Shelf and Sahul Shelf regions are shaded

recent update by Gan (1995) describe the mangroves of Matang in Perak, Malaysia, which have been under management for wood production since the 1900s (mainly for charcoal). Matang is widely regarded as the best-managed mangrove forest ecosystem in the world. Schuster (1952) gives an excellent account of the partial conversion of mangroves in Java to integrated mangrove fishponds, or tambaks,

for milkfish culture, with marine shrimps (Penaeidae) and mud crabs (*Scylla* spp.) being the secondary crop. The current tambak management system in Indonesia is described by Sukardjo (2000). While these represent good examples of the sustainable economic use of mangroves in Eastern Asia, it is really only within the past decade or so that all the functions, attributes, and values of mangroves (and other



Asia, Eastern, Coastal Ecology, Fig. 2 Distribution of mangrove swamp forest in Eastern Asia (heavy coastlines)

tropical coastal ecosystems) have become widely recognized and understood (e.g., White and Cruz-Trinidad 1998).

According to the World Mangrove Atlas, 41.5% (75,173 sq. km) of the world's total mangrove area (181,077 sq. km) are found in South and Southeast Asia. Indonesia (42,550 sq. km of mangroves) dominates with 23% of the world resource (Spalding et al. 1997). Mangroves occur throughout the tropical and subtropical regions of the world, but the Indo-Pacific region is clearly their center of diversity, with 63 out of the 90 or so genera of mangrove and associated plants occurring here. Mangrove forests extend into the colder temperate subregion of Eastern Asia as far as southern China (Hainan Island and Fujian Province), Taiwan, and the Ryukyu Islands of Japan (Fig. 2). The Ryukyus (26–27°N) are the northern limit for mangroves; here

temperatures in winter fall to 17–22 °C. At these northern extremes, the mangrove vegetation is much more stunted and limited in species compared to the magnificent mangrove forests of tropical Asia. The drop-off in mangrove species with latitude is shown clearly in China's Fujian Province. In southern Fujian, there are six mangrove species (*Aegiceras corniculatum*, *Avicennia marina*, *Acanthus ilicifolius*, *Bruguiera gymnorhiza*, *Excoecaria agallocha*, and *Kandelia candel*); this falls to four species in central Fujian and to only one species (*K. candel*) in northern Fujian (Peng and Wei 1983). Further north along the coastline of China, salt marsh vegetation, especially *Salicornia*, replaces mangrove as the dominant intertidal wetland community.

It is in Southeast Asia where mangroves have achieved their greatest development and diversity, especially along the

Asia, Eastern, Coastal Ecology,**Fig. 3** Tall, stilt-rooted

Rhizophora trees dominating a typical estuarine mangrove forest in Southeast Asian (Ranong, southern Thailand)



sheltered shores bordering the shallow seas of the Sunda Shelf within the Malay–Indonesia subregion (Fig. 3). This includes the coasts of Peninsular Malaysia and Sumatra adjoining the Melaka (Malacca) Straits, the Gulf of Thailand and the coast of Cambodia, the Mekong Delta of southern Vietnam, the northern coast of Java, and almost the entire coastline of Borneo (see Fig. 2). The prolific reproductive capacity of the main forest-forming Asian mangrove species is one reason for their great success throughout Southeast Asia (Fig. 4).

Under the more exposed conditions of the Bay of Bengal and South China Sea, it is in the great deltas of the Ganges–Brahmaputra (Bangladesh), Irrawaddy (Myanmar), and Red River (northern Vietnam) where mangroves are most extensive and luxuriant. In each case, massive deposition of alluvial sediments from the river systems creates mud banks along the shallow coastal zone that enable mangrove forests to colonize and flourish. At the same time, the presence of mangroves affects the hydrological regime by modifying tidal flows and velocities, influencing sedimentation and erosion patterns, and reducing the impact of waves along the coast (Wolanski and Ridd 1986; Kjerfve 1990). Thus, mangrove systems are highly dynamic, both in the short-term and over longer time periods. The longer-term dynamic changes in coastal land formation resulting from these processes are well illustrated historically, especially in the Sunda Shelf subregion. Palembang was a thriving port city when Marco Polo visited Sumatra in 1292, but since then coastal accretion has left Palembang more than 50 km inland (Macnae 1968).

The rates of “land-building” by mangroves are lower on more exposed coastlines due to less favorable current and/or climatic conditions. Mangroves rapidly colonize the mudbanks formed in the Irrawaddy Delta, but some sediments are carried



Asia, Eastern, Coastal Ecology, Fig. 4 Spear-like propagules of *Rhizophora mucronata* adapted for penetrating soft muddy sediments. Propagules give mangrove trees of the Family Rhizophoraceae, great capacity for self-generation and colonization

away eastwards and offshore by coastal currents generated during the monsoons. In the Mekong Delta, there is a net flow of sediments from east to west, and consequently some

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Fig. 5 A dense monoculture of planted *R. apiculata* trees in Ca Mau, southern Vietnam.

Mangrove forestry on the central platform is integrated with aquaculture in the surrounding water canals (foreground)

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Fig. 6 A dense stand of the palm *N. fruticans*, at the transition zone from mangroves to freshwater vegetation, Mekong Delta, Vietnam



parts of the eastern shore of the Cam Mau Peninsula are actually eroding, despite the great sediment loads being deposited by the Mekong River system. Moreover, Typhoon Linda, which struck southern Vietnam in November 1997, destroyed the last vestiges of the original mixed forest, as well as damaging many of the older mangrove forest plantations in Ca Mau. As in China, the colder climate in northern Vietnam, coupled with frequent typhoons, limit mangroves in the Red River Delta to low, shrubby vegetation (dominated by *Avicennia* spp. and *K. candel*). In contrast, the mangrove forests of southern Vietnam feature taller trees of *Rhizophora Sonneratia*, *Bruguiera*,

and *Ceriops* spp., including large tracts of plantation mangrove consisting of a single commercial species, *Rhizophora apiculata* integrated with canals for aquaculture (Fig. 5).

Nipa Swamp Forest

Along tropical riverbanks, and in low-lying muddy swamp-lands subjected to greater freshwater inundation, the nipa palm (*Nypa fruticans*) replaces mangrove forest as the characteristic plant community (Fig. 6). Nipa palms often intergrade with mangroves in estuarine transition zones near the upper limit of saltwater penetration. Large areas of nipa

swamp have been destroyed by land reclamation projects and for shrimp farming, but nipa remains an important wetland community in many Southeast Asian countries, both ecologically and economically. Nipa is still used widely in Southeast Asia as a traditional thatching material for house-building; it can also be processed for sugar and alcohol production (Chan and Salleh 1987).

Salt Marshes and Mudflats

Salt marshes replace mangroves in the coastal ecology of the temperate subregions of Eastern Asia, especially along the sheltered western coastline of the Korean Peninsula. The southern portion of the Yellow Sea, including the entire west coast of Korea, is dominated by tidal flats up to 10 km wide, with an operating tidal range of 4–10 m. This forms the Yellow Sea ecoregion that includes an estimated 2850 km² of tidal flats (Kellerman and Koh 1999). Salt marsh vegetation characterized by *Suaeda japonica* once flourished on the upper part of the shore, but much of this has been lost through land conversion for agriculture.

The tidal flats of the Yellow Sea ecoregion contain muddy, sandy, and mixed sediments; these areas are highly productive, with a rich benthic fauna of polychaetes, bivalve and gastropod mollusks, crustaceans, holothurians, and branchiopods (Koh and Shin 1988). The tidal flats are an extremely important habitat for migratory wading birds on their passage between Indonesia–Australia and Eastern Siberia–Alaska. Ecologically, the coastal habitats of the Yellow Sea have been likened to those in northern Europe, particularly the Wadden Sea (Kellerman and Koh 1999).

Salt marsh plants also replace mangroves locally in more arid conditions, even within tropical Southeast Asia. On dry muddy soils above the mangrove zone, or where coastal dikes and embankments have been constructed, the salt-tolerant creeper *Sesuvium portulacastrum* often covers such exposed habitats, its fleshy leaves being capable of retaining water even during prolonged droughts. In Vietnam, where virtually all the mangrove zone has been disturbed by sea dike and shrimp pond construction, several other marginal species colonize the supralittoral zone, including the shrubs *Clerodendron inerme* and *Pluchea indica* and the tree *Thespesia populnea*, which produces a hibiscus-like flower. Such high, arid habitats are hostile to most marine animals, but fiddler crabs (*Uca* spp., Family Ocypodidae) and mangrove sesamid crabs (Family Grapsidae) burrow deeply into the soil for protection. A few mangrove species, especially *E. agallocha* and *Lumnitzera racemosa* also tolerate the dry, acidic conditions on the slopes of coastal dikes. Local farmers in Vietnam often plant these mangrove trees, along with *Thespesia* and *Casuarina*, to provide shade and wind protection for their houses and farmland. Very poor people also gather *Sesuvium* for use as animal fodder.

Sandy Beaches

Eastern Asia includes some of the most magnificent sandy beaches in the world, many of which are still unspoiled, particularly those fringing remote coral islands. The world's longest beach is located on the Arakan coast (now Rhakine) of Myanmar, while the powdery white beaches of Boracay Island (Western Visayas, the Philippines) are regarded by many to be the world's finest. Beautiful sandy beaches also make up 90% of the eastern coastline of Peninsular Malaysia. Not surprisingly, the sandy beaches of Indonesia, Malaysia, Philippines, Thailand, and more recently Vietnam, are promoted strongly as a key attraction to tourists. Particularly renowned are the beaches on many tropical islands such as Bali and Lombok (Indonesia), Penang and the Lankawi Islands (Malaysia), Phuket and Koh Samui (Thailand), Boracay (Philippines), and Phu Quoc (Vietnam).

Beach sediments of volcanic origin occur in parts of Indonesia, such as eastern Java, creating beaches of black or gray sand. Beach ridges and sand dunes are not well developed in the humid tropical regions of Asia, but examples occur in parts of southern Java and southwestern Sumatra; here a natural beach ridge and dune vegetation of *Casuarina*, *Pandanus Calophyllum*, *Inophyllum*, and *Barringtonia* occurs, together with planted or self-established coconut palms. A similar sandy ridge and sand dune formation occurs much more extensively in monsoonal Asia, especially along parts of the coastline of Myanmar and Vietnam. Natural and planted *Casuarina equisetifolia* trees are a feature of the low sand ridges behind the beach slope. In many provinces of Vietnam planted *Casuarina* trees act as a wind break in front of coastal villages, as well as providing a traditional, and renewable, source of wood (Fig. 7). The slender form of the branches and leaves of this native conifer enables *Casuarina* to withstand the arid conditions and strong winds associated with sea-facing sandy beaches.

Within the upper intertidal zone on exposed shores with strong surf, the sand is too dry and unstable for most marine animals, but burrows of the ghost crab *Ocypode ceratophthalma* are a common site on the upper beach slope. This large ocypodid crab burrows deeply for protection from the sun and desiccation. Ghost crabs emerge at low tide to scavenge from the sand line, or to chase and capture smaller fiddler crabs (*Uca* spp.) or soldier crabs (*Dotilla* spp.) which occupy the lower, less-exposed intertidal zone. As the fastest running land crustacean, *Ocypode* is particularly well adapted to catch its prey and to escape predators.

Of the seven species of sea turtles found in the world, four species visit sandy beaches in Malaysia to lay their eggs. The beaches of Rantau Abang, in Trengganu on the east coast of Peninsular Malaysia, are one of only four main nesting sites worldwide known for the giant leatherhead turtle, *Dermochelys coriacea*, and the only nesting site in Eastern Asia. This rare

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Fig. 7 Planted *C. equisetifolia*
 trees serving to stabilize the upper
 beach zone and provide a
 windbreak, Nam Dinh Province,
 northern Vietnam



and endangered species is the world's largest living reptile, capable of reaching 2 m in length with a weight of more than 900 kg. It is also the only sea turtle without a hard shell; instead it has a leathery, ridged carapace. Although this unique species has a worldwide distribution, being found even in temperate seas, it only breeds in the tropics. After laying many large parchment-like eggs in a nest dug carefully in the sand at the top of the beach slope, the female turtle returns to the sea. As soon as the eggs hatch the baby turtles struggle out of the sand and crawl down the beach slope to enter the sea. Despite efforts to protect its nesting beaches in Malaysia, the giant leathery turtle has declined significantly over the past 30 years, disturbance to its nesting habitat being one of several critical factors.

The other turtles nesting on Malaysia's beaches are also endangered, having suffered severe decline due to egg harvesting and relentless fishing pressure because of their high value. It is also likely that their feeding habitats have declined significantly. The olive ridley turtle (*Lepidochelys olivacea*) is the world's smallest marine turtle; its diet includes crustaceans, mollusks, and jellyfish. The green turtle (*Chelonia mydas*) has a vegetarian diet of seagrasses and algae, though it may also consume sponges. The beautiful hawksbill turtle (*Eretmochelys imbricata*) also feeds on sponges. The hawksbill and ridley turtles are hunted for their shells, while the green turtle is prized for its meat and oil. The same four species of marine turtle occur in the Philippines (Gomez 1980), but they have declined greatly in numbers, just as in Malaysia, despite conservation measures including closed seasons for turtle fishing and egg-collecting (Palma 1993). Sea turtles are particularly difficult to conserve

because of their precise habitat requirements, as well as their high value to poachers. The disturbances to their historical breeding beaches, and the great decline in coral reefs and seagrass beds which provide sea turtles with feeding habitats, place these remarkable marine reptiles at great risk, even if hunting them can be controlled.

Coral Reefs

There are almost 100,000 sq. km of coral reefs in Southeast Asia, representing about one-third of the world's total coral resource. Like mangroves, corals show their greatest diversity in Southeast Asia, with 600 of the almost 800 reef-building coral species found in this tropical subregion of Eastern Asia. The center of coral diversity is eastern Indonesia, the Philippines, and the Spratly Islands with over 70 genera of hard coral represented. Indonesia and the Philippines, alone contain 77% of the region's coral reefs due to their multitude of islands featuring the sheltered, clear water conditions corals require. Coral reef fishes, crustaceans, and mollusks also show their greatest diversity in this subregion and up to 3000 animal species may be present on a single coral reef. A recent identification guide to corals of the world (Veron and Stafford-Smith 2000) provides an excellent reference to the corals of Eastern Asia. Several new coral species are described and there are probably many more species still to be identified.

Notwithstanding their great beauty, diversity, and ecological importance, the world's coral reefs have suffered a dramatic decline in recent years. About 10% may already have been degraded beyond recovery; a further 30% are likely to decline seriously within the next 20 years. Recently, the World Resources Institute estimated the potential threat to

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Fig. 8 A coral reef flat of hard and soft corals in Singapore; heavy coastal industries can be seen in the distance



coral reefs using standard criteria to calculate a “reefs at risk” indicator (Bryant et al. 1998). The reefs identified as being at greatest risk are in South and Southeast Asia, East Africa, and the Caribbean. Specifically, over 85% of the reefs of Indonesia and Malaysia, and over 90% of those in Cambodia, Singapore, the Philippines, Taiwan, and Vietnam are threatened (Fig. 8).

Seagrass Beds

Seagrasses are distributed widely through tropical and subtropical Eastern Asia, mainly as low intertidal to subtidal beds in shallow sandy bays, lagoons, and around near shore islands. They often form an associated wetland community below mangrove-fringed shores with sandy sediments, or occur interspersed with coral reefs. Like corals, seagrasses do not tolerate the muddy or turbid habitat conditions typical of estuaries and mudflats. In Vietnam, for example, where more than 5000 ha of seagrass beds containing 15 species have been identified, the main sites for seagrasses are north of the Red River Delta in Quang Ninh Province, in the central region of Vietnam (Hue and Khanh Hoa-Nha Trang) where there are many magnificent sandy bays and lagoons, and around the southern island of Phu Quoc (Vietnam Environmental Monitor 2002). Seagrass beds in Vietnam have become heavily degraded due to excessive harvesting for animal food and organic fertilizer, as well as destructive fishing practices (use of dynamite and cyanide) in seagrass and coral habitats.

Coastal reclamation projects have also caused widespread loss of Eastern Asia’s seagrass beds and coral reefs, but there is also growing evidence of decline due to natural causes,

including coral bleaching associated with warming of the seas and “wasting disease” in seagrasses. Haynes et al. (1998) concluded that most seagrass losses, both natural and anthropogenic, stem from reduced light intensity due to sedimentation and/or epiphytic development caused by water nutrient enhancement.

Coastal development activities commonly increase sediment loading in the surrounding waters, causing great harm to seagrass beds and coral reefs. One well-documented example is Hong Kong’s Chep Lap Kok Airport (Lee 1997). Increased sediment levels caused by the airport’s construction virtually wiped out the nearby beds of *Zostera japonica* and *Halophila ovata*. While *Halophila* has recovered well following the construction period, the *Zostera* beds have almost completely disappeared, although some transplanted patches have survived and enlarged. This difference between seagrass species may be explained by the observation that *Zostera* is mainly subtidal and intolerant of shading, whereas *Halophila* also occurs in the lower intertidal zone and can survive in relatively low light intensity.

Seagrass beds play various ecological roles, including sediment adsorption and nutrient absorption from seawater; this may serve an important protection function for adjacent coral reefs, since corals are even more sensitive to sediment and nutrient loading. Seagrasses also provide important nursery habitats for many fish species, including groupers fingerlings, for example, which are heavily collected throughout Southeast Asia for aquaculture. Seagrass beds are also critical feeding areas for the green turtle (*C. mydas*) and the rare sea mammal, the dugong (*Dugong dugon*), which feeds almost exclusively and selectively on various seagrasses, especially

Halodule and *Halophila*. Dugongs may choose these particular grasses in preference to other genera (*Thalassia* and *Enhalus*) because they are easily digested (*Halophila*) and rich in available nitrogen (*Halodule*). Dugongs are capable of digesting a large volume (8–15% of body weight per day) of low quality plant material (Reynolds and Odell 1991), an observation that explains why the great loss of seagrass beds in Eastern Asia has been so damaging to the survival prospects for this unique mammal. It has also been reported that grazing by dugongs and turtles plays an ecological role in seagrass communities by stimulating the growth of new leaves, thereby changing the structure and nutritional status of the seagrass bed (Aragones and Marsh 2000).

Exploitation of Coastal Resources

Human Pressure

Eastern Asia is a region where the great natural diversity of coastal topography, biological communities, and species has been transformed in many countries by immense human population and development pressures. Although human occupation and economic use of the coastal zone have a long tradition in the region, dating back centuries in the case of coastal farming and aquaculture, the pace of coastal development has accelerated since the 1980s, raising the need for habitat conservation and integrated coastal zone management as never before.

The effects of high population growth in the coastal zone, unsustainable exploitation of coastal resources, and conversion of coastal habitat for agriculture, aquaculture, and urban and industrial development, have now reached crisis levels. In reviewing the threatened status of wetlands in Asia in 1989, Wetlands International (an international NGO) noted that Southern and Eastern Asia (including the Indian subcontinent) had an average population density about eight times that of the rest of the world and was increasing at a rate of 55 million people annually. This led Wetlands International to conclude that, “Most of the threats to wetlands in Asia are a direct consequence of the need to feed and house this massive and ever-increasing number of human beings.” Unfortunately, this is a very accurate assessment of the problems facing Eastern Asia’s coastal ecosystems in general.

A good historical example is the Chakaria Sundarbans region of Bangladesh. Located on the eastern border of the Bay of Bengal, this is the eastern limit of the great Sunderbans mangrove forest shared between Bangladesh and India and the largest mangrove ecosystem in the world. One hundred years ago the Sundarbans was reserved forest, but over the next several decades the mangroves were exploited for wood and fishery products and then encroachment followed for human settlement and salt production. Mangrove conversion accelerated greatly during the 1980s due to the development

Asia, Eastern, Coastal Ecology, Table 1 Estimated loss of the original mangrove forest area in different countries. (Based on data from World Resources Institute 1996)

Country	Loss of mangrove forest (% loss by area)
Bangladesh	73
Brunei	17
Indonesia	45
Malaysia	32
Myanmar	58
Singapore	76
Thailand	87
Vietnam	62
Unweighted average for Asia	61

of coastal shrimp farming, until by 2001 all the remaining mangrove had been converted to shrimp ponds and human settlements (Hossain and Kwei-Lin 2001). Even by the 1980s, the Chakaria Sundarbans were considered too degraded to merit any special conservation efforts.

There has been a similar history of mangrove exploitation in Vietnam, where now less than 30% of the original mangrove area remains and virtually no pristine forest has survived. Only 60 years ago, mangroves in Vietnam covered an area of up to 400,000 ha (Maraund, 1943 cited by Hong and San 1993), of which about 250,000 ha flourished in the Mekong Delta. The greatest concentration of mangrove (150,000 ha) was in the Minh Hai Peninsula at the southern tip of Vietnam (now divided into Ca Mau and Bac Lieu provinces). The most recent estimate for the total area of mangrove forest remaining in Vietnam is only 110,000 ha (Vietnam Environmental Monitor 2002). Overall, the Eastern Asia region has lost more than 60% of its mangrove forest in just a few decades (Table 1).

It is not just Eastern Asia’s fragile tropical mangrove forests, seagrass beds, and coral reefs that have become severely depleted or degraded. Approximately 600 million people live in the area around the Yellow Sea, including major cities such as Shanghai, Seoul-Inchon, and Pyongyang-Nampo. The Yellow Sea is among the most heavily exploited coastal and marine areas of the world, with more than 100 species of fish, crustaceans, and cephalopods captured by its fisheries. Today, coastal habitats and fishery resources in the Yellow Sea are threatened by both land-based and sea-based infrastructures, by pollution, and by overfishing and other forms of unsustainable resource exploitation. More than 100 million tonnes of domestic sewage and about 530 million tonnes of industrial wastewater are discharged annually into the nearshore areas of the Yellow Sea. These discharges come from large coastal cities, shipping, and oil exploration; they contain many pollutants, including heavy metals (cadmium, lead, mercury), oil residues, and nitrogenous compounds.

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Fig. 9 Recreational use of a sandy beach in Johore, Peninsular Malaysia, with dense surrounding urban infrastructure – a scene typical of many beaches near large cities throughout Eastern Asia



Other economic uses of the coastal zone in Eastern Asia tend to reflect the stage of development of the countries concerned. Like Bangladesh and Vietnam, the more developed South-east Asian countries of Indonesia, Malaysia, Philippines, and Thailand have heavily exploited sheltered bays, lagoons, estuaries, and other mangrove and mud flat areas for coastal agriculture and aquaculture. There has also been localized industrial and urban development, as well as heavy investment in tourist infrastructure centered on islands and mainland sandy beach sites near major cities; for example, Pattaya and Hua Hin in Thailand, Port Dickson and Johore in Peninsular Malaysia (Fig. 9). Estimates for Thailand made by the Royal Thai Forestry Department in 1993 revealed that, of the original 372,448 ha of mangrove forest (measured in 1961), 133,812 ha (35.9%) had been converted into salt pans, mining, agriculture, and infrastructure, 64,991 ha (17.5%) was occupied by coastal shrimp farms, and 4961 ha (1.35%) had been converted to other uses (leaving only 45% undeveloped).

In the most populated and developed parts of Asia, especially Hong Kong, Singapore, and parts of Peninsular Malaysia, China, and Korea, coastal land is at such a premium that urban and industrial development (e.g., airports, container ports, oil-refineries) have almost completely replaced earlier natural resources use. Such economic developments have overtaken the coastal zone in a remarkably short period of time. Only 50 years ago Schuster (1952) noted “The greater part of down town Singapore is built on *Rhizophora* piles driven into the mud of the harbor region.” Today only carefully conserved vestiges of the original coastal habitat remain in Singapore (e.g., Fig. 8).

Coastal Aquaculture

Of the many human uses of the coastal zone, aquaculture stands out not only in terms of its scale and importance throughout the Eastern Asia region, but also because of its interaction with the ecology of the coastal environment (e.g., Primavera 1993). Unlike the conversion of coastal habitat to agriculture or urban and industrial uses, the production of human food by aquaculture is still critically dependent on the carrying capacity of the surrounding aquatic ecosystem. Where coastal aquaculture causes environmental degradation, there is a rapid negative feedback on the viability of the aquaculture production system itself, as the experiences with intensive shrimp farming in Eastern Asia show only too well (e.g., Smith 1999). The term “self-pollution” is sometimes used to describe this problem in coastal aquaculture (Phillips and Macintosh 1997).

Shrimp Farming

Shrimp farming has had a dramatic impact on the coastal ecology of Eastern Asia over the past 20 years or so. Although coastal aquaculture has a long history of operation in South-east Asia (Schuster 1952; Macintosh 1982), it was the commercialization of technology for breeding one particular species, *Penaeus monodon*, or the black tiger shrimp in Taiwan, which enabled shrimp seed (post-larvae) to be produced, then reared intensively in coastal shrimp farms. The world demand for farmed shrimp also rose significantly in the 1980s, led at that time by Japan, because the yields from shrimp capture fisheries had long since reached a ceiling level. From early pioneering research on the breeding

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Fig. 10 Intense use of the coastal zone for aquaculture; the photograph shows ponds and a complex of water pipes supporting fish and shellfish farming during the aquaculture boom in southern Taiwan in the 1990s



of *Marsupenaeus japonicus* in Japan, research then developed in China involving the local species *Fenneropenaeus chinensis* and in Taiwan with *P. monodon*. Cool climatic conditions and high costs had mitigated against shrimp farming in Japan involving the local species *M. japonicus*, but the black tiger shrimp was readily accepted by the Japanese market and Taiwanese shrimp producers rapidly spread their knowledge to other parts of tropical Asia where conditions for farming *P. monodon* were much better. During the 1980s, investors in Thailand, Philippines, Indonesia, and Malaysia rushed to adopt the new technology, converting mangrove forests, traditional shrimp ponds, salt pans, coastal rice fields, and (in the Philippines) sugar plantations, into intensive shrimp farms. From Southeast Asia, “shrimp fever” as it became known (Primavera 1993) spread to Vietnam and Bangladesh, countries with huge areas of coastal mangroves and mudflats considered more suitable for conversion to extensive, low-cost, shrimp ponds. Since 1995, records show that on average 138 million shrimp juveniles, or “seed” (mainly of *P. monodon* and *Macrobrachium rosenbergii*) have been collected annually from the Sundarbans reserve forest in Bangladesh to support coastal shrimp farming. Shrimp seed collecting is extremely destructive as many other species of fish and crustacean seed are caught and discarded by the shrimp fishers, but it is a form of livelihood for the great majority of families living in the Bangladesh Sundarbans and involve an estimated 225,000 shrimp seed collectors (Rouf and Jensen 2001).

Shrimp farming was also promoted on a massive scale in China, with shrimp farms occupying 160,000 ha from Hainan

Island in the south to Liaoning Province in the northern temperate zone. Production reached a peak of 200,000 t in 1992 before coastal water pollution and shrimp diseases had a devastating impact on output (Smith 1999). Shrimp farming has also developed in Korea, but as with its neighbors China and Taiwan, aquaculture has been badly affected by coastal industrialization and associated problems of water pollution (including self-pollution) and shrimp disease outbreaks (Fig. 10). The shrimp farms in Korea are confined almost entirely to the west coast where it is more sheltered. The majority of farms are located within the intertidal zone or near to the coastline. Reflecting the colder operating conditions in Korea compared to Southeast Asia, the main species cultured are the local shrimp *F. chinensis* followed by *M. japonicus*. Annual production levels are rather low (less than 1 t/ha), but the environmental limitations on shrimp production in Korea are offset by the high price farmers can obtain for live shrimp delivered to the restaurant trade, including live sales to Japan.

Mud Crab Culture

Poor coastal communities from Bangladesh to the Philippines are still heavily dependent on mangrove-based fisheries and aquaculture to support their livelihoods (Fig. 11). This includes crab fishing for *Portunus* (blue swimming crab) and *Scylla* (mud crab), as well as shrimp fisheries and shrimp culture. Mud crab culture has also developed rapidly as a secondary activity to shrimp farming within the mangrove forest ecosystem. It carries much lower investment costs and risks than shrimp farming and is therefore more suitable for



Asia, Eastern, Coastal Ecology, Fig. 11 A large mud crab (*Scylla paramamosain*) for sale in a local market in Vietnam. Mud crabs support important artisanal fisheries and coastal aquaculture production throughout Eastern Asia

poor families (Overton and Macintosh 1997). In the Lower Mekong Delta of Vietnam, farmers can now rear from 500 to 800 kg of mud crab per hectare in a simple mangrove pond (unpublished data). With a farm gate value of about USD3/kg, even a small crab pond can provide a poor Vietnamese family with an annual net income of USD200–300. This is a significant contribution to the economy of families living in remote coastal areas where the potential daily income from fishing, salt production, or agricultural laboring is only USD1–2.

Other Forms of Coastal Aquaculture

Many other types of aquaculture are also economically and ecologically important in the coastal zone of Eastern Asia. Economically, caged fish culture for valuable species such as groupers (*Epinephelus* spp.) and snappers (*Lutjanus* spp.) appears to be very attractive, but it carries many risks, especially from fish diseases and can be highly polluting due to the use of trash fish as the main type of feed. The environmental damage stemming from large-scale collection of wild fingerlings (especially groupers) from coral reef and seagrass habitats to support aquaculture is another great cause of concern. For large-scale aquaculture production, and in terms of their ecological impacts, the culture of bivalve mollusks and seaweeds has many advantages over most fish and crustacean species. Mollusks and large seaweeds are cultured in huge quantities by both China and the Republic of Korea. For example, marine aquaculture production in China reached

Asia, Eastern, Coastal Ecology, Fig. 12 The ark shell, or “blood cockle” (genus *Anadara*), one of the many bivalve mollusks found abundantly on coastal mudflats throughout Eastern Asia and important in aquaculture, especially in Malaysia and China



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Fig. 13 Tourists, long-tailed macaques, and mangrove forest in the Can Gio Biosphere Reserve, near Ho Chi Minh City, Vietnam. The reserve has conservation, sustainable development, education, and research functions



10.6 million tonnes in 2000, with 93% of this total contributed by bivalve mollusks and seaweeds. The main species of mollusks include oysters (*Crassostrea*), mussels (*Mytilus*), hard clams (*Meretrix meretrix*), short-necked clams (*Macra* spp.), scallops (*Clamys farreri*), and ark shells of the genus *Anadara* (Fig. 12). Gastropod mollusks of the genus *Haliotis*, better known as abalone, are more difficult to rear, but are in high demand as a delicacy. These snail-like animals occur naturally on subtidal rocks (especially in Korean waters) where they graze on algae. However, overfishing has decimated abalone populations because they are so valuable and easy for divers to harvest.

Seaweeds grow naturally on rocks in the lower intertidal and subtidal zones, especially in the warmer southerly regions of the Yellow Sea. The seaweed *Pelvetia siliquosa* occurs widely in the Shandong and Liadong regions of China, as well as around the Korean Peninsula. It grows most luxuriantly in Korean waters where it has been harvested for hundreds of years and exported to China. *Sargassum pallidum*, another edible seaweed, dominates in the west Yellow Sea, and *Plocamium telfairiae* is also common there. A cold water seaweed, the kelp *Laminaria japonica* has become the most important seaweed cultured in China. Introduced from Japan, kelp farming now occupies more than 3000 ha of China's coastal waters (yielding 10,000 t dry weight/year) and is supported by hatcheries producing young plants for transfer to growing frames in the sea once the water temperature drops below 20 °C.

Coastal Habitat Conservation and Restoration

Given the already very high human population pressure in many Asian countries, it is difficult to be optimistic about the future for coastal habitats and the communities of plants and animals they support. Coral reefs and seagrass beds are particularly sensitive to destructive fishing methods, pollution, and any increased sedimentation caused by coastal construction or other development activities. Even mangrove forests, which are more robust and relatively easy to manage on a sustainable basis by tree replanting, have disappeared at an alarming rate (Table 1).

There are some positive indications that coastal protected areas (e.g., biosphere reserves, nature parks, and sanctuaries) can preserve habitat and protect species biodiversity. In addition, protected areas can play a valuable role in raising public awareness of the need for such conservation efforts (Fig. 13). Habitat restoration, especially of mangroves, is also being supported very actively in several Asian countries. Thousands of hectares of mangrove have already been planted successfully in Bangladesh and in the Red River Delta and Mekong Delta of Vietnam in order to protect the coastal zone from storms, flooding, and erosion, and to accelerate the reclamation of land from the sea. Mangrove planting also leads to a higher diversity of animals and birds, and to improved fisheries production. However, the full ecological impact of mangrove reforestation need careful assessment, both in terms of the plant and animal communities that develop in mangrove

plantations and the physical changes that take place in the ecosystem (Macintosh et al. 2002).

In contrast to mangroves, it is very difficult to restore coral reefs and seagrass beds and most efforts to date have been only experimental in scale, involving, for example, the transplanting of healthy coral fragments or individual coral heads onto a degraded reef (e.g., Clark and Edwards 1995; Rinkevich 2000). For corals and seagrasses, strict conservation of the coastal habitats supporting these delicate plant and animal communities, including protection from pollution, should be the first priority. This is not simply a plea to protect these diverse coastal communities for biological interest; there are sound economic reasons for doing so. White and Cruz-Trinidad (1998), for example, estimated that the sustainable annual benefit from 1 ha of coral reef in the Philippines is in the range USD31,900 to USD113,000, including its fisheries, tourism, coastal protection, and biodiversity values. This suggests that, conservatively, coral reefs are contributing almost USD1 billion annually to the Philippines' economy, but equally such huge economic benefits will be lost if coral habitat continues to be destroyed.

Cross-References

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Asia, Eastern, Coastal Geomorphology

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The shores of eastern Asia, extending from Thailand and Indonesia to eastern Siberia and including the Philippines, Taiwan, and Japan, largely follow the tectonically active zones along the eastern and northern sides of the Eurasian plate (Inman and Nordstrom 1971). Large stretches of coastal area along Indonesia, the Philippines, Taiwan, Sakhalin, Kamchatka, and the East Asian island chains are therefore tectonically unstable. In these areas the structural trends are generally parallel to the coast, which was distinguished by Suess (1892) as the Pacific type. Outside these areas, away from the tectonically active collision zones, the coastal regions are generally more stable and the structural trends are usually not parallel to the coast, which corresponds to Suess' Atlantic type. This type is present along most of the Asian mainland from Thailand to Siberia.

D. Eisma: deceased.

Comparatively straight coasts, situated along mountain chains, sometimes with river deltas and local alluvial foreland, are found mainly in western Sumatra, southern Java, northern Vietnam, and the Pacific coast of Siberia. A drowned older topography with an irregular coastline is present along parts of southern Vietnam, the mainland coast north of the Red River, on the islands of eastern Indonesia, on northern Kalimantan (Borneo), the Philippines, the Shandong Peninsula in China, Japan, southern Sakhalin, and the eastern end of Siberia. These coasts are usually somewhat altered by the sea, with bays containing beaches, spits, and barriers, sometimes being filled up with sediment and partially surrounded by alluvial foreland. The main exceptions are the large river deltas, the west coast of Kamchatka that consists largely of sand barriers, some beach barriers in Japan and on Sakhalin, and some areas like eastern Taiwan where tectonic uplift matches the Holocene sea-level rise. Elsewhere the coast is predominantly depositional, consisting of beaches, spits, barriers, tombolos, mudflats, marshes, mangrove swamps, and coral reefs.

The general direction of beaches, spits, and barriers is related to the direction of the swell. Between northern Japan and Indonesia the swell comes chiefly from northern directions, and the beaches and spits face largely E–NE. Where the swell is southerly, as in the Indian Ocean and in the Pacific north of 45°N, the beaches face mainly SE–SW. A major distinction can be made between the coasts north and south of about 40°N (Davies 1973). North of this latitude storm waves are predominant and pebbles form an important part of the beach material. During the winter, sea ice (pack-ice) borders the coast as far south as Vladivostok and northern Japan. Further north, along the northern shore of the Sea of Okhotsk and north of the Kamchatka peninsula, permafrost reaches the sea, and from the eastern Siberian coast ice cliffs have been reported. South of 40°N the ocean swell is more important, and large areas from southern Japan to Malaysia are affected by tropical storms. In this area, especially south of 25°N, sandy beaches are predominant while pebbles are relatively unimportant as beach material.

Coral reefs and beach rock are confined to the tropical and subtropical zones – they are not present north of 35°N. Mangrove grows south of 31°N (on the Asian mainland coast south of 26°N) and dominates large stretches of coast in Indonesia, Malaysia, and Thailand, especially in areas that are sheltered from the ocean swell. An important part of the East Asian coastline consists of several large river deltas (those of the Mekong, Yang-tse-Kiang or Chang Jiang, and the Hoang-Ho or Yellow river) and a large number of smaller deltas often of still appreciable dimensions such as the deltas of the Chao Phraya, the Red River, and the Si Kiang. This preponderance of deltas is due to the high mechanical erosion on the mainland between Thailand and Korea

(>60 ton per km² per year), which, associated with the presence of large and numerous rivers, has resulted in extensive deltaic and estuarine deposition.

Tidal effects are especially important along the mainland coast from Taiwan to Kamchatka, where the tides have an appreciable range. Usually the tidal range in eastern Asia is less than 4 m (and around Japan mostly less than 2 m) but in the northern Sea of Okhotsk, in the Yellow Sea, and along the Chinese coast it reaches more than 6 m. Tidal marshes have been formed locally, as, for example, along the Gulf of Bohai, in southern Korea, and in Hangzhou Bay.

A striking feature on many coasts from Japan through southeast Asia to Indonesia is the evidence of raised coastlines at 0.5–1 m, about 2 m, and 5–6 m above present sea level. These are indicated by flat, sandy terraces, beach ridges, coral reefs, mollusks and/or barnacles, notches, benches, platforms, and beachrock. Similar raised coastlines are known from many other coasts in Asia, Africa, South America, and Australia (for references, see Tjia 1975; Clark et al. 1978). Daly (1934) and many others therefore assumed that during the Holocene, about 6,000–4,000 years ago, sea level stood about 5 m higher than at present. Others have contended that these features are related to temporary high water during storms, or to local uplift. Tjia (1975) has suggested that glacioeustatic uplift of the areas in the Northern Hemisphere that were glaciated during the Pleistocene, may account for the absence of any indications for high sea levels during the Holocene. C14 dates of raised reefs and mollusk shells from the Pacific and Indian Ocean, Southeast Asia, and Australia point to a slight Holocene transgression as well as to local uplift. Clark et al. (1978) indicate that regional differences in the Holocene sea-level rise are related to differences in crustal rebound after the Pleistocene through ice cap melting and the increasing water load of the oceans.

The general features of the east Asian coasts are summarized in Fig. 1, which is based essentially on the world map given in Valentin (1952), but with emphasis on the coastline characteristics. More detailed knowledge, however, is still lacking for large stretches of coast, especially in eastern Siberia and Southeast Asia.

Siberia

The Siberian coast east of the Taymir Peninsula is flat up to the Kolyma River estuary and shows a drowned topography with only a higher coast at the Laptev Straits, where bedrock is exposed. The flat coast is locally raised a few meters and here small cliffs are formed by intensive thermal abrasion. This is caused by the regular melting of ice and frozen deposits (permafrost) along the cliffs and removal by waves of the loose sediment. The coastal flats are flooded over large distances (up to 30 km) during wind-induced surges. During

offshore winds the muddy nearshore seafloor can be exposed over similar distances. Several large rivers reach the sea in this area such as the Lena, Indigirka, and Kolyma rivers, but only the Lena has a vast delta, which is largely a relict feature. The other rivers have long, partly filled-in estuaries and insignificant deltas. The islands off the coast (the New Siberian Islands and Wrangel Island) have a bedrock nucleus surrounded by loose but frozen sedimentary deposits. Thermal erosion is large and can annually remove up to 50 m from the cliffs.

East of the Kolyma River mouth the coast is partially rock, where the mountain ranges between Kolyma and the Bering Strait come near to the sea. There is usually a narrow coastal plain, several kilometers wide, and at Cape Billing a sand barrier extends for about 50 km into the sea with an up to 10 km wide lagoon on the landward side. Transverse sandbars divide the lagoon into several rounded lakes; their formation is related to the development of wind-induced circulation cells in the lagoon. The sand barrier has been formed in an area that is protected against pack-ice (probably by Wrangel Island). In general, longshore sediment drift is limited by the short period that the coast is free of ice. The sediment is mainly derived from reworked glacial deposits. The easternmost part of Siberian coast along the Bering Strait is characterized by high cliffs and fjords where the Chukhotsk range meets the coast (Zenkovich 1985).

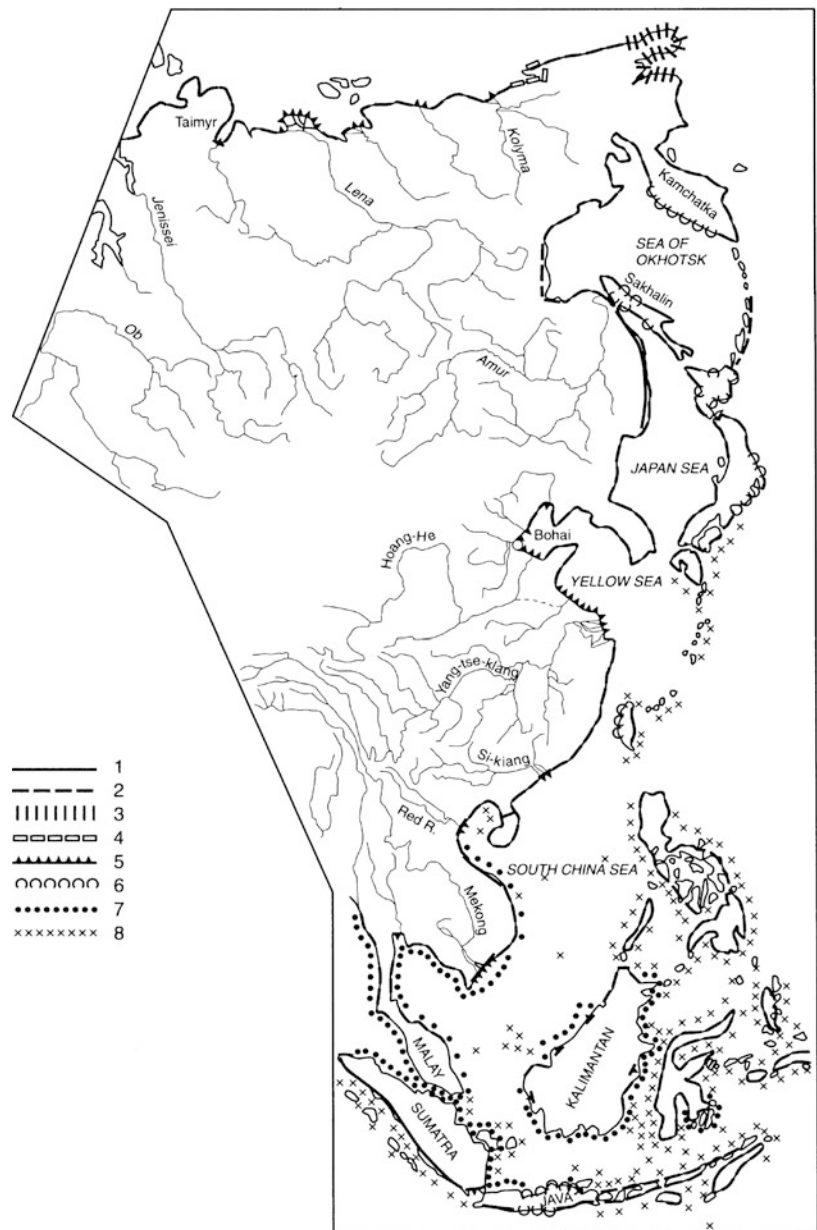
The Pacific coast of Siberia down to Korea extends from the arctic into the temperate zone and has complex tectonics with stable areas alternating with areas of emergence and subsidence. The coastal area is usually mountainous with rather short rivers (except the large Amur River) and narrow coastal plains. Waves and tides are variable: storm waves are up to 12 m in the Bering Strait and in the Sea of Okhotsk, while lower waves (normally up to 2–3 m, and up to 7 m during the winter) dominate in the other areas. Most of the coast is mesotidal, but locally, as in Penzinskaya Bay, becomes macrotidal (up to 12 m). There is a relatively large sediment supply to the coast, because of a wet, cold, or temperate climate in combination with easily weathered rock, such as volcanic deposits and Pleistocene fluvioglacial sediments. Barriers, coastal lagoons, and tombolos are common features, separated by promontories of rock or (unconsolidated) sediment (Kaplin 1985). The largely drowned topography has for a large part been filled in during the postglacial period. Along Sikhote Alin the coast is relatively straight and uninterrupted because of a low mountain range that lies parallel to the coast.

Japan, Sakhalin

The shores of Sakhalin and Japan consist of large stretches of ria coast, alternating with many small and several large stretches of

Asia, Eastern, Coastal Geomorphology, Fig. 1

Coastal morphology of eastern Asia. (1) Comparatively straight coasts along mountain chains, sometimes with river deltas and local alluvial foreland; (2) Drowned older topography with an irregular coastline, somewhat altered by the sea; (3) Fjords; (4) Tableland coast Table and coast with cliffs; (5) Large river deltas; (6) Predominantly spits, barriers, mud flats, marshes; (7) Mangrove swamp; (8) Coral reefs



beaches, spits, barriers, sand dunes, lagoons, and river outlets. Especially on northern Sakhalin, but also on Hokkaido and along Honshu, large parts of the coast are depositional alternating with rocky promontories. Locally, volcanism close to the coast has resulted in cliffs and beaches of volcanic material. Most of the coastal lowland in Japan is intensively used so that a large proportion of the coast is man-made: less than 50% has remained natural. Coastal terraces are widely distributed in northern Japan; in the south they occur mainly on the promontories along the Pacific coast. The terraces are predominantly present on the more easily abraded Tertiary and Quaternary formations, whereas the indented ria coasts have been developed mainly in folded Palaeozoic and Mesozoic rock. The highest terraces have been found in southwestern Hokkaido at

+585 m, but usually they reach to +150 m (Kosugi 1971). Below present sea level submerged flat surfaces have been found to depths of –150 and –190 m.

The terraces as well as raised coral reefs (Yabe and Sugiyama 1935) indicate that the Quaternary eustatic changes in sea level may have had an important effect on coastal development in Japan, but there is also much evidence for differential uplift, tilting, and local subsidence. Consequently, in an appreciable number of papers by Japanese authors the combined effects of both eustatic sea-level changes and crustal movements have been discussed (Richards and Fairbridge 1965; Yoshikawa et al. 1965; Richards 1970). The presence of ria coasts is generally seen as evidence of an overall subsidence since the

Pleistocene, but narrow terraces scattered along the rias suggest a more recent emergence. Some terraces (at +40 – +60 m) are widely distributed along the Japanese coast and are seen as the result of eustatic sea-level changes. Coastal terraces, formed at the same sea level, have also been found at different altitudes within short distances, even locally disappearing below present sea level, as a result of crustal movements. From this complex situation it has been deduced that the Holocene sea-level rise started around 18,000 BP and that between 6,000 and 3,000 BP sea level was 3–5 m higher than at present, which is indicated by former wave-cut notches, sea caves, benches, small terraces, marine deposits, and neolithic remains (Tada et al. 1952). Clark et al. (1978) have shown that these features can also be explained by crustal rebound after the Pleistocene and the increasing water load of the ocean.

A marked feature along the Japanese and Sakhalin coasts is the wide-spread occurrence of shore platforms (Takahashi 1974). These are most conspicuous along the coasts of the Pacific and the East China Sea, less on the Sea of Japan coasts, and least along the Seto Inland Sea. Their maximum width varies from 530 m along the Pacific to 80 m on the Sea of Japan coasts, which roughly corresponds to the wave conditions on each coast. The average width is 30–80 m and the maximum length about 4 km; they occur on all types of rock, but tend to be wider on Neogene mudstones, sandstone, and tuff-breccia.

The East Asia Mainland, Korea to Malaya

The mainland coast from Korea to the Malayan Peninsula is largely a drowned or ria-type coast, with abrasion at the headlands, accumulation (with sandy beaches, spits, and beach/ ridges) in the bays and, especially in China and southern Korea, numerous islands. Large estuaries and river deltas are present where major rivers reach the coast. In West and South Korea tidal flats have been formed in the bays and in the shallower parts between the islands (Guilcher 1976), which have been partly reclaimed. Coastal terraces are present along the East Korea coast up to 100–150 m and along the West Korea coast up to 30–60 m, which reflects the emergence of the east coast and the submergence of the west coast. Also from the area around Hong Kong (Williams 1971) coastal terraces and old beaches have been reported: they occur at +5 and +14 m (Berry 1961) and are probably also present elsewhere. At higher levels, up to 70 m, benches are found, below present sea level flat surfaces can be distinguished down to 60 m. Especially the terraces at +5 and +14 m are thought to be caused by eustatic sea-level changes. Flat surfaces at +130 and +230 m are probably not of marine origin.

Around the Gulf of Bohai, between the Liaodong and Shandong peninsulas, the coast has been filled with recent alluvium from the Huang He (Yellow River), Luan He, Liao, and a number of smaller rivers. Long stretches of tidal marshland have been formed. Along the northwestern shore, however, up to 80 km of alluvial deposits were eroded again in historical times (Von Wissmann 1940), probably because the Huang He then had a southerly course and no new sediment from that river reached the Gulf of Bohai at that time. The present situation dates from 1853, when the Huang He again took a northerly course. Between 1448 and 1853 it had been reaching the sea south of the Shandong Peninsula after having joined the Huai He River. While at present the silt from the Huang He moves northward anticlockwise through the Gulf of Bohai, the silt flowing out of the southern mouth before 1853 was transported further south as far as the Yang-tse (Chang Jiang) River mouth, which was deflected southward. In 400 years the coastline in this area moved 36–62 km eastward.

South of the Yang-tse-Kiang (Chang Jiang) a ria coast ends at the delta of the Si-Kiang, while further south the coast is interrupted by the deltas of the Red River, the Mekong, the Chao Phraya, and other smaller rivers. Tidal flats of variable widths have been formed along Hangzhou Bay, Wenzhou Bay, and numerous smaller bays and estuaries. Salt-marshes along the upper parts of the flats have mostly been reclaimed. South of Hong Kong mangrove swamps are present and fringing coral reefs become increasingly numerous; in particular around the Gulf of Thailand extensive coastal barriers and sandy beaches have been formed. From Dong Hoi to Cape Varella in Vietnam the coast is formed by a 15–40 km wide belt of barrier complexes, lagoons, and sand dunes; from Cape Varella to Cape Padaran the coast is again a typical ria coast. Terraces have been found in Vietnam and Cambodia between +2 and +80 m (Carbonnel 1964; Saurin 1965). The terrace at 2 m and raised reefs, mollusks (oysters), and conglomerates at the same level have been found all along the coasts of Vietnam and Cambodia. Other terraces occur more regionally; the 4 m terrace only in Vietnam, the +15 and +25 m terraces in southern Vietnam and Cambodia, and the terrace at 80 m only near Cape Padaran.

Taiwan, Philippines

Compared to the mainland coast, Taiwan is very different. Situated in the tectonically active zone it has been slowly uplifted during the Quaternary. This uplifting continues at present, although there is some evidence for regional temporary subsidence. Eustatic sea-level changes are considered to have had little effect on the coast, since the changes in sea level have been more than balanced by the amount of uplift

(Hsu 1962). On the Pacific side, the coast is steep and erosion predominates (as also on the southwestern and northern coasts), but on the west coast, which is sheltered from the ocean swell and the tropical storms as well as from the strong northeastern monsoon, a flat, marshy coast is formed with tidal flats. Coral reefs are present on the Pacific side as well as round the Ryukyu Islands further north and along the Pacific coasts of Kyushu and Shikoku.

The Philippine coasts, situated like those of eastern Taiwan in the tectonically active zone, are comparatively steep but intersected by river outlets, broad valleys, and inlets with beaches, surrounded by coral reefs. Although the area has been drowned during the Holocene transgression, the islands have been rising during the Pleistocene and Holocene and there is evidence of differential uplift. Numerous and often extensive marine terraces are present; on Sabtan Island and on Luzon the highest are at +180 m, on Bataan at +275 m, and in northern Mindanao at 360 m (Van Bemmelen 1970), but no correlations have been made.

Malay Peninsula

The Malay Peninsula, together with the eastern part of Sumatra, western and central Kalimantan (Borneo), and the Sunda-Java Sea, belongs to the Sunda landmass, which has probably been tectonically stable since the middle Pleistocene. The presence of beach ridges several meters above present sea level has been interpreted as being the result of recent uplift, but can also be explained by crustal rebound after the Pleistocene and the increasing water load of the oceans (Clark et al. 1978).

The east coast of Malaya is relatively smooth with shallow bays and few indentations. More than 80% of this coast is formed by sandy beaches interrupted by river outlets and small deltas; cliffs form about 10% of the coast (Nossin 1965; Swan 1968). Tidal swamps with mangrove are usually found landward of the beaches and sandbars. They are sheltered against the swell of the strong monsoon that comes from the northeast between October and April. During this period there is more erosion than accretion, especially between late October and January, when spring tides are at their highest. Accretion is widespread between February and September, when the tides are less effective and the wind blows mainly offshore. A minor period of erosion occurs from May to July when the spring tides reach secondary maxima.

The west coast of Malaya is for a large part covered with mangrove, which continues northward to the Irrawaddy delta and from there to the mouths of the Ganges. Locally, this mangrove belt is interrupted by cliffs, river outlets, and beaches. The islands around Singapore have been studied in detail by Swan (1971), who has drawn attention to the fact that in this

sheltered, low-energy coast the coastal forms are very different from those supposed to be characteristic for such coasts in the humid tropics. Mangrove swamps, tidal flats, beaches, and fringing reefs, are present as well as cliffs, caves, shore platforms, and beach conglomerate (ironstone). This diversity is due to the intense and continuous chemical weathering, which makes possible subaerial and marine erosion by small waves, and to differences in rock type and exposure. Coastal sediment is produced through cliff erosion, which takes place along both sheltered and exposed parts of the coast.

Indonesia

In Indonesia, mangrove swamps are present along the coasts of Sumatra, northern Java, along most of Kalimantan (Borneo), along northern Borneo (Sarawak), and at the southern end of Sulawesi (Celebes). All these areas are relatively sheltered against the ocean swell. Cliffs and beaches are locally present (Wall 1964), but although the rivers of Sumatra and Kalimantan carry large amounts of sediment to the sea, only the Kapuas and Pawan rivers, the Rajang and Baram rivers on the northern coast, and the Mahakam River in the east have built up sizeable deltas. Elsewhere, where sediment supply is abundant, broad alluvial lowlands have been formed. The coasts of Java are predominantly beaches interrupted by river outlets and deltas and locally by cliffs. However, along eastern Java, western Sumatra, and on most of the eastern Indonesian islands, the coasts are relatively steep. A few river deltas with associated swamps and alluvial foreland are present on the west coast of Sumatra and on the northeast coast of Java.

The main feature in western Indonesia is the large, drowned Sundaland (Molengraaff 1921). Two large submarine valley systems, being the continuation of the present rivers in Sumatra and Kalimantan, are present on the continental shelf. During the Pleistocene periods of low sea level, one system drained Sumatra and east Kalimantan, discharging into the South China Sea. The other system, draining Java and south Kalimantan, discharged south of Makassar Strait. A divide crossed the Sunda Sea between Sumatra and Kalimantan across Billiton and the Karimata Islands. The Sundaland plain has been dissected at least twice during the Pleistocene as a consequence of the eustatic lowering of sea level (Van Overeem 1960). In the whole area, including Malaya, eustatic terraces are present from 50 to 90 m (Tjia 1970).

On the former east coast of the Sundaland, a large barrier reef has been formed, stretching from Balikpapan on Kalimantan to the island of Sumbawa in the south. It is interrupted in many places and shows a large gap of about 100 km wide facing a deep embayment in the former

Sundaland. The main river outlet was probably situated here. The reef began as a late Pleistocene fringing reef and has grown upward with the gradual rise in sea level. Locally, it reaches the sea surface as separate coral islands. Similar fringing reefs of smaller dimensions are present around Sulawesi and the smaller islands in eastern Indonesia, along the west coast of Sumatra (where they show gaps in front of the river deltas), and further north along the Nicobar and Andaman Islands. In the Sunda Sea, coral growth is restricted to a number of isolated areas away from muddy river outlets (Kuenen 1933; Umbgrove 1947). In eastern Indonesia, numerous atolls and barrier reefs have grown upward from gradually subsiding submarine ridges and platforms, rising abruptly and steeply from a depth of 1,000–2,000 m. The effects of winds and waves on the reefs are conspicuous, especially in eastern Indonesia, where the reefs grow more vigorously on the windward side. Sea currents cause erosion and may shape a whole group of reefs. Solution of coral occurs within the tidal range. On the former northern coast of the Sundaland no barrier reef has been formed. Here the former coast was flat and gradually merged into an extensive sandy and probably muddy shelf. The water in this area was presumably too turbid for coral reef growth. This is also the case along the south coast of Irian Jaya (New Guinea), where large alluvial plains have been formed, covered along the coast by mangrove.

Cross-References

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Asia, Middle East, Coastal Ecology and Geomorphology

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Between the Bosphorus to the west and the Iran-Pakistan border to the east, the coasts of the Middle East may be divided into two different provinces, the Mediterranean one to the west and the Indian Ocean one to the east. The first province concerns the Aegean and Mediterranean coasts of Turkey on the one hand and the Levantine coast on the other hand. The second province includes three rather different sectors: the western coast of the Red Sea and the coasts of the Persian Gulf, two elongated and narrow inland seas, and the Indian Ocean coasts, comprising the south Arabian coast, the Gulf of Oman, and the Makran coasts. Each of these sectors presents its own particularities together with some common characteristics, whether in tectonics, climatology, ecology, or geomorphology (Figs. 1, 2, 4, 5, 6, 7, 8, 9).

The Coasts of Anatolia

Climate, Vegetation, Hydrology

Along the Aegean and Mediterranean coasts of Anatolia, the annual precipitation ranges from 400 mm in the plains, with a minimum in the sheltered areas like the Çukurova plain, to more than 1000 mm on the hills and flanks of mountains, notably in the Lycian sector. In winter, the Aegean and Mediterranean coastal areas are by far milder than in central Anatolia, with mean January temperature higher than 10 °C in sectors bordering the sea along the Mediterranean Sea and higher than 5 °C on the lower slopes of the hills and mountains. But everywhere freezing may occur, especially in the Aegean area. Summers are very hot; mean August temperatures exceed 25 °C and maximum absolute temperatures reach 40 °C or more, particularly in the Cilician plain. Rainfall occurs in winter and the summer is dry for about 6 months. Vegetation is much more xerophytic than in the West Mediterranean province. The seaward slopes of the coastal mountain ridges are truly Mediterranean in

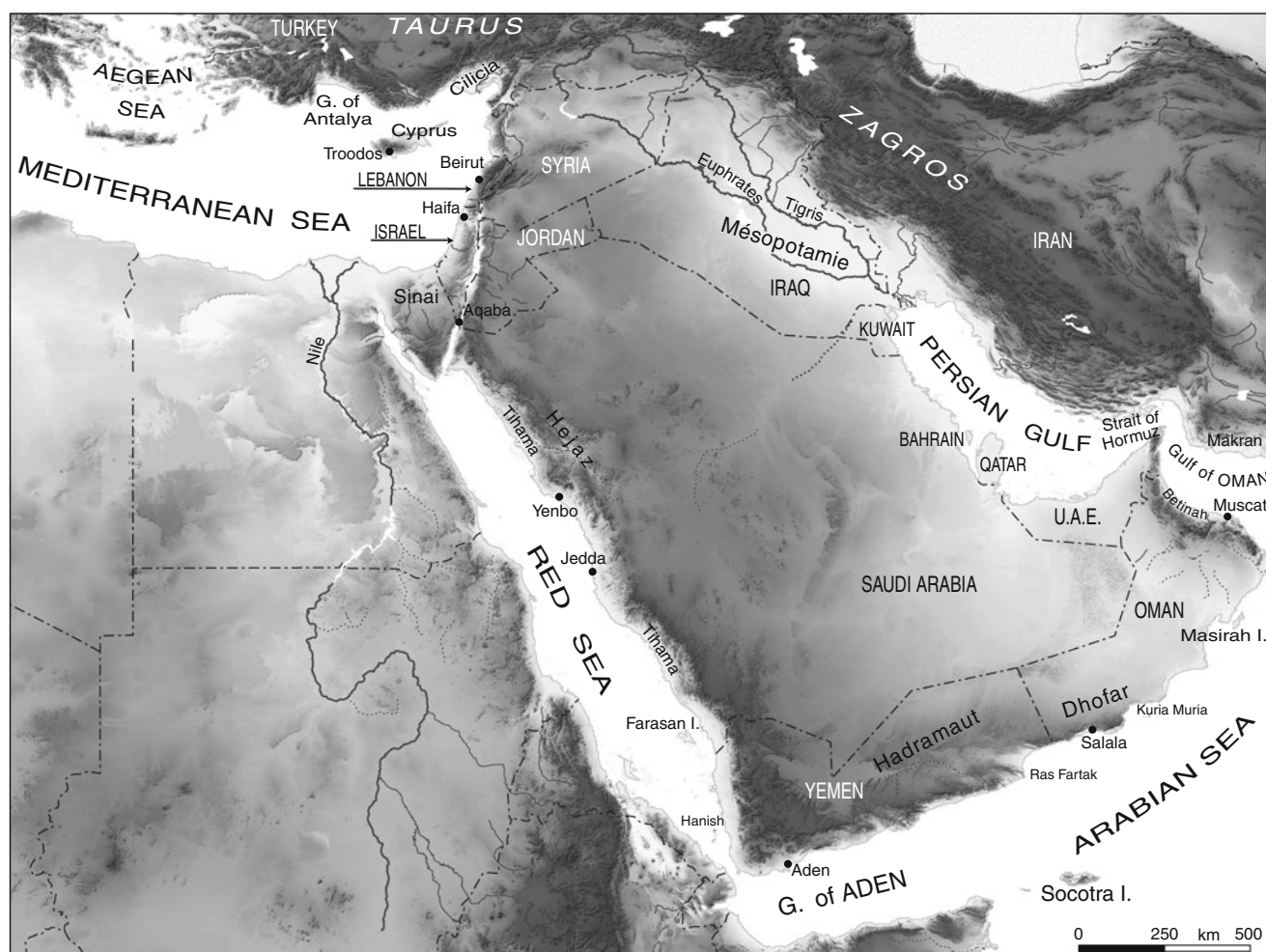
vegetation and in climate. Along the Amanus Mountains, the evergreen maquis zone of *Quercus calliprinos* and *Pistacia palaestina* sometimes starts at sea level and extends to an altitude of about 700 m. In the Lycian-Cilician sector, the eu-Mediterranean belt is confined to a very narrow ribbon with the same plant association of the evergreen maquis. In the Aegean sector, the evergreen maquis mainly comprises the association *Pistacia lentiscus*, *Olea eumpea*, and *Myrtus* (Zohary 1973).

The ratio of precipitation/evaporation shows a deficit and the eastern Mediterranean Sea receives water coming from the Black Sea through the Bosphorus and Dardanelles Straits, the salinity in the Aegean Sea is notably lower than in the eastern Mediterranean; in August, it is less than 38‰ in the north, less than 39‰ in the south of the Aegean Sea, whereas it is more than 39‰ in the eastern Mediterranean along the Turkish and Levantine coasts. Marine currents are generally north-south in the Aegean Sea and east-west along the Mediterranean coast of Turkey. On the whole, sea water temperatures increase from the north to the south in the Aegean Sea and from the west to the east in the Mediterranean, rising for example in August from 25 °C along the south-west coast of Anatolia to over 28 °C along the Cilician coast.

Regional Features

The Turkish Aegean coastline is very long (3484 km) and indented. It presents a great variety of lithological units, with alternation of crystalline massifs and Mesozoic geosynclinal belts including mostly limestone, flysch, and ophiolitic formations. This area is tectonically controlled by a highly complex structure of east-west oriented active horsts and grabens. Peninsulas, islands, rocky cliffs, and a narrow continental shelf correspond to the horsts. In the grabens there are wide continental shelves prolonging vast deltaic plains bordered with marshes, dunes, and sandy beaches. In the north, the Sea of Marmara, related to the greater Anatolian fault, is very deep (> 1300 m) (Fig. 1).

During the Middle Miocene severe tectonic phases occurred and grabens formed. These grabens were submerged by marine transgressions during the Quaternary, and during the glacial periods rivers deepened their courses. During the Holocene, when the sea-level rise had reached its present level about 6000 years ago, major marine embayments occurred in several depressions where the main rivers ended; such as the Gediz, the Küçük Menderes, and the Büyük Menderes. Many important cities of the Hellenistic to Roman times were located on the immediate coastline and closely associated with the sea, such as Ephes, Priene, Milatus, or Heracleia; in the vicinity of Ancient Troy, the marine embayment protruded nearly 10 km south of the site. But, quickly, deltaic progradation and floodplain aggradation, caused by deforestation and severe erosion on the steep slopes

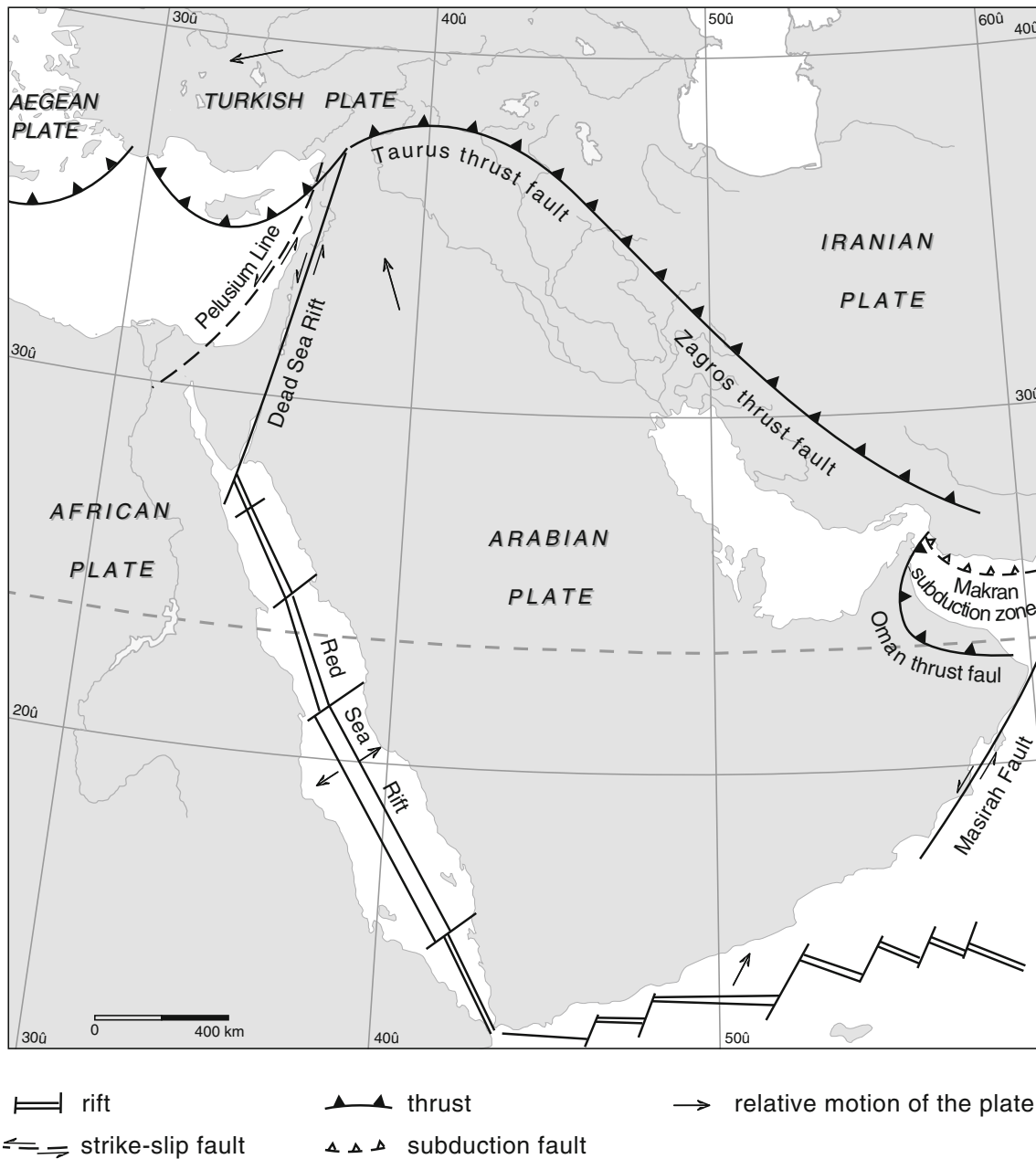


Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 1 Location map

of this mountainous country, markedly changed the landscape: the depressions were infilled with prograding river alluvium and deltas. These towns were separated from the sea; Ephes in the plain of the Küçük Menderes and Priene or Milatus in the Büyük Menderes basin, and the site of Troy is now about 6 km from the sea. Sometimes, for instance, in the Gediz Basin and the Izmir area, it seems that the evolution was more complex and mainly controlled by intense tectonics. The progradation was less marked where there is no major drainage system to provide sediment, for instance, in the south as in the Gökova Gulf. Now, progradation has largely halted mainly where the grabens widen rapidly into larger coastal embayments and modern deltas are wave dominated (Kraft et al. 1980; Erol 1985; Aksu et al. 1987; Hakyemez et al. 1999; Kayan 1999).

The Mediterranean coastline of Turkey (1707 km) is more regular and rectilinear than the Aegean coast, because of its longitudinal structure. The coastline is subparallel to the Taurus Mountains consisting of Paleozoic to Tertiary folded formations where carbonate rocks outcrop largely,

favoring numerous karstic features. West of Antalya, the Taurus Mountains, S-N oriented, rise sharply and steeply out of the sea with high cliffs and only very small deltaic coastal plains. From Antalya to Alanya, there is a major embayment with a wide coastal plain edged by extensive coastal dunes and sandy beaches with frequent beach-rock formations; the only exception are the travertine cliffs at Antalya. Marine Pleistocene terraces occur to the east of the Antalya Bay at heights from 5 m to 20–30 m or even higher. Here, as in the Hatay area, beaches and bio-erosional notches and benches are frequent, between +0.80 and +2.5 m, testifying to late Holocene uplifts (Pirazzoli 1986; Kayan 1994). East of Alanya, once more, the Taurus Mountains rise directly out of the sea with very steep flanks and cliffs in Paleozoic schists and Mesozoic limestones. Scattered sandy beaches exist at the mouth of the rivers but elsewhere the coastline is dominated by plunging cliffs or by pebble beaches at the foot of the cliffs. The main river is the Göksu, characterized by a small lobed delta plain. East of Göksu, between the SW-NE Taurus Mountains and the N-S



Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 2 Tectonic map

Amanus mountains (or Nur Mountains), there is the large Çukurova plain (or Cilician plain) which is a progradational plain built by two big rivers, the Seyhan to the west and the Ceyhan to the east. The plain, formerly swampy or marshy, but now largely drained and dry, is edged by a series of broad Holocene accretion beaches. At the foot of the Amanus Mountains, the rectangular-shaped Gulf of Iskenderun is structurally controlled. At the very southern end of the coast of Turkey, the Orontes River crosses the Amanus Mountains in a narrow defile and ends in a small delta near the ancient harbor of Seleucia ad Orontes (Erol 1985).

Almost everywhere, the coastline is now receding owing to the taking of sediments on beaches and dunes, the setting up of dams, and slight raising of the sea level, and slabs of beach rock are more and more frequent on the beaches.

Cyprus Island lies south of Turkey. Its coastline, about 450 km long, is partly steep and rocky in the north (notably in the Karpos Peninsula) and in the SW (at the foot of Troodos massif) and partly beach-fringed, particularly in the southeast, alongside the Mesogea central plain, mainly between Limassol and Famagusta where there are some brackish lagoons (Fig. 2).

The Levantine Coast

Climate and Marine Hydrology

The climate is clearly Mediterranean, with hot and dry summers, and rainy and mild winters and the winds mainly blowing from the west. Average rainfall decreases from the north (ca. 1000 mm north of Latakia) to the south (313 mm at Gaza). Average air temperature along the coast usually ranges from 11 °C to 14 °C in January (with minima reaching the freezing point) and from 23 °C to 26 °C in August with absolute maximum temperatures over 45 °C. Tidal range does not exceed 40 cm. There is an anticlockwise east Mediterranean current, which moves south-north along the Levantine coast. Longshore currents occur too, influenced by coastal morphology and wind direction, but the resultant shifting is to the north. All these currents carry the Nile sediments northward up to south Lebanon. Westerly and especially southwest winds are dominant. As the fetch is large, the coast undergoes strong wave action. In winter, storms are sometimes severe: in January 1968, with wind speeds of almost 150 km per hour, wave heights reached 7 m off the Beirut region.

The living world of the eastern coasts of the Mediterranean is impoverished and the littoral fauna and flora are poorly diversified in comparison to that of the western Mediterranean as a result of the extremely low biological productivity of the Levantine basin, mainly since the opening of the Suez Canal and the erection of the Upper Assouan Dam. Since the opening of the Suez Canal, over 300 migrant species have invaded the Mediterranean from the Red Sea at a faster and faster rate; nine new species of fish arrived between 1902 and 1939, but 22 from 1965 to 1987. Among the “Lessepsian” crustaceans, decapods are particularly representative, such as *Myra fugax* or *Atergatis roseus*, and *Charybdis longicollis* represents up to 70% of the sandy-muddy bottoms. Among the Serpulids, seven Lessepsian species have been identified, belonging to three genera, and 33 algae are possibly lessepsian (Por 1982, 1990).

Coastal Morphology

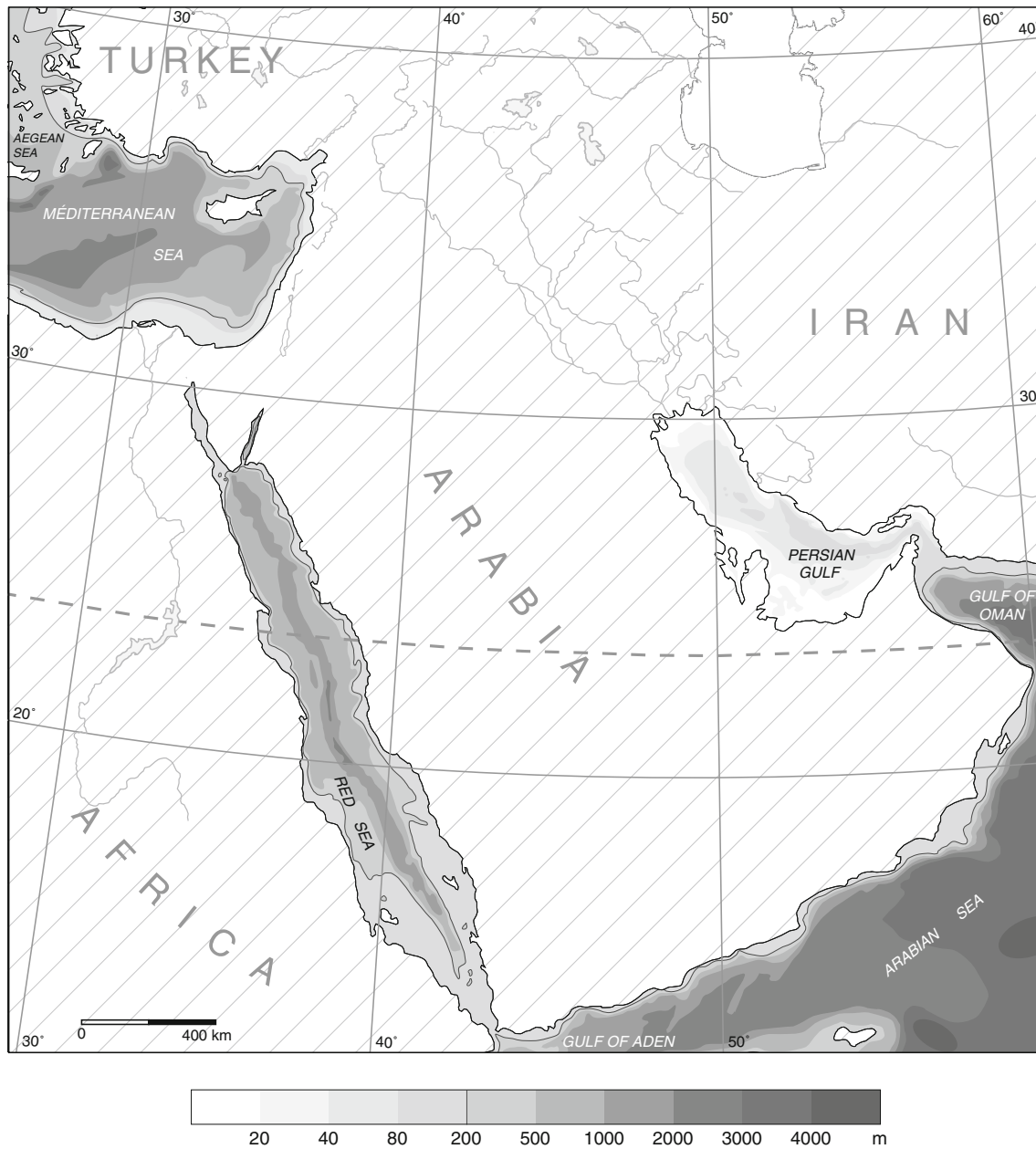
The rectilinear Levantine coast is open to the western winds and swells. The only noteworthy irregularities of the coast are such peninsulas as Ras el-Bassit in Syria, Ras el-Mina (near Tripoli), Ras Chekka, Ras Beirut, and Tyr in Lebanon, and Mount Carmel in Israel. Some of them (Tripoli or Tyr) are ancient rocky islands more or less recently connected to the mainland by a sandy bar (tombolo). Bays are generally small, the most important being the Akkar Bay, on the Syria-Lebanon frontier, and the Haifa Bay in Israel. Transverse faults, often east-west, account for the main features of the coastline. The north Levantine coast runs parallel to the

folded coastal mountain range of Jabal Ansarieh in Syria and Mount Lebanon (culminating at 3083 m) in Lebanon. The mountain range approaches very close to the Mediterranean Sea, mainly in Lebanon where, in the Jounieh Bay, for example, the altitude reaches 600 m less than 1 km from the sea as the crow flies. Nevertheless, high cliffs are uncommon, except in the Jabal Akra in the north of Syria, or at Ras Chekka or Ras Beirut in Lebanon. In the northern Levant, the continental shelf is narrow and broken by very deep and steep canyons, as in the Beirut Bay; in the south, the continental shelf gets wider and wider and its slope is very gentle. The coastal plain is discontinuous and narrow in the north Levant, widening in south Lebanon, between Saida and Tyr, and even more south of the Mount Carmel; getting wider and wider to the south, it reaches a width of 40 km near Gaza. On the other hand, north of Haifa, the coast is Relicts of Pleistocene *kurkar* (dune calcarenite, called *ramleh* in Arabic) ridges form several parallel alignments in the coastal plain south of Mount Carmel, witnesses of several coastal dunes associated with transgressive coastlines; the waves sometimes attack the westernmost *kurkar*, cutting it into a cliff preceded by a wide rocky platform (Nir 1982). The youngest one also occurs as small islands or islets off the coast of Palestine, Lebanon (at Sidon and near Tripoli), and Syria (Arados) (Fig. 3).

Syria, Lebanon, and Mount Carmel show fine examples of little deformed, but strongly uplifted Pleistocene coastlines up to 300 m above present sea level (Sanlaville 1977), resembling wide horizontal flat platforms edged with indurated beach sediments, sometimes containing paleolithic artifacts. The last interglacial deposits, between 6 m and 12 m, contain a rich Tyrrhenian fauna; at Naame, 15 km south of Beirut, 14 species living on the western coast of Africa but missing today on the Mediterranean coast, were found, such as *Strombus bubonius*, *Cantharus viverratus*, *Conus testudinarius*, *Arca afra*, *Cymatium costatum*, etc. (Fleisch et al. 1981). Holocene marine deposits are also frequent, indicating recent oscillations of the sea level or tectonic movements up to 2 m, but the aggradation beaches were often clearly in phase with fluvial discharges controlled by climatic oscillations or anthropic factors (Sanlaville et al. 1997).

Erosion and Pollution

Some decades ago, beach rock occurred here and there along the Levantine coast but it is now widespread all along the coast, in relation to strong beach erosion which results in the disappearance of the beach sand and the exhumation of an ancient slightly consolidated buried beach, then strongly indurated when exhumed. Everywhere the coast is retreating, sometimes rapidly, and beach sediments and dune sands disappear quickly. This coast is also badly polluted and the landscapes are greatly altered by sea walls, breakwaters, harbors, and all sorts of concrete buildings. Because of the westerly swell, the presence of slabs of beach rock at the



Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 3 Bathymetric map

lower edge of beaches or the presence of deep and sharp lapiaz on the rocky coastal areas, and severe pollution, this coast is not as attractive as it was, despite good climatic conditions (Fig. 4).

The Red Sea Coast

Structure, Climate, and Marine Hydrology

The Red Sea occupies a complex rift system between the African and Arabian plates, when separation occurred in the mid-Tertiary times. The Red Sea opened into the Indian

Ocean in Miocene times and remained open into the Mediterranean until the early Quaternary. The beginning graben event reactivated erosion and due to the penetration of the sea, carbonates and evaporites were deposited, with coarse-clastic intercalations. With the closure of the Mediterranean connection and tectonic and eustatic changes affecting the Bab al Mandeb narrows, the Red Sea became isolated during the later Quaternary and, because of intense evaporation, greatly diminished in volume with a dramatic impact on shore-zone morphology and ecology.

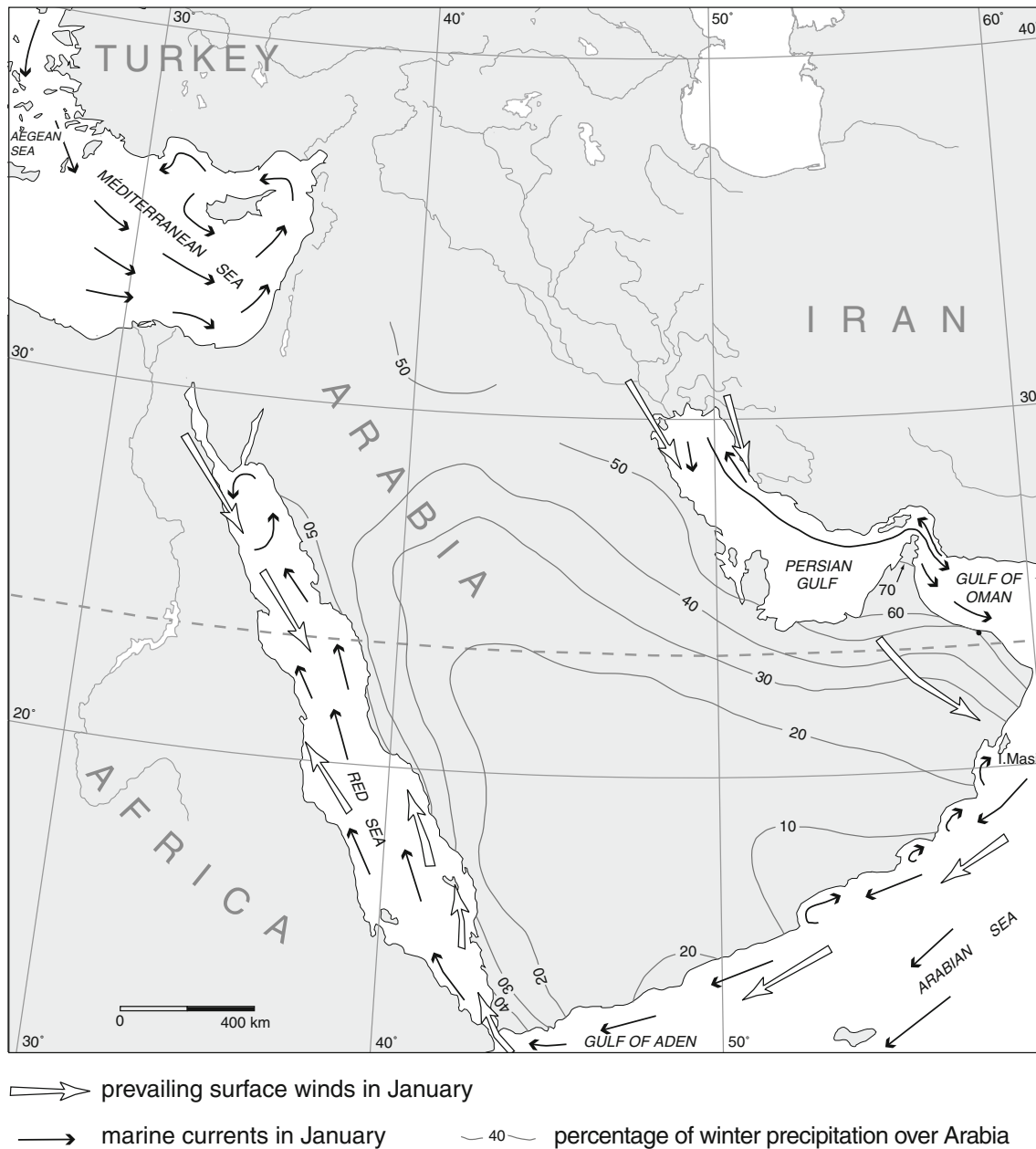
Connection between the waters of the Red Sea and the Indian Ocean is hindered by a shallow submarine sill in the



Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 4 Sea surface salinity (‰ in August)

Bab el Mandeb narrows; the narrows are only 20 km wide and near the Hanish Island, the sill does not exceed –137 m. In the north, the Suez Canal established an artificial but very weak contact with the Mediterranean Sea. The climate of the area is hot and arid. In February, the temperature in the coastal plain (Tihama) varies from 12 °C in the north to over 26 °C in the south (23.5 °C in Jeddah), and in August it is over 30 °C (Jeddah, 32.4 °C). The Tihama receives less than 50 mm of rain. North of Jeddah, the rainfall does not reach 100 mm in the mountains (occurring in winter and spring), but it is more important in the south: over 400 mm in the Asir and over 600 mm in Yemen (in summer and spring) where mountains are very high. But water is lost before reaching the sea, so the Red Sea does not receive freshwater, since evaporation is very

high (more than 2 m per year). The shortage must be supplied by the Indian Ocean through the Bab al Mandeb, where three currents are superposed in summer and two in winter as a result of water density and seasonal winds. So the Red Sea is warm and very salty. Surface sea water temperature stands between 28 °C and 31 °C in the northern part in August (under 28 °C in the Aqaba Gulf) and rises above 31 °C in the southern part. In February, the surface sea water temperature rises from the north (less than 20 °C in the Aqaba Gulf) to the south (26 °C). Salinity is very high in summer, increasing regularly from the south (37‰) to the north (over 41‰ in the Gulf of Suez). Sea water temperature and transparency are very favorable to coral reefs and the mangal vegetation is well developed, dominated by *Avicennia marina*; along the



Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 5 Winds and currents in January

southern Red Sea coast there is also found *Rhizophora mucronata*, *Bruguiera gymnorrhiza*, and *Ceriops tagal* (Glover 1998). Winds are predominantly from the NW all over the Red Sea in summer; in winter, NW winds occur in the northern part of the Red Sea but SE winds blow in the southern part. The marine currents are oriented north-south in summer and south-north in winter. Tidal range is insignificant; 0.10 m in Jeddah, but atmospheric pressure and winds can bring about variations of the sea level up to 0.90 m, which is very dangerous for boats due to coral reefs (Fig. 5).

Coastal Morphology

The landscape of the region is predominantly mountainous; the highest peaks are often in excess of 2500 m, mainly in the south, in Asir or Yemen where mountains reach a height of over 3000 m and up to 3760 m at the Jabal Nabi Shuaib, in Yemen. In places, the strongly dissected mountains come right down to the sea. But generally a flat plain, 5–40 km wide, constitutes an almost 2000 km long marginal corridor, called the Tihama, following the border of the Arabian Mountains. Plutonic, volcanic, and metamorphic rocks outcrop in the mountains. In the Tihama, the basement is generally

hidden below Tertiary and Quaternary rocks. Tertiary sediments, which can reach a thickness of over 2000 m, generally present a division into three series: Oligocene, Lower, and Middle Miocene limestones and conglomerates, Middle and Upper Miocene evaporites, and Upper Miocene-Pliocene limestones and clastic red series. Frequently, Tertiary rocks are exposed only at the edge of the basement and in a few places in the coastal plain where mainly Quaternary deposits outcrop, constituted of gravel terraces, extensive alluvial cones and uplifted coral reefs.

Coming out of the mountains, the alluvial surfaces fall relatively steeply to the coast. They can be recognized by their covering of pebbles and gravels with particularly dark desert varnish. Upstream, near the basement, a terrace consists of a thick sequence of coarse gravels and downstream mainly of fine sediments. There are generally several stepped alluvial terraces, the oldest characterized by coarse pebbles and a very dark patina, the youngest by a sandy to gravelly material and a lighter color. Relatively coarse windblown sands often occur in the upstream part of the coastal plain. Dating the fluvial terraces is rather difficult, but relative dating is sometimes possible. Between Umm Lajj and Yanbu, the basalts from the Jabal Nabah, K/Ar dated 1.4 ± 0.6 My, flowed into wadis that were already cut into the oldest terrace, which might be Lower Quaternary or Upper Pliocene and show evidence of vertical block movements; the formation of the middle terrace antedates the Umm Lajj basalts (0.4 ± 0.2 My). The age of the youngest gravel terrace can be estimated, due thanks to its direct connection to the 6–8 m marine terrace, which belongs to the last interglacial (Hötzl et al. 1984).

Ancient coral reefs outcrop in the lower part of the coastal plain, in places slightly covered with gravels. The main reef body, very flat, is the 6–8 m marine terrace. Toward the sea, the main reef itself outcrops and, behind it, increasingly lagoonal facies. This ancient reef can be attributed to the last interglacial (isotopic stage 5e) and exhibits no real deformation. An older one, faulted, is probably 250 ky old (isotopic stage 7). A Holocene erosion step at 1–2 m above m.s.l., only a few meters wide, is carved into the edges of the 6–8 m reef, even on its landward side. The reef terraces form striking isolated erosional remnants and more or less broad table-like crests. Depressions and channels developed behind the reefs, which, subsequently, were sometimes occupied by bays (Sharm) when the sea level rose later. The best example is Sharm Yanbu with its narrow deep channel and its two branches representing the continuation of two wadi channels. The protective reef plate kept this crest from being eroded, while the loose material behind it was excavated. Such back-coast depressions recur repeatedly. These bays are the result of the last drop in sea level in the latest Pleistocene and the later increase in the Holocene. On the eastern edge of the reef

crest, an inter-fingering of coral and beach formations with clastic continental sediments can often be seen. This interfinger-ing points to the existence of a humid climate during the interglacial period. Carbon-14 dated remnants of trees indicate heavier-than-today precipitation during the so-called “Neolithic Pluvial” (Fig. 6).

Coral Reefs

Coral reefs are mainly of the fringing type; fringing reefs in a very nearshore position in front of steep cliffs, or fringing reefs with a lagoon within shallow water. They grow close to the mainland and are absent from wadi mouths. Barrier reefs and even atolls also occur, mainly in the central and southern part of the Red Sea. The outlines of the reef crest and the orientation of the foreslopes follow the tectonic pattern. Spectacular drop-offs are widespread and represent predominantly horst and graben structures parallel to the rift.

In the Aqaba Gulf, the fringing reefs exhibit a characteristic biozonation (Dullo and Montaggioni 1998): (1) a local and poorly developed beach; (2) the back-reef zone, a sandy depression between the beach and the reef-flat, with a maximum width of 50 m and a depth of 1–2 m, the bottom sediment is colonized by scattered coral heads (*Stylophora*, *Serkitopom*, *Platygyra*, *Mitteporea*), alcyonarian colonies (*Lithophyton*, *Cladiella*, *Sinularid*) and seagrass beds (*Halophila*, *Halodule*); (3) the reef-flat zone, about 20 m wide, this is a dead coral pavement bordered by a 1 m rear-reef step. Coral communities are composed of small-sized colonies (*Stylophora*, *Seriatopom*, *Acropord*) associated with hydrocorals (*Mitteporea*). The reef front forms a nearly vertical drop-off, 2–1 m high, with *Mitteporea* and branching red algae; (4) the outer slope zone, with loose sedimentary slopes (sandy talus and patches of *Halophila*) or coral-reef slopes, with their upper parts dominated by branching growth forms (*Stylophora*, *Seriatopora*, *Acropora*, *Echinopord*). Along the coast of Arabia, the fringing reef constitutes an almost continuous belt and often has wide back-reef zones.

On the central and southern coast of Arabia, coral reefs also occur as offshore knoll reef platforms. In the vicinity of Jeddah, about 3 km west of the coastline, the offshore reef consists of a shallow platform covering an area of 800 km², bounded both shorewards and seawards by a vertical escarpment dropping rapidly to depths of 400–800 m, marking the edge of the Red Sea trough. The platform is occupied by scattered reefs separating sandy bottoms. The present reefs come from Holocene reefs settled on Pleistocene reef limestones. The coral community of the innermost part is dominated by *Stylophora pistillata*. Coralline algae are represented by massive branching and crustose forms (*Lithothamnium*, *Spongites*). The upper parts of the steeply reef flanks are settled by *Millopora*.



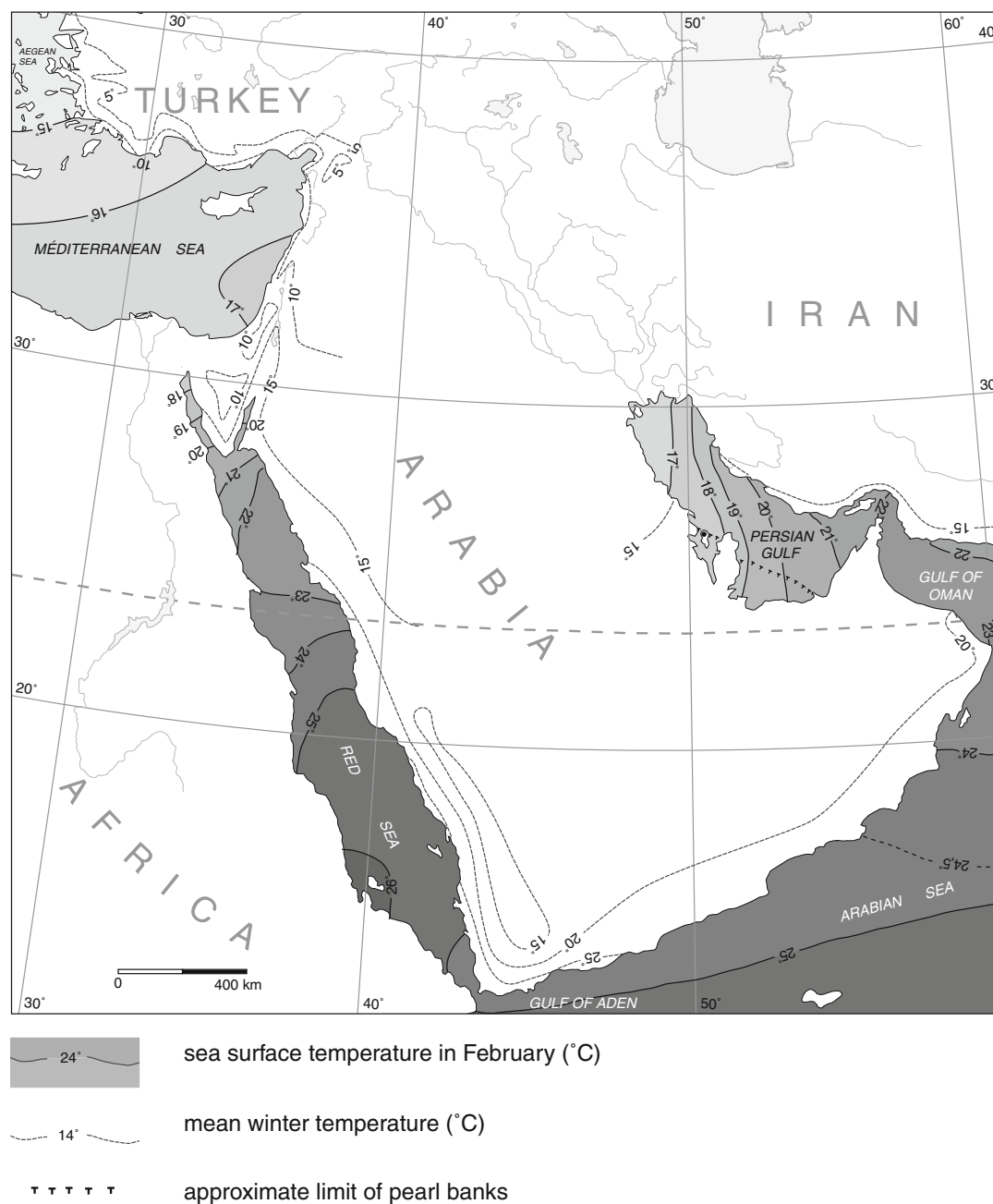
Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 6 Winds and currents in July

In the southern part of the Red Sea, where the sea reaches its maximum width of 360 km, remarkably flat shoals (called *shab*) occur. The best example is the Farasan bank which extends from 16°N to 22°N. Lat. and reaches a width of up to 120 km. The sea is almost always less than 100 m deep and often only a few meters deep. The two main islands, which were very important for pearling at the beginning of the twentieth century, are 60 km long and 8 km wide (al Kabir) and 35 km long and 10 km wide (Sajid). These islands are flat

and low (under 20 m), except in a very few places where heights of up to 75 m are to be found, in domed eminences where gypsum and anhydrite outcrop as a result of halokinesis (Dabbagh et al. 1984).

Regional Features

The Gulf of Aqaba is the continuation of the Red Sea rift, formed as a consequence of transform movements along the Aqaba-Levant structure. This narrow gulf slopes steeply to



Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 7 Mean winter temperatures

depths of over 1800 m. In the north of the Red Sea, the coastline is very straight, free from larger indentations, with a trend of 140/150°, identical to that of the entire graben. Sigmoid curves are typical of the central and southern parts of the Red Sea coast. The 200 m isobath shows significant deviations. It is seldom parallel to the coast, underwater ridges, and troughs repeatedly show the conjugated 70° course indicated by transform faults. The main wadis are generally related to young NE striking faults and block-faults (Fig. 7).

The Wadi al Hamdh, with a catchment extending almost to al Madinah, is the most important drainage system in the Hejaz Mountains. Its load of sediment considerably contributed to the creation of a broad aggradation plain south of al Wajh. Even under the present arid conditions, its water, enriched with mud and rubble, regularly reaches the coastal plain, and sometimes the sea. A continuous reef belt has developed some 30 km away from the coast (Al-Sayari et al. 1984).

Because of a wide break in the recent coral reef along the coast, Jeddah is one of the few places on the Red Sea coast where a large harbor could be located. Isobaths show an E-W channel from Wadi Fatimah into the Red Sea. The upper course of Wadi Fatimah extends into the massive basalt flows of the Harrat Rahat and the remaining catchment is made up mainly of Precambrian rocks. The middle course of Wadi Fatimah is to be considered as a fault-bounded graben. The presence of plutonic and metamorphic rocks limits the potential aquifers to the alluvions of the Wadi Fatimah which is an early source of water for the Jeddah city, as the ruins of a qanat(conduit) system show (Hacker et al. 1984).

The plain of Jizan forms a narrow NW-oriented coastal strip. The strongly dissected Asir and Yemen high mountains overlook the whole area. At their base, parallel foothills, consisting of metamorphic schists or volcanic rocks reach heights of between 200 and 700 m. The plain itself averages about 40 km in width and is composed of alluvial deposits, except for some volcanic cones and the salt dome (diapir) of Jizan (Müller 1984). Several important intermittent wadis flow across the plain, fed by the summer and spring rain falling in the mountains, but only their upper courses are perennial. Many volcanic intrusions, which occurred during the development of the Red Sea rift, are characteristic of the plain. Given the good state of preservation of some cinder cones, volcanic activity has continued up to recent geologic times. The coast is preceded by a reef over almost its entire length. Immediately adjacent to the coast is a sabkha zone, some kilometers wide, consisting of silty, clayey sediments with high salt content. On its eastern border, this zone passes into the alluvial accumulation of the middle terrace (Purser and Bosence 1998).

The Persian Gulf Coast

Hydrology, Climate, and Structure

The Persian Gulf has an average depth of 35 m and a maximum depth of 100 m near its narrow entrance, the Strait of Hormuz, 60 km wide. Tidal range varies from 1 m at Dubai, 2.5 m at Bahrain, and 3.5 m at Kuwait to a maximum of 5 m in the Clarence Strait, between Qeshm Island and the mainland, and the tidal range averages ca. 1 m at Basra, 150 km upstream on the Shatt al Arab. Regional tidal currents are oriented approximately parallel to the axis of the Gulf. Tidal velocity may exceed 60 cm/s, their bidirectional movements favoring the development of spectacular oolitic tidal deltas (Purser and Seibold 1973).

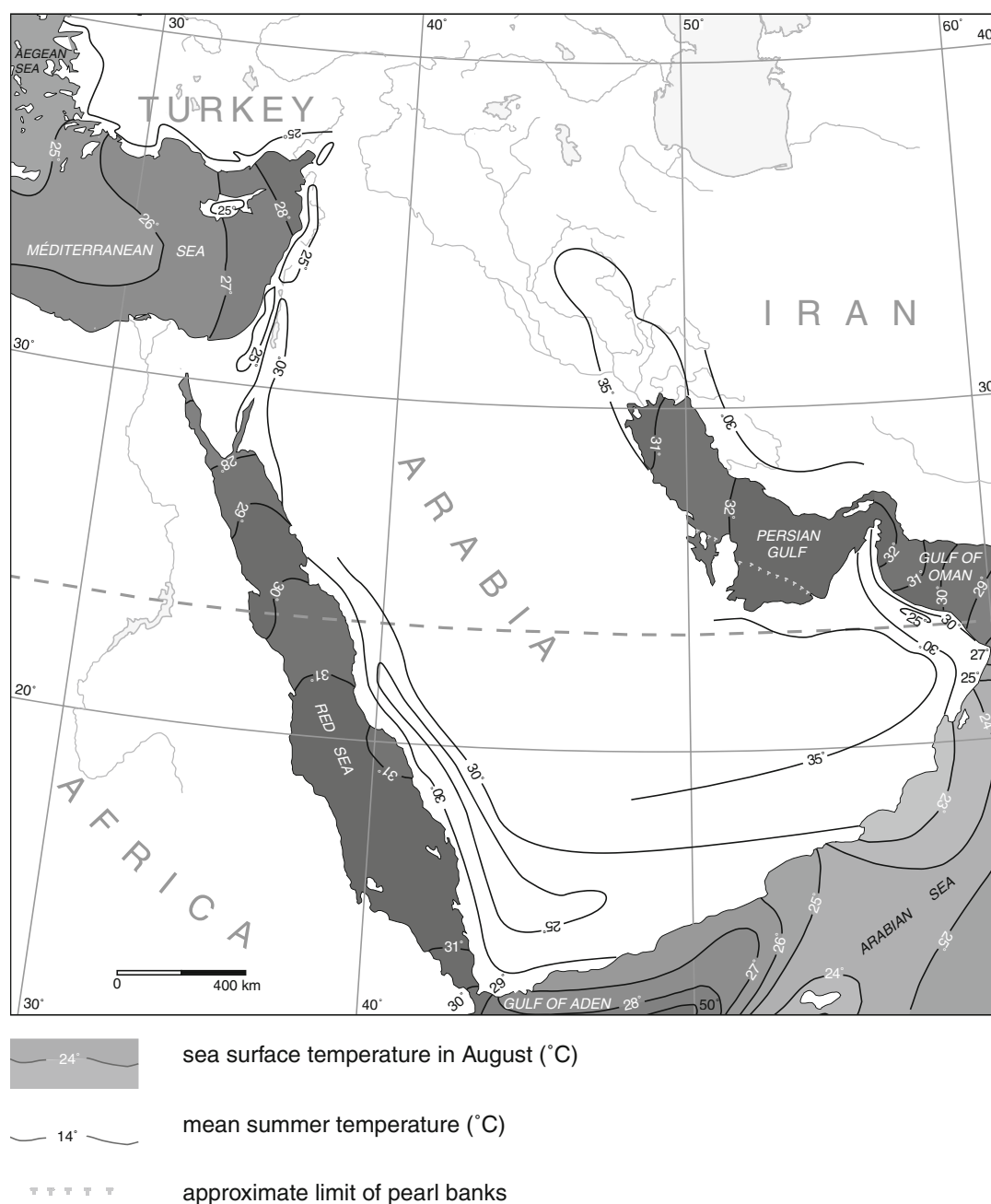
The considerable loss of water in the Gulf through evaporation is not compensated by precipitation and river inflow, the only notable inflow of fresh water coming from the Shatt

al Arab, in fact mainly from the Karun (ca. 6000 m³/s). As a result, compensatory waters come from the Arabian Sea, with surface current moving anticlockwise along the Iranian coast and coming down the Arabian coast, determining the distribution of salinity, temperatures, and nutrients. In summer, surface water attains temperatures over 32 °C in the central part but very much higher in the Arabian lagoons. Temperatures in winter fall below 20 °C in the northern part of the Gulf and even under 18 °C north of Bahrain, where corals and mangrove cannot live (the absolute minimum is −5 °C at Abadan). The range of variation in water temperature tends to increase away from the entrance of the Persian Gulf. The salinity of surface waters is high along the northern Iranian coast (39‰ – 40‰) but diminishes down to 30‰ west of the mouth of the Shatt al Arab, increasing then to the south along the Arabian coast (more than 38‰ east of Qatar and even 60‰ in the Gulf of Salwa) (Fig. 8).

The climate of the Persian Gulf area is hot and semiarid to arid. Mean monthly temperatures increase from 14 °C at Kuwait to 18 °C at Abu Dhabi in January and stand ca. 34 °C in July all along the Arabian coast. Extreme temperatures are cool in winter and torrid in summer, the maximum absolute temperature reaches 49.2 °C at Sharjah and 52 °C at Abqaiq. Mean annual rainfall is less than 100 mm on the Arabian coast, slightly higher on the Lower Mesopotamian Plain and on the Iranian coast (ranging from 100 to 200 mm with a maximum of 230 mm at Bushire), rains occurring mainly in winter with a second peak in spring. The prevailing “Shamal” wind blows down the axis of the Gulf from the NW, then from the north to the south. Summer months are essentially calm, but during the winter the Shamal may be very strong. Mangrove stands are found at intervals along the Persian Gulf and *Avicennia* is the single species because it is the most resistant to low and high temperatures and to an intense evaporation resulting in high soil salinities. The northernmost site is in Bahrain and the greatest areas occur in the United Arab Emirates and in the Clarence Strait, in Iran, but the mangrove environment was richer and more extensive during the Middle Holocene and *Terebralia palustris*, frequent at that time, is now extinct (Glover 1998).

The small but numerous streams draining the Zagros ranges are characterized by rough flash-floods and deliver a significant amount of terrigenous sediments. In contrast, the Arabian side of the basin is completely lacking in fluvial influx so almost pure carbonate sediments predominate along the Arabian coastline. Oolitic sediments are forming both in exposed and sheltered environments and are particularly abundant in the southern part of the Gulf.

From the peak of the last glaciation until about 14,000 year. BP, the Gulf was free of marine influence. By 14,000 year. BP the Strait of Hormuz had opened up as a narrow waterway and by about 12,500 year. ago marine incursion

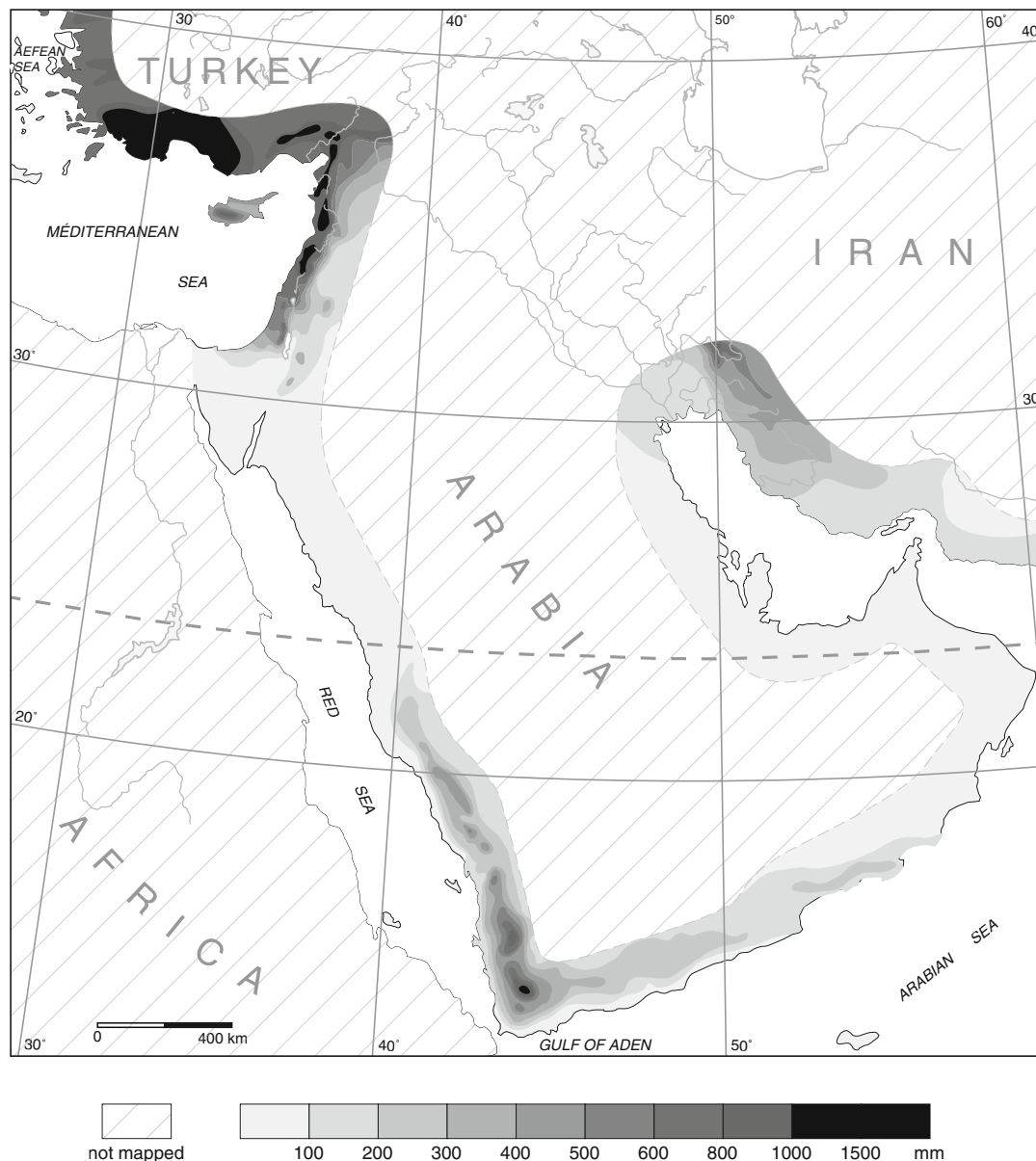


Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 8 Mean summer temperatures

into the central basin had started. The present coastline was reached shortly before 6000 year. BP but the relative sea level then rose 1–2 m above its present level (Lambeck 1996).

The difference in slope between the two flanks of the Gulf reflects the fundamentally different tectonic histories of Iran and Arabia. Its elongate axis separates two distinct geological provinces, the stable Arabian foreland and the unstable Iranian fold belt. The foundations of the modern Persian Gulf were largely laid in the Plio-Pleistocene Zagros orogeny.

The collision of the Arabian and Euro-Asian plates resulted in the formation both of a broad structural depression, about 2000 km long, which extended from Syria to Hormuz and which eventually became an extensive sedimentary basin, and of the Zagros Mountains consisting in anticlines hugely folded with steeply dipping flanks, characterized by a NW-SE oriented belt, with mainly Late Cenozoic sedimentary formations in which clays, silts, and carbonates are predominant. The tectonic instability results in frequent earthquakes (Fig. 9).



Asia, Middle East, Coastal Ecology and Geomorphology, Fig. 9 Mean annual precipitation in coastal areas

The low-lying Arabian coast and adjacent shallow seafloor are characterized by large, low-dipping anticlines with N-S or NE-SW trends, for example, in Bahrain and Qatar. In the Persian Gulf there are some 20 islands, as well as numerous submarine highs or shoals. Most of the islands and submarine shoals are clearly due to salt diapirism which, affecting the early Palaeozoic Hormuz salt series, was very active during the Plio-Pleistocene period and down to recent times (Kassler 1973). The NW-SE Arabian coastline and the straight coastline from Abu Dhabi to Ras al-Khaimah may be partly controlled by faulting.

Regional Features

The Shatt al Arab deltaic system. The extreme north of the Gulf corresponds with the deltaic system of the Shatt al Arab, the combined delta of the Euphrates, Tigris, and Karun. At the latitude of Basra, the two wide alluvial fans of Wadi Batin (Pleistocene) and Karun River (Holocene) restrict the width of the Mesopotamian plain, being responsible for the upstream development of extensive palustrine environments (Hammar Lake and marshes), where the Euphrates and Tigris deliver most of their water and suspended sediment. After loss of their suspended sediment in the marshes and shallow lakes,

the clear waters are drained to the Shatt al Arab (Purser et al. 1982). In fact, much of the fluvial waters and suspended sediment of the Shatt al Arab come from the Karun River. South of the two big alluvial fans, the Shatt al Arab forms a major deltaic system, dominated by strong tidal conditions, with large sebkhas flanking the waterways, wide tidal areas, sebkhas, marshes, and islands (such as the large Bubiyan Island). During the latter period of the flooding of the Gulf, in the mid-Holocene, much of these areas was likely to have been a shallow marine- lagoonal environment when sea level rose 1 or 2 m above its present level (Sanlaville 1989, 2000).

The Arabian coast is low-lying, with only a slight relief corresponding to the broad gentle anticlines that trend N-S or NE-SW. South of the large Bay of Kuwait and down to the Gulf of Bahrain, the coast is low and almost rectilinear, with rocky promontories, wide sandy beaches, and extensive silted swamps. The sandy point of Ras Tanurah extends SSE for about 10 km. The Dammam Peninsula is related to an Eocene limestone anticline, as is Bahrain Island, rising to 122 m at Jabal Dukhan. The Gulf of Bahrain between Bahrain and the mainland is very shallow with extensive submerged spits and bars separated by long channels, while the Bahrain archipelago is surrounded by wide, partly coralline tidal flats, which are largely exposed at low tide. Near Bu Ashira, in Bahrain, a mangrove swamp is formed of thick but not very deep mud and small mangrove bushes where the fauna consists mostly of gastropods, especially of the Potamididae and Planaxidae families (Smythe 1972). In the Qatar Peninsula, Eocene dolomite and limestone form massive and continuous outcrops forming cliffs separated from the sea by a strip of sebkhas, several kilometers wide. At the maximum of the Holocene transgression, the cliffs constituted an irregular coastline. After stabilization of the present sea level, fine sediment began to fill and regularize the embayments, creating hook-shaped spits at their southern ends and wide supratidal flats eventually formed on a complicated tidal channel system. Today, unusually strong easterly winds combine with high spring tides to flood the sebkhas as far inland as the ancient cliff. After a few weeks, evaporation of the flood water produces crusts of salt, gypsum, and dried algal mats which are progressively blown away (Shinn 1973).

Most of the beaches are composed of carbonate sand and the bulk of this material is of biological origin, made up of small fragments of shells and corals. This beach complex includes subtidal and supratidal zones. On the eastern coast of Saudi Arabia, tidal flats occupy 30–40% in space of the numerous large or small bays along the coastline. They consist of mud and very fine sand deposited in bays and sheltered zones where wave energy is low and also of rock flats composed of mud. Sands, and shell fragments cemented into beachrock known locally as *farush* (Basson et al. 1977).

The mud flats generally consist of very fine, silty sediments which can be divided into several zones:

1. The marsh grass zone with plants such as *Phragmites communis*, *Aeluropus lagopoides*, *Bienertia cycloptera*.
2. The halophyte zone covered at high tide by only a few centimeters of water. It is the uppermost portion of the true intertidal region covered with *Arthrocnemon macrostachyum* and *Halocnemon strobilaceum*, with a far-stretching root system. In the mud surface there are sometimes to be seen the round holes of the crab *Cleistoma dotilliforme*.
3. The mangrove zone with *A. marina*, with trees 1 or 2 m high, with their characteristic pneumatophores.
4. The algal mat zone composed of cemented sediment with blue-green algae and diatoms. Microscopic animals live in this mat, including gastropods, ostracods, nematods, flatworms, copepods, and oligochaete worms.
5. The *Macrophthalmus* zone between the mangrove zone and the lowtide level. These crabs (*Macrophthalmus depressus* and *Macrophthalmus grandidieri*) live in a burrow in the mud.
6. The Cerithidae zone whose density can reach 2100 individuals per square meter, with sometimes *Murex kuesterianus*.
7. Tidal creeks or drainage channels, dug by ebb tide, are a characteristic feature of the muddy tidal flats, with the presence of tide pools, about 2 or 3 m wide, 50 cm deep, and 10–30 m long, subject to great fluctuations of temperature and salinity. These tidal creeks are inhabited by a great variety of animals (shrimps, fishes, swimming crabs), which find there a shelter from the predators and rich food.
8. The sand flats frequently have a grayish color due to the presence of organic matter. At low tide, one can observe ripple marks produced by the ebbing water. Often, sands constitute a pellicular layer overlying the rock or a beach rock. The tidal channels are not strongly marked in that zone, where many crabs live, such as *Ocypode saratan*, *M. depressus*, *M. grandidieri* *Scopimera scabricauda*.
9. Rock flats also exist, for example, in the Gulf of Salwah, in bays where beach rock find exceptional conditions of formation. The *farush* consists of broken shells, sand, and mud cemented together. This type of environment presents very great biodiversity with polychaete worms, gastropods (*Thais* sp., *Monilea* sp., *Trochus* sp., *Littorina* sp.) pelecypods (*Pinctada* sp., *Isognomon* sp., *Anomia* sp., *Barabatia* sp.) and decapod Crustacea (*Alpheus* sp., *Pilumnus* sp., *Xantho* sp.), and also algae such as *Enteromorpha* sp., *Rhizoclonium kochianum*, *Achnanthes* sp.

Middle Holocene beach sediments are to be found 1 or 2 m above the present sea level, indicating a higher sea level than today. Pleistocene marine calcarenites are also known in that area where some paleo-wadis (W. Ar Rimah, W. As Sahba, and W. Ad Dawasir) have built gravel fans of enormous size.

The relatively linear character of the northern Arabian coast is modified by the Qatar Peninsula which strongly

influences the marine currents and patterns of sedimentation on the SE side of the Persian Gulf. To the east of the Qatar Peninsula there is a broad, shallow area, 10–20 m deep, studded with numerous shoals and salt dome islands. An irregular bathymetric ridge, the *Great Pearl Bank Barrier*, extends eastwards from Qatar along the central part of the United Arab Emirates coast, whose coastline is characterized mainly by low, evaporitic, supratidal flats, which locally attain 10 km in width, and by storm beaches in more exposed settings.

The western and central parts of the Trucial coast are protected laterally by Qatar Peninsula and the Great Pearl Bank barrier. The coastline is characterized by a complex of islands and peninsulas. Each island is growing by accretion around a Pleistocene rock core. Accretion also occurs as tails of sediments, which locally form tombolos. The inter-island channels end seawards in spectacular oolite deltas. Small swamps, colonized by the black mangrove *A. marina* and other halo-phytic plants, have developed, favoring the deposition of carbonate muds. The shallowness of the lagoons, together with the protection provided by the coastal barrier complex has led to a very active intertidal flat accretion which has produced the wide sabkha plain, approximately 5 km wide, during the last 4000 years. The high salinity and temperature of the water, together with the extreme aridity, have led to the extensive development of dolomite and other evaporite minerals (mainly gypsum and anhydrite) within the supratidal sediments (Evans et al. 1969; Purser and Evans 1973). As the lagoons become completely filled, aeolian coastal sands may transgress over them.

The northern part of the coast is unprotected and directly faces the entire length of the Gulf. It suffers from the effects of maximum wave fetch and a strong eastward longshore transport, which resulted in the development of storm beaches backed by coastal dunes mainly composed of skeletal carbonate sands, and the construction of major spit systems; between Sharjah and Ras al Khaimah, a series of subparallel spits has prograded the coastline some 5–10 km seaward, bringing vast lagoons into existence, the biggest being the Umm al Qowayn lagoon, which shelters vast mangroves. Further inland, the extensive quartzose dune fields of the Arabian desert extend up to the alluvial fans which skirt the Oman mountains. The NE section of the Trucial coast end in the cliffs and rocky shorelines of the Musandam Peninsula. They are mainly composed of limestone and are drained by a series of deeply incised wadis which terminate in spectacular alluvial fans, some of which reaching the coast.

The northern Persian Gulf coast is mountainous, with ridges up to 1500 m high formed by hugely folded, anticlines but the intensity of the folding diminishes markedly seawards where Pliocene folding produced regularly spaced elongate synclines and anticlines. In the south many anticlines are cut by salt diapirs. The chain of islands in the Gulf (Kharg,

Shuaib, Qeshm, etc.) forms part of the Zagros foothill belt. The Zagros is seismically active, earthquakes are quite frequent and much detrital sediment is deposited in the form of alluvial fans delivered through short, ephemeral streams perpendicular to the main structural axis.

North of Bushehr, the coast is at first swampy, then low and sandy. South of Bushehr, the Iranian shores are essentially linear and rocky, the mountains rising abruptly above the sea, with narrow coastal plains associated with the estuaries of numerous small rivers flowing down the Zagros mountains. The coastal valleys yield evidence of two distinct fills separated by a phase of incision. According to artifacts, the deposition of the older ceased by 6000 BC. The process of the younger began about 1250 years ago. Most of the streams are now incising their channels (Vita-Finzi 1978).

The Mehran River, oriented parallel to the Gulf, is situated along the axis of a syncline and has developed a big marine alluvial-fan delta, pro-grading into the Clarence Strait, between the mainland and Qeshm Island. Its very flat fan is dominated by silty and sandy deposits covered upstream by scattered vegetation of *Chenopodiaceae* and *Graminae*. Downstream there are mud banks with large desiccation polygons, the shores of the strait and of the smallest tidal creeks being populated by a dense belt of mangroves (*Avicennia*). The mud banks area, 15 km wide, includes a series of extensive, mostly bare tidal flats with a characteristic network of tidal channels (Baltzer and Purser 1990).

The Persian Gulf area is the foremost producer and exporter of oil in the world and is now experiencing huge transformations. The landscapes have been strongly altered by the creation of harbors, channels, and artificial islands, the filling of lagoons and bays, and the construction of roads and towns. Oil pollution is a great danger and the equilibrium of the sandy shores is threatened.

The Arabian Sea Coasts

Climate and Hydrology

The coastline of the Arabian Sea, along the southern coast of Arabia, the Gulf of Oman and the Iranian Makran, is very hot and arid. Mean temperature in January is $>20^{\circ}\text{C}$ (22.8°C at Salalah and 25.6°C at Aden) and the maximum absolute temperature in August is $>45^{\circ}\text{C}$ (46.7°C at Mascate and 47.2°C at Masirah). Mean annual rainfall is <100 mm (38 mm at Aden, 94 mm at Masirah) and the rainfall variability is extremely high. Winter is the main rainfall period, except on the coast of Hadramaut and Dhofar where rainfall occurs mainly in summer. Precipitation over the coastal area of Salalah is usually in the form of a fine drizzle with daily totals seldom exceeding 5 mm but the presence of fog (on average 54 days per year) has a marked effect on vegetation growth.

As a whole, the Arabian Sea is very deep with a rather narrow continental shelf. Coral reefs exist intermittently where the continental shelf is wider. The tidal range is generally between 2 and 4 m. All during the year, the coast is buffeted by the heavy Indian Ocean swell. Winds and currents are parallel to the coastline in accordance with the monsoon circulation. The NE monsoon begins in October but is mainly characteristic between November and March, with a maximum in December.

and January. Winds are moderate and the sea propitious to navigation. The SW monsoon is longer and stronger. It begins in April and is very strong from June to September. Then the sea is very rough and navigation is dangerous so the fishers do not leave the harbors. Strong southwesterly winds blow along the southern coast of the Arabian Peninsula, forcing the ocean surface in a northeasterly direction and causing a compensatory upwelling of cold subsurface water. Along the Dhofar coast, between Ras Fartak and Masirah, sea surface temperature can be up to 5 °C cooler than the ambient offshore values. These cold waters are rich in nutrients that trigger phytoplankton blooms and make for a great wealth of fish, 70% of the biological sediment is deposited during the summer monsoon months. Tropical storms and cyclones occur from time to time, almost entirely confined to May–June and October–November and cross the coast of Yemen and mainly Oman, generally between Salalah and Masirah, about once every 3 years. They occur more frequently along the Makran coast but are fewer in the Gulf of Oman. Well-developed mangrove stands are found at intervals along the Arabian Sea shore, mainly in the bays or lagoons.

Regional Features

The southern coast of Arabia is more than 2600 km long between the Bab el-Mandeb and Rass el-Hadd. It corresponds to a young rift border; its structure is very complex and the coast is often bordered with high mountains. Volcanic rocks outcrop here and there, particularly in the western part where they appear as islands and promontories, for instance in the Perim Island, at the entrance of the Red Sea, in the Aden district where extinct volcanoes are connected to the continent by tombolos, or in the Shuqra area. The Old Aden settled in a crater whose rim reaches the height of 551 m at the Jabal Shamsan. Narrow plains bordered by sandy beaches alternate with high rocky cliffs. At Ras Fartak, vertical cliffs reach up to 580 m high. About 450 km to the south, Socotra is a barren high (1519 m) island of limestone and granite, 112 km long and 15 km wide with generally steep coasts and reef-fringed promontories. In eastern Dhofar, the Kurya Muria archipelago is edged with fringing reefs. The eastern part of the southern Arabian coast is low, sandy, and barren, notably in the Wahiba Sands area, in Oman. This coastal area belongs to the Sudanian vegetation region, well differentiated from the adjacent Sahara-Arabian region by hundreds of genera and

thousands of species that are not to be found in the Holoarctic, especially different species of *Acacia*, mainly *Acacia tortilis*, *Capparis decidua*, *Calotropis procera*, *Panicum turgidum*, *Indigofera spinosa*, etc. (Zohary 1973).

The Arabian coast of the Gulf of Oman, between Ras al-Hadd and the Strait of Hormuz, in the Musendama Peninsula, sweeps in an arc, some 650 km long, dominated by an isolated folded mountain range, the Jabal al-Hajar, whose highest point rises to 3035 m. At either end of this range the mountains plunge straight into the sea and the coast is rocky and cliffy, with deep water inlets. In the center of the arc the mountains are separated from the sea by the great Batina Plain, some 280 km long and 20–30 km wide. The coast is sandy with a strip of sand dunes. Behind it, the piedmont plain is covered with silts, gravels, and conglomerate laid down in a now semi-fossilized drainage system. During the Quaternary, the deposition of wadi sediments occurred during the interglacial pluvials, whereas erosional processes predominated during the cooler aridials and numerous sea-level terraces were formed during the interglacials; traces of three coastlines dating from the last interglacial can be found and the Holocene sea level was higher than the present one (Hanns 1998).

The Makran coast is part of an accretionary wedge of late Cretaceous to Holocene sediments which accumulated near an oceanic subduction margin, between two major strike-slip fault zones, the left-lateral Chaman fault zone to the east and, to the west, the right-lateral Zendan fault, which separates the simply folded Zagros Ranges from the turbidites and ophiolitic melanges of the Inner Makran.

The Makran ranges rise irregularly from the coastal plain in a series of ridges underlain by folded sedimentary rocks. Along the coast, outcrop mainly Pliocene thick, neritic, and monotonous sequences of calcareous mudstones (the so-called *Chatti mudstones* of Pakistan). They have been locally affected by gentle folding but the beds are often sufficiently undeformed to produce extensive and almost horizontal platforms. So, the coast is marked by a series of prominent headlands (Konarak, Chah Bahar, Gavater, etc.) separated by low areas. These rocky headlands rise as isolated flat-topped hills formed from marine sandstone, conglomerate, and coquina of the Ormara and Jiwani Plio-Pleistocene formations. The Konarak terrace, one of the best developed, is an 18 km long and 300–900 m wide platform. Most of the scientists who investigated this area were strongly interested in these platforms which exhibit spectacular stepped shore platforms and raised beaches (Page et al. 1979; Snead 1993; Reyss et al. 1998). On Qeshm Island as many as 18 marine terraces were identified, up to 220 m in elevation, and as many as 19 levels, up to 246 m, near Chah Bahar. The lowest levels, less than 4 m high, were 14C dated as Holocene; for higher levels, up to 26 m in elevation. Uranium series analyses gave apparent ages between 100 and 140 ka BP, and therefore may be ascribed to oxygen-isotopic stage 5e (or 5c). These dates

show that uplift rates did not exceed 0.2 mm/year during the late Quaternary, except in areas where salt domes exist, for example, on Qeshm Island where uplift rate was faster (6 mm/year). Deposits of the oldest levels are not suitable for dating because the shells and corals are very recrystallized. These marine terraces give evidence of the complex interrelationships of tectonism, coastal erosion and sedimentation, and eustatic sea-level changes (Reyss et al. 1998).

However, the main features of the coastal area could be the low-lying alluvial plains forming a coastal strip, 5 to more than 20 km wide, between the coastline and the mountains (Page et al. 1979). Several stepped terraces partly rocky and partly alluvial exist at the foot of the Makran ranges. The streams have eroded the siltstone and mudstone bedrock into low-lying pediments and built huge deltas. Locally, the plain is covered with sand dunes. Typical mud volcanoes outcrop as well in Makran as on the coastal plain of the Persian Gulf. In the coastal plain there are many shore deposits, series of prograding accretion beach ridges, isolating lagoons and marshes which shelter vast mangroves, before being filled by silt. In some places, ancient islands have been connected to the mainland by one or two spits which finally form a tombolo (Ras Tang, Konarak). So, the 19.3 km-long Gurdin terrace paralleling the coastline is connected with the coastal plain by a wide 12.8 km tombolo, displaying well-developed beach ridges. The Makran coast is unprotected and during the SW monsoon and especially in tropical storms and cyclones, the ocean and surf are extremely rough. Erosion continues to affect the headlands, but elsewhere progradation seems to have dominated along the flat sandy areas since the maximum of the postglacial transgression. Radiocarbon dates for shell samples collected in the central part of the tombolo of Konarak and elsewhere on the ancient beaches of the Makran coast gave ages between 4870 ± 100 year. BP and 6255 ± 320 year. BP (Page et al. 1979; Vita-Finzi 1981). The same observations were made on beach deposits of the Pakistan Makran (Sanlaville et al. 1991). So, in Makran the coastline has been prograding since the Middle Holocene owing to both slight uplift and marine or alluvial sedimentation.

Cross-References

- Artificial Islands
- Beachrock
- Climate Patterns in the Coastal Zone
- Coastal Climate
- Coral Reef Coasts
- Coral Reefs
- Desert Coasts
- Mangroves, Ecology
- Oil Spills
- Paleocoastlines

- Sharm Coasts
- Tidal Creeks
- Trottoirs

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Asteroid-Impact Coasts

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A highly unusual category of general coastal morphology is created by asteroid impacts at some time in the geologic past. Asteroid impact creates a circular or ovoid crater beneath which is a brecciated zone that extends many thousand meters below the former surface of the Earth's crust. The eroded relics of these ancient craters have been called "*astrob-lemes*" (Dietz 1961).

Three coastal areas in North America are believed to owe their morphology, at least in part, to asteroid impact. They are:

1. *Chesapeake Bay, and adjacent areas of Maryland and Virginia*. The coast is characterized by an unusual pattern of a drowned dendritic drainage system, that is to say, organized like the branches of a well-shaped tree. It is fed by the valleys of the Susquehanna, Potomac, and Rappahanock rivers. The asteroid or "bolide" struck 35.2 (+/–0.3) million years ago, in Late Eocene times in soft coastal plain and shelf sediments which to the impacting object, about 3–5 km in diameter and travelling at about 80,000 km/h, would have had the consistency of warm butter. Molten glass shards (cooled as tektites) were strewn over 9 million km². Penetrating about 600 m of youthful sediments it continued to a limited depth in the crystalline basement. The principal crater is about 85 km in diameter, and overlying (postimpact) sediments have slowly sagged into the depression and the surrounding area. In the center of this astrobleme, there is an inner depression with a peak in the center (just as in some other impact sites such as the Miocene Ries crater of Germany). A much larger area extending up to 90 km from the principal crater rim is marked by a belt of radial fractures spreading out over much of the Salisbury Embayment. A polymictous debris

Rhodes W. Fairbridge: deceased.

plume reached NE, up the Atlantic coast at least 400 km. The subsurface structures have been proven by extensive seismic profiling and drilling (Poag 1997). In the 35 Myr since the impact, the site has been the focus of differential compaction and subsidence (with eustatic revivals) that determines the general form of today's coastline. Another but much smaller asteroid fragment at the same time created the 25-km-wide Toms Canyon crater 140 km east of Atlantic City.

2. *Gulf of Mexico (Chicxulub crater)*. This impact occurred near the west side of the Yucatan peninsula of Mexico about 65.2 (+/−0.4) million years ago, generating a crater 180 km across. Its ejected debris has been traced to a semicircular rim extending in an arc through central Texas which suggests an incoming orbit heading NNW. This event was destined to become the Cretaceous/Tertiary boundary, marking a revolution involving major biotic extinctions, notably the dinosaurs, ammonites, and other classes (75% of all marine species). While the western and northern borders of the Gulf of Mexico had already developed as a broad arc in Mesozoic times, its southern border is marked by a meridional fault system that today truncates the western borders of Yucatan, a nearly horizontal platform of Miocene limestones. The impact theory, initiated by W. and L. Alvarez in the late 1970s (from evidence in Italy) triggered an intense interest that has been well reviewed by Marvin (1990) and expanded to events throughout the entire Phanerozoic (Rampino and Haggerty 1996). It seems likely that anomalous patterns of coastlines worldwide should be explored for possible astrobleme ancestry.
3. *Hudson Bay*. The present form of this embayment is an epicontinental sea bounded particularly in the southeast by an extraordinary arcuate coast that coincides with an Archaean/ Proterozoic boundary injected by lopolithic sills of diabase. An ancient astrobleme is suggested by Dietz (1961).

Cross-References

- [Changing Sea Levels](#)
- [Estuaries](#)
- [Sea-Level and Climate Change](#)
- [Submerging Coasts](#)

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Atlantic Ocean Islands, Coastal Ecology

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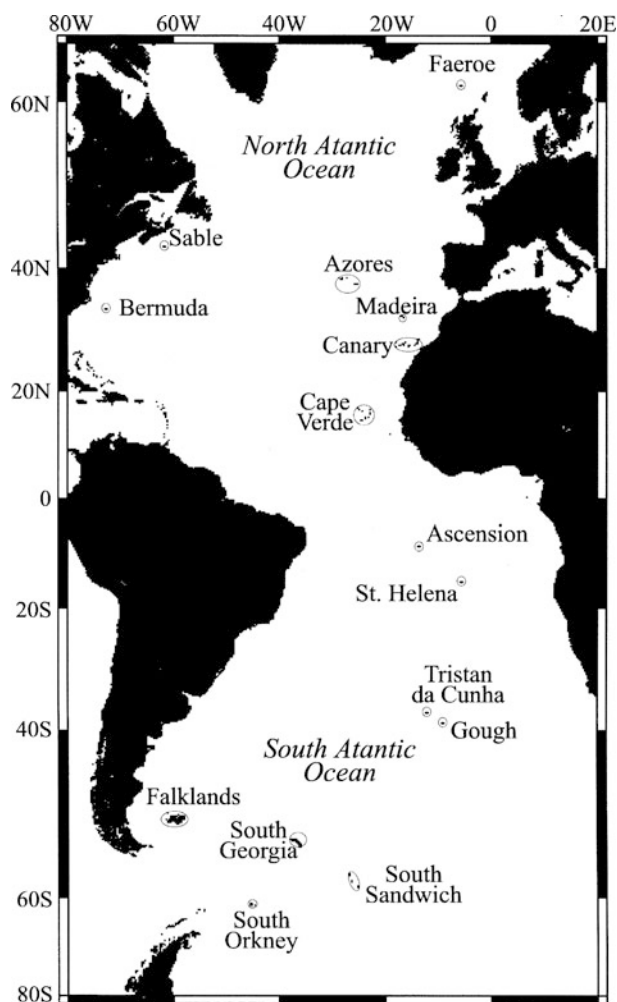
While many islands lie in Atlantic waters, this entry will focus only on a few of the most significant to illustrate a variety of insular habitats from the Northern and Southern Hemispheres (Fig. 1). Except for the Faeroe, Azores, Canaries, Bermudas, Ascension, Gough, and the islands of the Scotia Arc, relatively little is known about the coastal ecology of most of the islands scattered in the Atlantic. When available, data on the general biodiversity, the intertidal and subtidal zones, and the seasonal variation of the plankton will be presented, together with some aspects of the terrestrial ecology relevant to the coast and typical of islands.

The fauna and flora of most islands have several features that distinguish them from those of continents. Many of these are related to difficulties of dispersal. Organisms that can disperse well are more likely to be found on islands than those that cannot. Species of plants and animals on islands show equilibrium between new arrivals and local extinctions. Generally, the number of species is greater on large islands than small islands, varying in function of their isolation from other landmasses (Table 1). Thus, islands usually present an impoverished terrestrial and marine wildlife compared to equal areas of mainland (Williamson 1981).

Faeroe (Denmark)

The Faeroe group comprises 1398 km² of rugged and mountainous islands located just north of the United Kingdom, on the edge of the Rockall Rise. The coastline is indented, with fjord-like inlets, the marine life is abundant and the climate is mainly temperate oceanic.

As their situation in the transitional zone between the North Atlantic and the Arctic, the Faeroe Islands harbor a flora and fauna made up of both boreal and arctic species (Blindheim 1989). The phytoplankton productivity of this area is considered to be one of the greatest in the world oceans. The blooms occur in March and consist mainly of diatoms (Braarud et al. 1958). Compared to the spring bloom, the summer abundance of pelagic algae is rather poor, but more diverse (Blindheim 1989). The zooplankton bloom begins around March–April, in conjunction with the



Atlantic Ocean Islands, Coastal Ecology, Fig. 1 Map of the Atlantic Ocean locating the islands discussed in this entry

Atlantic Ocean Islands, Coastal Ecology, Table 1 Isolation index of different islands or group of islands in the Atlantic Ocean (compiled from the United Nations Environment Programme)

Islands	Isolation index ^a
Bermuda	91
Azores	74–77
Madeira	66
Canary	12–41
St. Helena	113
Ascension	119
Tristan da Cunha	78
Gough	125
Falklands	59
South Georgia	113
South Sandwich	71–76
South Orkney	28–31

^aThe isolation index is the sum of the square roots of the distances to the nearest equivalent or larger island, the nearest island group or archipelago and the nearest continent. A range indicates the isolation variation among the different islands composing an archipelago or group of islands

phytoplanktonic bloom. The summer pelagic fauna is dominated by copepods such as *Calanus finmarchicus*, *Pseudocalanus minutus*, and also by the arrow worms *Sagitta elegans* and *Eukrohnia hamata*. The Euphausiacea represent another important component of the zooplankton. Common pelagic fish in the waters off Faeroe are herrings, Greenland halibuts, blue whittings, redfish, cods, and haddocks.

On the vertical slopes of the coast, from the lowest water spring tide mark, every centimeter of the subtidal benthic zone is covered in a profuse carpet of life. Kelp and other algae are gradually replaced by sea anemones and sponges in deeper water.

Marine birds like the puffin *Fatercula arctica* and marine mammals such as the pilot whale *Globicephala melaena* and the white-beaked dolphin *Lagenorhynchus albirostris* are abundant.

Sable (Canada)

Sable Island is located approximately 300 km southeast of the Canadian province of Nova-Scotia. The island itself is about 40 km long and 1 km wide, hence its temperate climate is markedly influenced by the sea. Ocean waters maintain winter air near freezing and keep summer temperatures below what would be considered room temperature. Fierce ocean currents sweep around the island causing the peculiar shifting of sand that constantly changes its contours. Between the dunes are numerous depressions usually filled with freshwater, and supporting a variety of aquatic plants and insects. A total of 179 species of terrestrial plants can be found on Sable Island. Along the coast, the moving sand is stabilized by the marram *Ammophila breviligulata*, the sandwort *Honckenya peploides*, and the beach pea *Lathyrus maritimus*.

Since all the shores of the island are sandy, benthic invertebrate diversity is low, although the sand dollar *Echinarachnius parma*, shrimp *Crangon septemspinosa*, and gastropod *Lunatia heros* are observed commonly. Among fish larvae, the pollock *Pollachius virens* and haddock *Melanogrammus aeglefinus* have been found around Sable Island. The American sand lace, smooth skate, yellow-tail flounder, Atlantic cod, Atlantic mackerel, witch flounder, and spiny dogfish are also present.

Sable Island is the only known nesting place for the ipwich sparrow *Passerculus sandwichensis princeps*. Over 2500 pairs of terns, mostly Arctic terns *Sterna paradisaea*, nest on the island along with over 500 pairs of black-backed gulls *Larus marinus* and 2000 pairs of herring gulls *Larus argentatus*. A few sandpipers *Calidris* spp. and semipalmated plovers *Charadrius semipalmatus* also nest on the island and broods of black ducks *Anas rubripes* and red-breasted mergansers *Mergus sarrator* are produced near the ponds. During

migration and after strong gales, nearly any species of bird in eastern North America may show up on Sable Island. About 300 Sable Island horses, introduced in 1738, are now living wild on the island. Sable Island also supports the largest breeding colonies of gray seals *Halichoerus grypus* and harbor seals *Phoca vitulina* in the western Atlantic Ocean. Harbor seals use Sable Island as a mating ground (Coltman et al. 1999).

Not far from Sable Island is the Sable Gully, which is the largest marine canyon in eastern Canada, and is considered an important habitat for a wide diversity of marine life, including over 200 vulnerable northern bottlenose whales (*Hyperoodon ampullatus*).

Bermuda

The island group of Bermuda is located in the western central region of the North Atlantic oceanic gyre known as the Sargasso Sea, approximately 1250 km southeast of New York City. Bermuda comprises about 130 islands and islets covering a land of just over 50 km². The climate is wet subtropical. These islands are characterized by tidal and non-tidal ponds, rocky coasts, sandy beaches, and shallow marine bays. They harbor the most northerly mangroves in the world and the most northerly coral reefs in the Atlantic, including coral/algal and algal/vermetid reefs.

The ocean temperature of the surface waters off Bermuda varies seasonally between 18 °C and 28 °C. This island group can be considered a semi-tropical area where ocean conditions differ markedly from those encountered further south in the true tropics. In Bermuda, winter cooling and mixing of the surface layers occurs and a seasonal thermocline develops from late spring through fall (Morris et al. 1977). The phytoplankton of the Sargasso Sea that surrounds Bermuda shows low concentrations, especially during spring and summer with about 3000 cells L⁻¹. The density of phytoplankton reaches a maximum in fall and winter with 50,000 cells L⁻¹. In summer, the phytoplankton is mainly composed of coccolithophores, dinoflagellates, and diatoms. The main species, especially between late fall and spring, is the coccolithophore *Coccolithus huxleyi*. A spring bloom of diatoms, 200,000 cells L⁻¹, occurs around April. Dinoflagellates dominate in late summer concurrent with the lowest productivity of the year.

The zooplankton of the waters surrounding the Bermuda reef platform and of the inshore waters are distinct from the oceanic species. The larvae of numerous benthic invertebrates are a significant component of the reef and inshore zooplankton, which is 5 to 10 times more abundant than in the Sargasso Sea, fluctuating around 2000 individuals m⁻³. Maximum zooplankton abundance generally occurs after the phytoplanktonic bloom in late spring-summer and

again in winter, following the spring and fall increases of phytoplankton. Both zooplankton peaks are of only short duration and are followed by rapid population declines. Copepods are the most abundant members of the inshore populations and dominate in summer and fall, representing about 70–95% of all species present. In winter, the appendicularian *Oikopleura graciloides* dominate and fish eggs are prevalent in spring. Copepods nauplii and gastropod veligers are abundant in spring and again in early winter. Though less abundant, medusae, cladocerans, decapod and fish larvae are also part of the zooplankton of Bermuda (Morris et al. 1977).

The benthic habitats of Bermuda comprise rocky shores of the intertidal and subtidal areas, gullies, crevices and caves, muddy bottoms, mangroves, sandy-bottoms, and coral reefs. The coral reefs of Bermuda can further be divided into four different types: the ledge flat reef, the patch reefs, the coral knobs, and boiler reefs (Morris et al. 1977). The muddy substrata, mainly located in shallow protected areas, are associated with mangroves but also occur in the deepest bassin depressions. The sublittoral zones around Bermuda, including the surroundings of coral reefs, are mostly sandy. Numerous algae and sea grasses cover the shallows and offer home or shelter to various invertebrates. Burrowing species are the most abundant group of invertebrates present.

The rocky shore, sublittoral, and coral reef habitats shelter the greatest number of species. Overall, 118 common species are known from the rocky-shore subtidal area, including several algae, cnidarians, mollusks, and echinoderms. The rocky intertidal zone harbors 82 common species, dominated by algae, mollusks, and crustaceans. Many of these species live in the lower fringe or directly in tide pools. In the coral reefs, there are about 79 common species, mainly corals, but also algae, sponges, mollusks, and crustaceans. Mangrove swamps sustain about 21 species, dominated by algae, shrubs, mollusks, crustaceans, and bryozoans (Morris et al. 1977).

Macronesia

Macronesia encompasses the islands of Azores, Madeira, Canaries, and Cape Verde, which are located in the northeastern Atlantic.

The Azores consist of 10 Portuguese islands covering 2244 km² and having a highly volcanic topography with many peaks and hollows (Bird 1985). They lie about 1500 km off the main coast of Europe and their climate is moist-moderate with abundant rainfall.

Currently, 274 species of algae are listed in the Azores, including 45 Chlorophyta, 52 Phaeophyta, and 177 Rhodophyta, among which are 10 endemic species. Although the Azores

share algal biotopes with the Atlantic coast of mainland Europe, intertidal and subtidal seashores in the archipelago mostly lack the functionally important community of fucoids and laminarians widespread in the North Atlantic. Large brown algae such as *Fucus spiralis* are sporadic in the Azores, while *Laminaria ochroleuca* is known from only one location. Other large brown algae (*Cystoseira* spp., *Sargassum* spp.) occur in deep sheltered pools, lagoons, and some subtidal biotopes.

The intertidal areas of the Azores typically present three distinct zones. The highest band is dominated by littorinids (*Littorina striata*, *Melaraphe neritoides*). The upper boundary of the second zone is marked by the presence of barnacles *Chthamalus stellatus* and some small shoots of *Ulva rigida*, *Gelidium pusillum*, *Gelidium microdon*, *Ralfsia verrucosa*, and *Enteromorpha* spp. The third zone comprises an extensive area dominated by *G. pusillum* and *Centroceras capillacea*. In the adjacent subtidal area, the algal community changes along the depth gradient. *Ulva* spp., *Hypnea musciformis*, and *Asparagopsis armata* dominate the shallow depths. The 10 m depth community is characterized by the algae *Dictyota dichotoma*, *Halopteris* spp., and *Stypocaulon scoparia*, while *Zonaria tournefortii* is dominant at 30 m depth. With the exception of mollusks *Bittium* spp., that can be found everywhere, differences in the invertebrate fauna are also observed along the depth gradient. Sponges are an example of this: *Tethya aurantium*, *Halichondria panicea*, *Hymeniacidon sanguinea*, and *Timea unistellata* occur in the intertidal zone, whereas *Petrosia ficiformis*, *Erylus discophorus*, *Myxilla macrosigma*, and *Clathrina coriacea* are found between 10 and 15 m depth (Neto and Avila 1999).

The southern oyster drill *Stramonita (Thais) haemastoma* preys intensively on the populations of *Mytilus edulis*, *Patella candei*, *Patella aspera*, *Crassostrea virginica*, *Ostrea equetris*, *Ischadium recurvum*, *Balanus eburneus*, *Balanus amphitrite*, and *Chthamalus stellatus* which colonize the intertidal zone of the Azores. Sea urchins *Arbacia lixula* also common on the rocky shores.

Over 175 species of mesopelagic fish are believed to occur around the Azores. The most widely spread families are included in the Stomiidae, Gonostomatidae, Sternoptychidae, and Paralepididae. The yellowmouth barracuda *Sphyræna viridensis* is a common pelagic coastal predator, mostly preying on juvenile horse mackerels *Trachurus picturatus*.

Large numbers of juvenile loggerhead sea turtles *Caretta caretta* thrive off the coast of the Azores. The sperm whale *Physeter macrocephalus* is the most common whale in Azorian waters. A total of 23 species of cetaceans have been observed, including the false killer whale *Pseudorca crassidens*, Risso dolphin *Grampus griseus*, bottlenose dolphin *Tursiops truncatus*, common dolphin *Delphinus delphis*, striped dolphin *Stenella coeruleoalba* and spotted dolphin *Stenella frontalis*. *Balaneoptera acurostrata* and

Globicephala macrorhynchus have also been recorded in winter and early spring.

The Portuguese archipelago of Madeira consists of one large and several small islands lying some 600 km off the coast of Morocco and representing a total land area of 810 km², mostly made up of extinct volcanoes. The climate is semi-arid subtropical. The slopes are steep, but well vegetated and terminated in high coastal cliffs. Beaches are mainly gravely, but beach borders are also found. The coastal marine fauna of Madeira is a mixture of species with Mediterranean affinities and distinctly tropical species that often reach their northern limit there (Wirtz 1999).

Little is known of the precise marine ecology of this group. The sea urchin *Diadema antillarum* is an extensive herbivore that preys on algal bed. Thirty-eight species of fish are easily observed in the coastal waters of Madeira. The endangered Madeira soft-plumaged petrel *Pterodroma mollis madeira* occurs there, as well as the Mediterranean monk seal *Monachus monachus*, which breeds on the island. In fact, Madeira is one of the most important remaining localities for this species threatened with extinction.

The Canary group comprises 7500 km² of dry islands spreading close to the coast of northwest Africa. These Spanish islands are mostly surrounded by sandy beaches and rocky shores and their climate is semi-arid subtropical.

In the Canary Islands, the chlorophyll concentrations range from 0.09 mg m⁻³ to 0.25 mg m⁻³, with deep maxima between 75 and 100 m depth. The biomass in the shallower water corresponds to 2000–28,000 cells L near the coast (Ojeda 1996). Some coastal lagoons are characterized by the conspicuous presence of Cyanophyta and Chlorophyta up to 10⁵ cells ml⁻¹, thus imparting a green coloration to the water.

Primary productivity is high in the water off Canary Islands. The northern portion of the islands supports a high phytoplankton production where the Canary current collides with the coast. The biomass is not large but the turnover is pronounced. Bacteria associated with phytoplankton are abundant at the outer edge of the main phytoplankton concentration. Zooplankton accumulates in the marginal southern areas and is quite important in the upwelling area in the upper 50 m of the water column. The zooplankton community is generally composed of copepods, but mysidacea are usually important in all regions. Zooplankton also includes amphipods, ostracods, and mysids.

The sea urchin *Diadema antillarum* is commonly observed. Falcon et al. (1996) found 76 species of coastal fish around the Canary Islands, the most common being *Abudefduf luridus*, *Canthigaster rostrata*, *Chromis limbatus*, *Sparisoma cretense*, and *Thalassoma pavo*. Lorenzo (1995) indicated that many waders (*Charadrius alexandrinus*, *Calidris alba*, *Arenaria interpres*) were found in intertidal lava platforms and in supratidal lagoons of the Canary Islands.

St. Helena and Ascension (United Kingdom)

St. Helena is an isolated volcanic peak rising 4400 m from the seafloor, slightly east of the mid-Atlantic ridge, about 1900 km off the coast of Angola. It covers 125 km² as a deeply eroded volcanic cone. It has a dry subtropical climate. The bay on the south coast has a beach backed by calcareous dunes, which have been driven far inland by the southeast trade winds. The coastal areas are rugged and barren whereas the higher elevations in the center of the island present a lush vegetation (Bird 1985).

St. Helena's isolated position in the South Atlantic has given rise to an unusual and remarkable land and marine flora and fauna. There are 60 species of flowering plants, of which 49 are endemic. Thirteen endemic species of ferns are present, almost all on the verge of extinction. Of 1100 land invertebrates, 400 are unique to St. Helena. At least six unique land birds once occurred on St. Helena, only one, the wirebird *Charadrius sanctae helenae*, survives today. Also, modest seabird populations nest on isolated cliffs and rocks. The island is home to the largest species of earwig, *Labidura herculeana*. Endemic species of shrimps *Typhlatya rogersi* and *Procaris ascensionis* are present in rock pools. Ten shore fishes are found only around the island, and 16 more are found only here and at Ascension. The green turtle *Chelonia mydas*, and the hawksbill turtle *Eretomochelys imbricata* may breed along the coast of St. Helena.

Ascension is a 97 km² volcanic island isolated in the south Atlantic. The peak of a dormant volcano rises 3000 m above the seafloor and a further 859 m above sea level to the summit of Green Mountain. The vegetation is sparse with scattered guano deposits and the island has a dry tropical climate. Imposing sea cliffs line the more exposed eastern coast (Bird 1985). Beaches, intertidal rock pools, and rocky shores are mainly composed of bare rock, regularly covered by a layer of crustose coralline algae that has cemented all rubble into a uniform surface.

The general biodiversity of Ascension is not high, not only because of its isolation but also due to the absence of lagoons, estuaries, seagrass beds, mangroves, and coral reefs (Manning and Fenner 1990). Price and John (1978) concluded that the primary facet presented by the intertidal and sublittoral fringe on Ascension was still that of stark bare rock, sometimes abutting or adjoining clean sand. Other factors such as strong tidal surge, supplemented by rollers, enormous waves that assault the island, influence the shore and shallow depth colonization. Finally, the blackfish *Melichthys niger* is another important factor regulating the presence of other species, as it deters colonization by grazing all surfaces that it can reach (Day 1983), including the crustose algae described earlier. Stands of algae and colonies of sabellariid worms, barnacles, and solitary coral are limited to crevices, blowholes, and isolated tide pools where blackfish cannot penetrate (Manning and Fenner 1990).

The sandy beaches of Ascension are high-energy areas made up of very coarse sand. Only digging crustaceans *Hippa testudinaria* have been found in the sand and crabs *Grapsus* spp. have been seen roaming over the open beaches at night. Isolated tide pools with rims of protective rock are often inhabited by dense beds of oysters *Saccostrea cucullata* (Manning and Fenner 1990).

Much of Ascension's global conservation importance comes from the island's remoteness, which has produced one of the most remarkable island floras and faunas in the world. It is of world significance for its 11 species of breeding seabirds, especially the unique Ascension Island frigate bird *Fregata aquila*. There are six unique species of land plants, nine of marine fish and shellfish, and over twenty of land invertebrates. Some 27 species of amphipods have been described in the pool habitats along the shore, and Manning and Fenner (1990) observed 74 species of decapods on Ascension. Of these, 41 occur in the western Atlantic, and most are common.

The echinoderm fauna of Ascension comprises 25 species, including brittle stars, sea cucumbers, sea urchins, and starfish. Eight of these species are amphi-Atlantic, one species is restricted to Ascension, three species are also known in the western Atlantic, four species are also found in eastern Atlantic, five species are circumtropical and four species are also found in St. Helena. These findings suggest that species dispersion by larvae is mediated by surface and subsurface transport of planktonic stages (Pawson 1978).

About 81 species of shore fishes are known, including 11 endemic species. Ascension has one of the most important breeding populations of green turtle *C. mydas* in the world. The green turtles mate and lay their eggs on sandy beaches, whereas hawksbill turtles *E. imbricata* less frequently seen along the shores.

Tristan da Cunha and Gough (United Kingdom)

Tristan da Cunha is a circular volcanic island of 103 km² rising 2062 m above sea level. The island is bleak and barren, but zoned vegetation occurs above 1350 m. It has a wet temperate climate and is surrounded by rocky shores.

Gough is a deeply dissected volcanic island located 350 km southeast of Tristan da Cunha. High cliffs with hanging valleys and waterfalls surround the surface area of 65 km². The luxuriant evergreen scrub forest thins out above the 300 m contour. The climate is a cool-temperate oceanic. The flora of Gough is typical of southern cold-temperate oceanic islands, with relatively low species diversity, and a large preponderance of ferns and cryptogams.

Both Tristan da Cunha and Gough are characterized by their isolation and high level of endemism near the shore. However, with increasing depth, the amount of faunal similarity increases between regions across the South Atlantic.

The marine area can be split into two distinct algal zones. From sea level to 5 m depth, algae consist mainly of bull kelp *Durvillea Antarctica*, and beyond 20 m are dominated by *Laminaria pallida* and giant kelp *Macrocystis pyrifera*. Forty species of algae have been recorded, of which two species are endemic to Gough. Most littoral species found at Gough are widespread on other southern ocean islands, and 79 invertebrate species have been recorded. The absence of limpets and bivalves in the littoral and subtidal zones is noted. Sea urchins *Arbacia dufronii* are abundant in the marine area, as are whelks *Argobuccinum* sp., chitons, starfishes, sea anemones, bryozoans, barnacles, slipper limpets, nudibranchs and sponges. Important marine species include the Tristan rock lobster *Jasus tristani* (from close inshore to 400 m depth around Gough), and octopi. Both are economically exploited by fishermen under close regulation. About 51 species of fish are known to occur in nearshore waters of Tristan da Cunha and Gough.

Gough has been described as a strong contender for the title of “most important seabird colony in the world,” with 54 species recorded in total, of which 22 breed on the island and 20 are seabirds. About 48% of the world’s population of northern rockhopper penguin *Eudyptes chrysocome moseleyi* breed at Gough. Atlantic petrel *Pterodroma incerta* is endemic to Gough and the Tristan group of islands. Gough also harbors a major colony of the great shearwater *Puffinus gravis* with up to three million pairs breeding on the island. The main southern ocean breeding sites of little shearwater *Puffinus assimilis* are Tristan da Cunha and Gough Island, with breeding pairs numbering several million. The wandering albatross *Diomedea exulans dabbenena* is virtually restricted to Gough, with up to 2000 breeding pairs. The only survivors of southern giant petrel *Macronectes giganteus* also breed on Gough, with an estimated 100–150 pairs. Gough moorhen *Gallinula comeri* is found in fern bush vegetation areas, and estimates of population size vary from 300–500 pairs to 2000–3000 pairs. About 200 pairs of Gough finch *Rowettia goughensis* have been recorded, although recent estimates suggest that there may be 1000 pairs. Other seabirds breed on these islands, such as the yellow-nosed albatross *Diomedea chlororhynchos chlororhynchos* and the great shearwater *P. gravis*. The penguin *Eudyptes crestatata* is present on Tristan.

Subantarctic fur seals *Arctocephalus tropicalis* (200,000 individuals and increasing), and southern elephant seals *Mirounga leonina* (about 100 individuals) are the only two native breeding mammals. The former breed at beaches all round the island, whereas the latter are restricted to the island’s sheltered east coast. Two other marine mammals are found in the area, namely the southern right whale *Eubalaena glacialis australis* and the dusky dolphin *Lagenorhynchus obscurus*. Reptiles, amphibians,

freshwater fish, and native terrestrial mammals are absent, although the introduced house mouse *Mus musculus* is widespread and abundant.

Falklands (United Kingdom)

The Falklands include two main islands and several hundreds smaller islands and islets that cover about 12,000 km². The coastline is lengthy and highly indented with numerous rias at the mouths of winding valleys. The landscape is treeless, with tussocky grasses and extensive peat bog mantles (Bird 1985). Some rocky coasts possess offshore kelp beds of *Macrocystis*. The climate is dry cool temperate.

The Falklands are exceptionally rich in marine life, including benthic and pelagic forms. Also, about 85% of the world population of the black-browed albatross (*Diomedea melanophris*), and the largest concentration of rockhopper penguins (*E. chrysocome*) are found on the islands. A total of 63 species of sea and land birds occur, including the night heron *Nycticorax nycticorax cyanocephalus*, ashy-headed goose *Chloephaga poliocephala*, kelp goose *Chloephaga hybrida malvinarum*, Garnot’s ground tyrant *Muscisaxicola muscisaxicola macloviana*, and the common diving petrel *Pelecanoides urinatrix berard*. The penguins *Pygoscelis papua*, *E. crestatata*, *Eudyptes chrysolophus*, and *Spheniscus magellanicus* are present on the Falklands. The islands are breeding grounds for sea lions (*Otaria flavescens*), elephant seals (*Mirounga leonina*) and fur seals (*Arctocephalus gazella*). Fifteen species of whales and dolphins occur in the surrounding seas. Pelagic invertebrates such as the squid *Ilex argentinus* are abundant near the coast and can be considered the basic diet of birds and mammals.

Scotia Arc Islands (United Kingdom)

South Georgia, South Sandwich, and South Orkney Islands combined represent about 10,000 km² of land that peaks over 2000 m. They are known as the islands of the Scotia Arc. Most of them rise steeply from the sea and are rugged and mountainous. Their coastlines are generally cliffed and rocky, with many inlets and some bay beaches.

South Georgia is a remote island located 1330 km from any other land. It is largely barren and has steep, glacier-covered mountains. The climate is wet subpolar; it is variable, with mostly westerly winds throughout the year, interspersed with periods of calm; nearly all precipitation falls as snow. The South Georgia fauna is depauperated and consists mainly of direct developers. The intertidal fauna is of geologically recent origin. It is suggested that South Georgia shores stem from colonization by rafting from remote sources. There are

abundant rocky shores with *Macrosystis* kelp offshore. Some 24 species of breeding seabirds and 26 visitors, 6 species of seals, and 10 endemic bottom dwelling fish are found around South Georgia. Penguins *Aptenodytes patagonicus*, *E. chrysolophus*, *Pygocelis antarctica* are present.

The South Sandwich Islands are of volcanic origin with some active volcanoes. They have a wet polar climate, and sparse vegetation of lower plants (lichens and mosses), richer around fumaroles with dense bryophyte communities. The penguins *A. patagonicus*, *E. chrysolophus*, *Pygocelis papua*, and *Pygocelis adeliae* are present.

The South Orkney Islands are located between the Antarctic Peninsula and South Georgia and they have a subpolar climate. The littoral zone is either lifeless or very poor. Scarce diatoms and a few species of chitons and gastropods are found along the littoral. At low tide, there are green and red algae in small rock pools, and amphipods and planarians under stones. At around 2 or 3 m depth, green and red algae appear with an increasing number of fauna including starfish, urchins, sponges, and ascidians. At around 8–10 m depth, new species of starfish appear and the general animal biomass increases with depth. At 30 m, ascidians, sea urchins, starfish, and brittle stars form vast colonies. The penguins *P. antarctica*, *P. papua* and *P. adeliae* are present around South Orkney.

Overall, the biota of the intertidal region is poor, even during summer, in the Scotia Arc Islands. Due to the rigorous conditions, most biota is confined to lower levels of the shore (Stephenson and Stephenson 1972). The midlittoral zone that is most exposed to air is characterized by the presence of the *Porphyra* algae followed by *Ulothrix* and *Urospora* along the center of the zone. There are also algae in the tide pools and crevices. Limpets *Patinigera polaris* occur in crevices, whereas amphipods, nemertines, and flatworms are found clustered under stones and rocks. The tide pools of the midlittoral are home to many Antarctic species of algae such as *Leptosomia*, *Iridaea*, *Adenocytis* that can also extend their distribution to the infralittoral fringe area with *Desmarestia*, *Curdia*, *Monostroma*, and *Plocamium*. The infralittoral fringe is characterized by numerous algae, mainly of the genera *Desmarestia* and *Ascosiera*. The huge *Phyllogigas grandifolius* dominates this area and its blades harbor a small fauna comprised, among others, of worms, sponges, hydroids, and mollusks (Stephenson and Stephenson 1972).

There is no supralittoral fringe in many places, the midlittoral fringe has a very restricted population but there is an increasing number of species in the infralittoral zone (Stephenson and Stephenson 1972). Several animals such as sponges, alcyonarians, pycnogonids, amphipods, isopods, polychaetes, and nemertine worms are giant representative of their group (Knox 1970).

The chlorophyll abundance in the surface waters shows a high degree of variability. The chlorophyll in South Orkney

fluctuates around 4 mg m^{-3} in summer. High pigment concentrations have been found especially between South Orkney and the South Sandwich Islands. The main species that compose the phytoplankton biomass are diatoms. *Chaetoceros neglectus*, *Chaetoceros dictyota*, *Chaetoceros atlanticus*, *Nitzschia* sp., *Corethron criophilum* are the most common reaching densities around 10^5 cells L (Zernova 1970). The standing crop of zooplankton in the subantarctic represents 55 g m^{-3} between 0 and 50 m depth. Around South Georgia, the relationship between the density of phytoplankton and zooplankton is inverse (Knox 1970). Copepods reach maximum densities at around 600 m depth, decreasing continuously toward the surface. The main species of zooplankton are *Euphausia superba*, *Euphausia frigida*, *Thysanoessa* spp., *Parathemisto* spp., and *Salpa fusiformis*. The main krill biomass is observed in areas of the island arc. The complicated underwater relief and coastline of this area leads to the formation of numerous gyres and local eddy currents trapping the krill (Makarov et al. 1970).

Krill has been found to be an important part of the diet of numerous fish living around the islands, including members of the families Rajidae, Paralepididae, Myctophidae, Scopelarchidae, Muraenolepididae, Gadidae, Moridae, Macruridae. The occurrence of various fish in the epipelagic waters of the Scotia Arc Islands, which are described by many authors as a highly productive krill zone, are also profitable to whales, seals, birds, and Antarctic fishes, but also to subantarctic and subtropical epipelagic species of the southern hemisphere. The crabeater, leopard, Ross and some other seals can be found in the Scotia Arc Islands.

Cross-References

- ▶ [Antarctica, Coastal Ecology and Geomorphology](#)
- ▶ [Arctic, Coastal Ecology](#)
- ▶ [Atlantic Ocean Islands, Coastal Geomorphology](#)
- ▶ [Caribbean Islands, Coastal Ecology and Geomorphology](#)

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Atlantic Ocean Islands, Coastal Geomorphology

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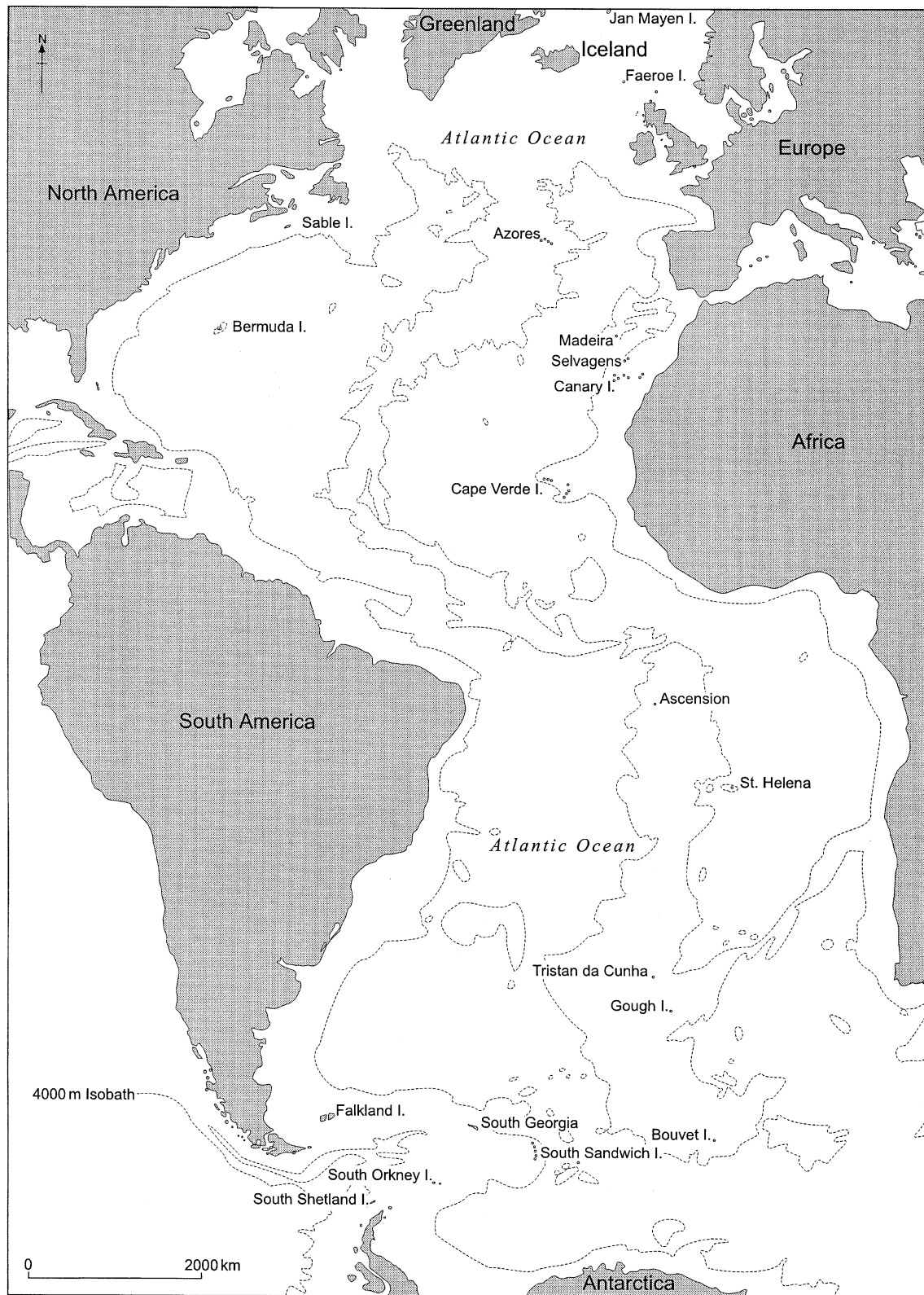
From south to north, the islands of the Atlantic covered here include the Scotia Arc (South Shetland, South Orkney, South Sandwich, and South Georgia), Bouvet, the Falklands, Tristan da Cunha and Gough, St. Helena and Ascension, Macronesia

(Cape Verde, Canaries, Madeira, and Azores), Bermuda, Sable, Faeroes, and Jan Mayen (Fig. 1). The Caribbean islands, Iceland, Great Britain, and Ireland are covered elsewhere.

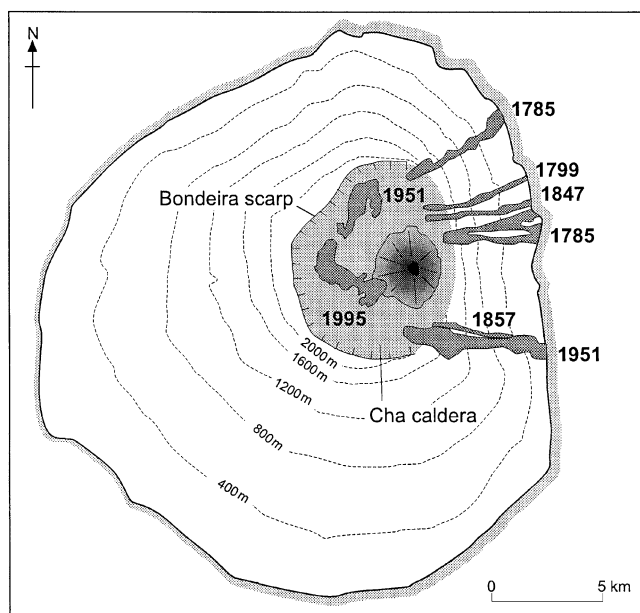
Setting

With only a few exceptions such as Bermuda and Sable, it is clear from the spatial distribution of almost all of the other individual islands or island groups in the Atlantic Ocean that most owe both their location and coastal morphology to a turbulent tectonic and often volcanic past. The tectonic opening of the Atlantic Ocean is the key factor in this history. Commencing in the Late Jurassic some 140 million years ago and fully open some 65 million years ago (Hansom and Gordon 1998), the Atlantic basin is characterized by the westward movement of both American plates and the eastward movement of the African and Eurasian plates. Basaltic magma upwelling into the crustal gap in the mid-ocean spreading center has over time produced a long submarine ridge composed of a series of fissure volcanoes known as the mid-Atlantic Ridge. In places the volume of upwelling magma has been sufficient to allow the construction of volcanoes that extend to the ocean surface and form the individual volcanic islands and island groups of Bouvet, Gough, Tristan da Cunha, St. Helena, Ascension, Macronesia, Kelard, and Jan Mayen. Elsewhere, where more than two plates and spreading directions are involved, this relatively simple mid-Atlantic geometry is more complex. For example, the development of the Scotia Arc in the South Atlantic was contingent on the opening of the Drake Passage and the eastward movement of several micro-plates along a progressively elongating arc. The boundaries of the Scotia Arc are marked in the south by the South Shetland and South Orkney Islands, in the north by South Georgia and in the east by the still-extending volcanic island arc of the South Sandwich Islands (Hansom and Gordon 1998). As a result, the rocks of South Georgia, South Shetlands, and South Orkney are mainly composed of fragments of older rocks and intruded volcanics, most of which were created elsewhere and subsequently transported to their present sites. By contrast, 70% of the rocks of the South Sandwich Islands are recently extruded basalts (Baker 1990).

This tectonic and volcanic past has also ensured that, with exception of parts of the Scotia Arc in the South Atlantic and Bermuda and Sable Island in the North Atlantic, almost all of the remaining mid-ocean islands are of volcanic origin (Bird 1985a, b). Many remain as active volcanoes that rise from deep water and their coastlines are mainly characterized by steep cliffs cut into lava, ash or hyaloclastite. Since many of the volcanoes themselves are youthful, the coastlines are also young and in a state of continuing adjustment in response to



Atlantic Ocean Islands, Coastal Geomorphology, Fig. 1 Location of the Atlantic Ocean Islands



Atlantic Ocean Islands, Coastal Geomorphology, Fig. 2 The island of Fogo, Cape Verde group is an active volcano and has a steep coast that is characteristic of many of the Mid-Atlantic Islands. 18–20th century lavaflows indicated as map

marine activity and ongoing volcanic events. For example, Fogo, in the Cape Verdes, is an active stratovolcano with a symmetrical cone rising out of the ocean (Fig. 2). Its slopes are composed mainly of subaerial lava flows, some of which are dated, and all of which end in marine cliffs, sometimes with small cliff-foot beaches composed of basalt boulders and gravels.

The Scotia Arc

Like most Antarctic coasts, the coasts of the Scotia Arc islands are mainly cliffed and rocky. Many of the cliffs are cut in glacier ice and beaches are regionally rare (Hansom and Gordon 1998). Rapid cliff retreat rates of 1 cm a^{-1} have been calculated for the cliffs of Fildes Peninsula in the South Shetland Islands and are thought to be the result of efficient frost shattering (Hansom 1983a). Since all of these coasts are affected by sea ice for up to eight months of the year, the more normal coastal processes associated with wave and tide operate intermittently and often in conjunction with floating ice fragments (Hansom and Kirk 1989). This means that the relative order associated with wave processes is annually swept away and replaced by the bulldozing and grinding action of floating ice so that beach sediment is moved in a nonselective way to produce disorganized fabrics and landforms. In the South Shetland Islands much of the coast is either rock or ice-cliff, although some glacier termini rest on a low rocky platform at sea level. There are important exceptions, however, and significant beach

development occurs on the peninsulas that protrude beyond the ice cover. The Fildes and Byers Peninsulas on King George Island and Livingston Island, respectively, are adorned with extensive emerged beach deposits, some of which connect offshore islets and skerries to the shore platforms of the main island (Fig. 3). Prominent shore platforms occur at sea level and at altitudes of 3–8 m, 11–17 m, 28–50 m and 85–100 m above sea level and are cut into a variety of rock types (Hansom 1983a). Although gravel beaches are found up to 200 m above sea level the most extensive are sited on the lower shore platforms at altitudes of 2.5–3 m, 5.50–6 m, 8–10.5 m, 12–13.5 m, and 16.5–18.5 m above sea level, as well as at present sea level (John and Sugden 1971; Hansom 1983a). Since the extensive sub-horizontal shore platforms that occur at present sea level in sheltered locations give way to steeper, ramp-like platforms in more exposed locations, Hansom (1983a) related their development to the interplay between the frequency of floating ice blocks that horizontally abrade the surface and wave activity that produces ramps. The upper part of the sub-horizontal platforms within sheltered bays is often adorned with an undulating layer of boulders that have been smoothed and compacted into extensive boulder pavements by the action of floating ice blocks (see Boulder Pavement) (Hansom 1983b). Three of the islands, Penguin, Bridgeman, and Deception, are relatively new and active volcanoes and erosion of the friable rocks of the outer coast has produced cliffs of a variety of heights. The inner crater of Deception Island has been inundated by the sea and has several sand and gravel beaches that are frequented by penguins and tourists alike, in sharp contrast to the steep rocky outer coast. Whereas the extensive beaches of parts of the South Shetland Islands allow for easy landing for both wildlife and boats, the coastline of the South Orkney Islands is more formidable. Three of the four largest islands, Coronation, Powell, and Laurie are dominated by large permanent ice caps spilling down steep rocky cliffs which plunge below sea level. A few bouldery beaches occur, for example, on Laurie Island. The remaining island, Signy is characterized by ancient schists and amphibolites that have been glacially scoured to produce a relatively low and rolling coastal plain with a rocky sloping shore albeit with only a few bouldery beaches. The coast of the South Sandwich Islands is as inhospitable as that of South Orkney, since only a small proportion of land is free of ice and there are few beaches or easy landing points. The South Sandwich group consists of a 400-km-long volcanic island arc caused by the subduction of the South Sandwich plate beneath the South American plate. Rising from an 8,428 m-deep ocean trench, the eleven main islands are all volcanoes made up of 60% lava and 40% tephra (Baker 1990). The coastline is steep and cliffed but because many of the islands are still active volcanoes, erosion of softer tephra has produced a crenulate coastline with infrequent small bouldery beaches. All of the islands except Zavodovski are characterized by reefs and



Atlantic Ocean Islands, Coastal Geomorphology, Fig. 3 Ardley Island, in the South Shetland Islands, is connected to one of the main islands by an intertidal sand tombolo and is adorned with raised beaches whose altitudes are the same as those on the adjacent islands.

skerries and these are especially well-developed between smaller island clusters, such as in the extreme south at Thule, Cook and Bellingshausen islands. Low altitude bare coastal plains do occur on some islands, such as Zavodovski and Bellingshausen, but few scientists have investigated these islands and so, the nature of the plains is unknown. In the strong westerly ocean circulation of the Scotia Sea, pumice from South Sandwich is known to travel great distances. The pumice from a submarine eruption close to Zavodovski Island in March 1962 reached the beaches of New Zealand in late 1963, having been transported east within the Antarctic Circumpolar Current at $11\text{--}17\text{ cm s}^{-1}$ (Hansom and Gordon 1998).

South Georgia on the northern limb of the Scotia Arc, is a 170-km-long and 40-km-wide sequence of volcanoclastic sediments folded into a mountain range that rises out of the Atlantic Ocean to heights of over 2,700 m. It is not volcanically active but is heavily glacierized and has all the characteristics of a glacially eroded coastal fold mountain range. Long ridge-like peninsulas separated by deep glacial troughs jut out into the sea. The south coast is more heavily glacierized than the north and is extremely rocky and dominated by steep cliffs cut in ice or rock. The few small cliff-foot gravel or sandy beaches that occur are swept by storm waves from the southwest and occasionally by storm surges (Hansom 1981). On the north-facing coast, the glacier cover is more restricted and a significant amount of ice-free ground exists, particularly on the peninsulas. The peninsulas are flanked by high and steep cliffs but these often have narrow

shore platforms at their base together with numerous offshore islets and skerries. Some fragments of emerged shore platforms are known to occur at altitudes of 2, 5, 6–7, and 20–50 m above sea level but a systematic study of distribution and altitude is lacking (Clapperton 1971). The lower of the shore platforms are often adorned with emerged beaches at 2–4 m and 6–7.5 m above sea level (Clapperton et al. 1978). Within the northern bays and fjords of South Georgia lie the most extensive beaches of the Antarctic region. Fed by numerous debris-laden glaciers and short glacifluvial streams, substantial areas of the sheltered inner fjord heads have become infilled by glacifluvial outwash plains fringed by extensive sand and gravel beaches (Hansom and Kirk 1989). The largest of these outwash beaches occur at Salisbury Plain, Fortuna Bay, Cumberland Bay and St. Andrews Bay (Fig. 4) (Gordon and Hansom 1986). Some of the inner reaches of the fjords of South Georgia are affected by sufficiently large quantities of floating ice calved from adjacent glaciers to allow the development of intertidal boulder pavements similar to those produced by sea ice further south (Hansom 1983b).

Falkland Islands

The coastline of the Falkland Islands, like the islands themselves, is low, flat, and reminiscent of the coastline of the Outer Hebrides of Scotland. Although hundreds of smaller islands exist, the main island group comprises West and East



Atlantic Ocean Islands, Coastal Geomorphology, Fig. 4 Saint Andrews Bay, South Georgia, is fed by glacial sediment and is characteristic of the extensive outwash beaches of the north coast of the island

Falkland separated by Falkland Sound. The intricate and crenulate nature of the coastline is probably more related to the submergence and faulting of the underlying Devonian, Carboniferous, and Permian limestones and sandstones, rather than to the efficiency or otherwise of marine erosion and deposition. In spite of the majority of the coastline being rocky, steep and high cliffs are mainly absent except in the extreme west where 100–200 m cliffs occur at Beaver Island. Everywhere and especially within the broad embayments and inlets, there are innumerable small islets and skerries whose detailed morphology and outline appear to be structurally controlled. A notable feature of the rocky coastline is the abundance of dense stands of giant kelp in the nearshore. On the western flank of the fault that controls Falkland Sound, narrow coast-parallel outcrops of hard and softer rock have been eroded to produce an intricate series of small headlands composed of more resistant rock separated by small arcuate bays cut into less resistant beds. The result is

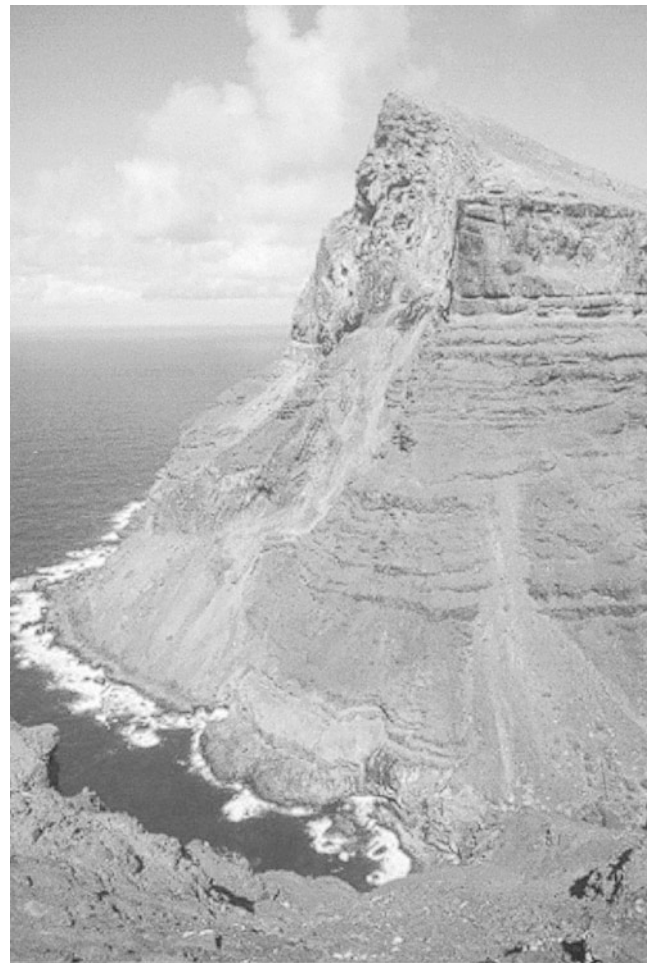
an essentially linear coastline stretching the entire length of Falkland Sound, broken by regular and uniform bays where the outer rocks have been breached. The crenulate nature of the coastline and extensive areas of nearshore shallow water means that a wide range of sheltered locations exists for potential beach development. Streams, although abundant and rarely dry, do not appear to contribute significantly to beaches, since there is a lack of beach development adjacent to the mouths of the creeks and streams entering tidal waters. Beaches of sand and gravels tend to be located in either outer coast sites where ocean waves gain access or within inlets where more open aspects allow wind-generated local waves to produce beaches. The high proportion of shell-sand of the beaches points to an important biological input from nearshore shallow waters. Significant beach development on the outer coast occurs at Bull Point, Lively Island, and at Bertha's Beach where sandy spits and tombolos connect small islets to mainland East Falkland. Paloma Beach, Elephant Beach, and Concordia Beach in the north of East Falkland are all examples of large open coast sandy beaches where strong winds have resulted in blown sand spreads over inland areas. In West Falkland, large beaches occur within the inlet of Whitsand Bay. Unfortunately, as a result of indiscriminate mine-laying during the Falklands War, several of the beaches of East Falkland are now unsafe and access is restricted.

Bouvet, Gough, Tristan da Cunha, St. Helena and Ascension

Bouvet Island (Fig. 1) stands at the southern end of the mid-Atlantic Ridge, the slopes of the central cone terminating on all sides in precipitous cliffs that descend abruptly to sea level. Probably a complex of volcanic cones, the island rises almost symmetrically to a flat ice-covered dome 935 m high. The ice covers the eastern side of the island where it reaches the sea in an ice cliff 122 m high. The north and west sides are free of ice but are much steeper. Between 1955 and 1958, a new 0.2 km² rock platform emerged at 25 m above sea level as a result of either eruption or rock falls associated with tremors (Stonehouse 1972). Since little low-level ground was previously to be found around the coast of Bouvet, the new addition has become a prime-breeding site for penguins, petrels, and seals. Gough Island (Fig. 1) lies 2,300 km to the east of the Cape of Good Hope, South Africa and is the eroded summit of a Tertiary volcano. Although the island is mountainous, rising to 910 m above sea level, the slopes are cut by deep gullies or gulches and the coastline is characterized by steep cliffs that appear to have undergone erosion to produce a variety of narrow boulder beaches at the foot of the cliffs. There are also numerous islets, stacks, and skerries that mostly lie within 100 m of the coast. Tristan da Cunha lies 350 km to the northwest of Gough. It is a circular island of

98 km² with four other small islands close by. The base of the central volcanic peak lies 3,700 m deep on the seabed and the summit rises to almost 2,100 m above sea level. The lower slopes are almost entirely lava and form high cliffs that surround the island. Emerged beaches, platforms, and caves occur at 5 m above sea level on Tristan but 12 m above sea level on the small adjacent islands, indicating differential tectonic uplift in the area (Bird 1985b). In some places, small and inaccessible beaches occur, some of which are sandy. A small plateau at the foot of the main cliffs on the northwest side at Herald Point provides the site for Edinburgh, the only inhabited part of the island. The cliffs are lower along the Edinburgh coast, allowing access to the sea via small gullies. A narrow gravel beach has developed along this stretch of coast, fed by material eroded from the cliffs. In the months following the 1961 eruption at Big Point, on the updrift section of this coast, rapid recession of the cliff was noted (10 m in 8 weeks). This contributed to accretion of the downdrift gravel beaches which formed a spit enclosing a small lagoon (Bird 1985b).

St. Helena (Fig. 1) is 122 km² in area and lies in mid-ocean, 2,900 km from Brazil and 1,900 km from the west coast of Africa. Like Tristan and Gough, the coastline comprises high stepped and sometimes vertical cliffs cut by steep v-sided valleys. Emerged shore platforms stand at 4–6 m above sea level (Bird 1985b) and the numerous offshore islets and skerries common on the south and west coast, such as at Egg Island and Cockburn Battery, are almost absent from the north and east coast. The highest cliffs occur at Flagstaff Bay in the north and at Man and Horse Cliffs in the southwest where they reach 580 m. The vertical steps in the coastal cliffs often correspond to differences in the composition of the outcropping of lava flows. At Great Stone Top, Turks Cap, and Prosperous Bay on the east coast, thick flows of trachyte form vertical faces, which cascade debris onto stone chutes below (Fig. 5). Some of these extend into the sea to form fringing boulder beaches but nowhere on the outer coast are these well developed or extensive. Only in favored locations where the deep gulches reach the coast do small beaches, some of sand, occur, such as at Lemon Valley Bay. At Sandy Cove on the south coast, the sandy beach is backed by calcareous sand dunes. In mid-ocean some 1,300 km north of St. Helena lies Ascension Island, a Holocene stratovolcano with no known historic eruptions (Simkin and Siebert 1994). Although more arid than St. Helena, the coasts of both islands are similarly high and surrounded by steep cliffs cut into volcanic lava. At Boatswain Bird islet, a natural arch has been eroded into a 98 m high monolithic stack of trachyte (Bird 1985b). However, Ascension also has superb white sand beaches derived from shell and coral. Outside of the Caribbean area, the Atlantic Ocean is not noted for its corals although small structures are known from the coast of western Africa (Trenhaile 1997). Although little seems to be known of



Atlantic Ocean Islands, Coastal Geomorphology, Fig. 5 The steep cliffs of Saint Helena are capped by outcrops of trachyte at Great Stone Top on the south east coast. (Photo courtesy of Barry Weaver)

the nature of the sediment supply to the Ascension beaches, the carbonates can only come from shells and coralline algae (such as *Lithothamnion*) growing in the narrow shallow water fringing the island. Erosion of adjacent rock shores and terrestrial sediment in-washed by infrequent rains probably account for the remainder of the beach sediment.

Macronesia

The island groups comprising Macronesia stretch lie in a long line that extends north between the coast of West Africa and the Portuguese coast (Fig. 1). All of the island groups are volcanic in origin and many remain active volcanoes. Four main stratovolcanoes occur in the arid Cape Verde Islands, Fogo, Brava, Santo Antao, and San Vicente. Fogo is 2,829 m high yet only 24 km wide and so the steep seaward slopes end abruptly in retreating cliffs. The volcano has erupted 10 times between 1500 and 1995 each one sending long

streams of lava down the flanks into the ocean and altering the coastal geometry. A similar picture characterizes the other islands of the group, each being surrounded by steep cliffs cut by deep gullies at the foot of which occur infrequent gravelly beaches. Emerged features have been noted at six levels up to 100 m above sea level. Some bays contain fringing algal reefs and in some locations boulders have been moved inland by wave activity by up to 200 m (Bird 1985b). The Canary Islands comprise seven main islands, six of which host volcanoes. Some of these have erupted as recently as 1971. The coastline of the Canaries resembles that of the other mid-ocean volcanic islands in as much as the central volcanic spine of the islands falls steeply to a mainly rocky and cliff coast cut by deeply incised ravines. For example, in the northwest of Tenerife at Acantilado de Los Gigantes, sub-vertical cliffs reach 500 m. The occurrence of such cliffs has constrained some of the tourist-related expansion that many of the coastal towns and villages have undergone in recent years. Some cliffs have been formed in sediments brought down by torrents in the ravines. These deposits have since been tectonically uplifted together with the boulder beaches that once fronted them. The north coast of Gran Canaria has good examples of such cliffs, together with well-developed shore platforms a few meters above sea level. However, the volcanism of the Canaries has in places produced several long and low craggy volcanic peninsulas that have allowed beach development to occur. Fuerteventura, Lanzarote, Tenerife, and Gran Canaria all have sandy beaches some with extensive sand dunes. The long white sand beach at San Andres, on the southeast coast of Tenerife, has been artificially nourished with sand brought from the Sahara Desert. Between Maspalomas and Playa del Ingles, in the south of Gran Canaria, 328 ha of fine shell sand have accumulated at the mouth of the Fataga ravine to enclose a freshwater lagoon. On account of the relative aridity, most dunes in the Canaries are sparsely vegetated and the unfixed dunes at Maspalomas advance east towards the lighthouse at a meter per year. Some of the dunes have succumbed to tourist development, such as at Playa de las Canteras in the north of Gran Canaria. In the easternmost part of the islands, the supply of beach sediment from the near-shore and the inland ravines is augmented from a more exotic source. Fuerteventura and Lanzarote are affected by the summer *scirocco*, a hot wind from the Sahara 90 km to the east, which carries large quantities of dust and desert sand to the islands. Locally known as the *kalima*, it turns the day into twilight and regularly covers surfaces with a thin layer of sediment. Over the centuries, this sediment has been an important source of sand for the beaches of the easternmost islands. In the south, Fuerteventura narrows at the Pared isthmus where sandy beaches extend eastward to connect the original island of Jandia to the main island by what is now a low and narrow peninsula. On the eastern shore, the sands

of the 26 km-long Playa de Sotavento (Sp: leeward) are protected from the dominant westerly waves. On the western side, the Playas of Cafete, Pared and Barlovento (Sp: windward) are more exposed with strong currents and undertows. Nearby Lanzarote is a similarly dramatic landscape of lava fields and steep cliffs with intervening sand beaches, such as the sweeping Puerto del Carmen beach on the east coast. Accretion is common at many of the valley-mouth inlets.

Maderia comprises the main island itself, together with the smaller island of Porto Santo, the nearby *Ilhas Desertas*, together with the uninhabited *Selvagens* to the south. All of these islands are volcanic in origin but have not been active in the last 1.5 million years. As a result the, main volcanic core of Madeira, together with its plateau-like lateral subsidiary vents, has become eroded to produce deeply incised valleys and gorges running down to the sea. Between the mouths of the river valleys are high cliffs of vertical columns of basalt with layers of red and yellow tufa exposed in places. The 580 m plunging cliffs of Cabo Giroa, west of Funchal, are among Europe's highest, but high cliffs are found everywhere on the coast of Madiera. In the north, fragments of shore-platform occur as well as small islets and skerries, particularly near Faial. There are several well-developed stacks near Ponta do Sao Lourenco in the northeast. Small beaches of rounded gray basalt gravel occur at several places, particularly where river mouths occur. The only sandy beach is at Prainha, on the extreme east of the island where the sand is mainly basalt. Shell sand occurs on the low plateau area in the extreme east of the island. The nearby island of Porto Santo is 14 km long and 5 km wide, its generally low volcanic profile veneered by thin deposits of sand, calcareous sandstone and clay. The cliffs of the north coast reach 100 m but the south coast is dominated by an 8 km long white sand beach fronted by a shallow sandy-floored bay backed by low cliffs of cemented sands. Protected from the main force of southwesterly storms by Madeira, from the west by the small island of Baixo and from the north by its own cliffs, the south coast of Porto Santo is relatively sheltered. It appears to have been subject to sediment accumulation over a substantial period of time derived both from a combination of shell sand from shallow nearshore surfaces, aeolian sand blown west on the *scirocco* from the Sahara to the east and from topsoil erosion caused by early deforestation. The *Selvagens* are composed of limestone capped by lava and ash and are cliffed to 100 m in the north but with a gentler beach-fringed coast in the south. Emerged beaches and dune calcarenites occur on Pleistocene marine terraces at 3 and 7 m above sea level (Bird 1985b).

The most northerly of the islands of Macronesia, the nine islands of the Azores are a widely separated group of mountainous but fertile islands which share the steep nature of many of the mid-oceanic islands but also have long beaches and many fishing harbors. The coast of the largest island, Sao

Atlantic Ocean Islands, Coastal Geomorphology, Fig. 6

Glacial troughs have been cut into horizontally bedded basalts and subsequently flooded by Holocene sea level rise to produce cliffs that are characteristic of the inner coast of the Faeroese fjords. The plunging cliffs of the Faeroese outer coast are everywhere exposed to high energy Atlantic waves



Miguel, is a microcosm of the Azores coastline. Where the coast is backed by higher volcanic peaks, it is characterized by steep cliffs fronted by patchy low shore platforms, offshore stacks and islets, such as at Mosteiros. Emerged beaches and platforms occur at various heights up to 60 m (Bird 1985b). However, the highly indented coast also has short sandy beaches between cliff headlands. Where the hinterland is lower, longer sand beaches occur such as at Praia dos Moinhos and at Populo where small pine-clad sand dunes occur. Several coastal features of note occur on the other islands. Pico Island is 42 km long and dominated by a 2351 m cone-shaped strato-volcano of the same name. Various historical lava flows have extended the coastline and now have been eroded back to form low cliffs. In places sheltered from the dominant westerly waves, smaller volcanic forms survive on the coast, such as at Barca on the island of Graciosa where a volcanic cone has been sectioned by marine erosion to form bays fringed by crumbling cliffs of ash that cascade onto boulder beaches below. The 1957/58 eruption at Capelhino on Faial showered the adjacent coast in ash and contributed to accretion of the beach. At Porto Pim, also on Faial, a substantial sandy spit has developed (Bird 1985a). Such is the power of the surf reaching the Azores that Pico has been chosen as the site of a pilot plant to produce energy from waves.

Faeroes and Jan Mayen

The rugged outer coast of the 16 main islands that comprise the Faeroes is the result of the deep incision of past glaciers into a thick pile of horizontally bedded flood basalts. These are slowly eroding, especially on the west and north coasts, due to constant exposure to high-energy westerly and

northerly North Atlantic waves. The outer coast is dominated by high and sub-vertical cliffs that reach 820 m at Hestmul in the north of the island of Kunoy. The rate of recession is unknown but is higher on the seaward facing slopes where cliffs have developed that tend to be steeper than immediately adjacent landward-facing glaciated slopes, such as at Tindhølmur in the extreme west. The slopes of the inner coast are the sides of over-steepened glacial troughs that have been flooded by Holocene sea-level rise and are more subdued and better vegetated than the cliffs of the outer coast (Fig. 6). Small skerries, islets, and stacks (Fr: drangur) are common as are deep clefts and geos (Fr: gjogvs) where dykes and faults have been differentially eroded into the horizontal basalts of the west coast. Small, often structurally controlled, ramped shore platforms are common around most of the coast but tend to be more fragmented and smaller on the more exposed west and south coast. More continuous and better-developed platforms occur near the entrances to the fjords, particularly in the east. For example, well-developed shore platforms occur on the east coast of Vidoy. There are very few beaches in the Faeroes and the ones that do occur are almost exclusively found at the sheltered fjord heads. In such locations, wave approach is unidirectional and the limited amounts of sediment supplied by the small streams accumulate to produce small pocket beaches composed of black basalt sand (such as at Tjørnuvík, Funningsfjordur, and Vidareidi) and occasionally backed by low sand dunes (such as at Sandur and Saksun).

The coastline of Jan Mayen is dominated by the volcanic bulk of Beerenberg, a 2,277 m stratovolcano that comprises the northern half of the 54-km-long island (Norsk Polarinstitutt 1959). The coastline of Nord-Jan is steep and rugged, comprising 200–400-m-high cliffs eroded in ash, lava and tephra as well as into the ice walls of tidewater glaciers

that spill down from the central crater. The coastline shows signs of recent advance as a result of lava tongues from eruptions in 1970 and 1985, which have extended into the ocean. These recent eruptions and flows are an extension of the mode of construction of the island over an estimated 700,000 years. Such additions to the coastline have resulted in the base of preexisting cliffs becoming buried by newer lavas and it is onto this platform that more recent glacier terminal moraines have been deposited, such as at Smithbreen in the northeast of the island (Kinsman and Sheard 1962). At Kroksletta, near the northern cape, 4,000-year-old moraines that rest on top of an emerged shore platform and beach at 8–10 m above sea level have been buried by subsequent lava flows. The protruding seaward edge of the Kroksletta lava is now cliffed to a height of 5–13 m (Kinsman and Sheard 1962). The south part of Jan Mayen is a hilly ridge of scoria craters and mounds and trachyte domes whose lower elevations and older age has produced a coastline of low gradient rocky platforms connected by gravelly beaches. In several places around the Sud-Jan coastline a prominent rock platform is present upon which is sited emerged beach gravels. This feature is particularly well developed on the north coast of Sud-Jan at Sorbukta, Engelsbukta, and Haugenstranda. Sediment supply to the central section coastline that joins Sud-Jan to Nord-Jan appears to be healthy both from erosion of the lava cliffs to the west and from shallow water sources: the 10 m depth contour lies about 0.7 km offshore. This has resulted in the construction of a substantial barrier beach

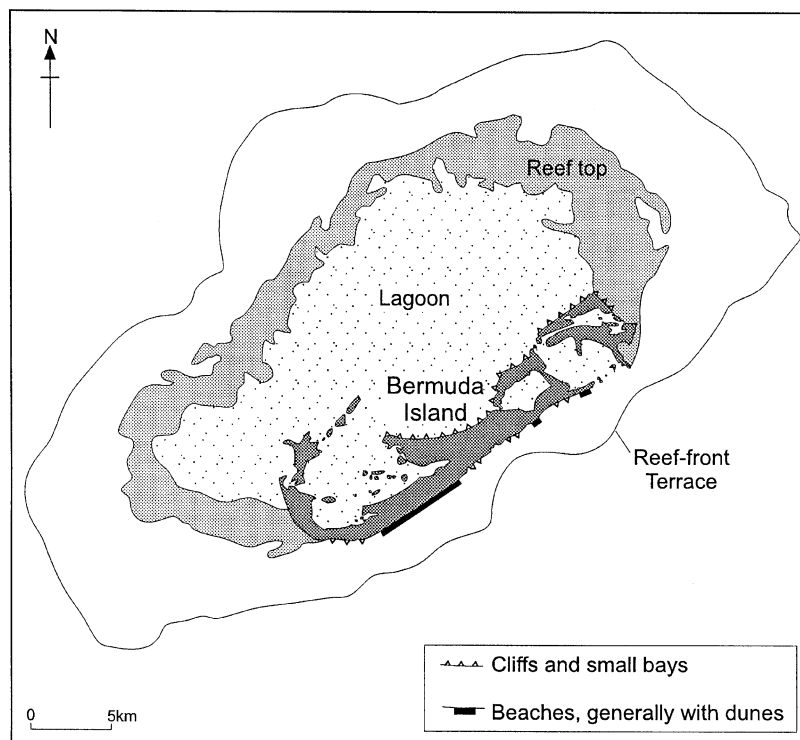
enclosing a lagoon along an 11-km-long stretch of south-central coast at Lagunevollen, with a smaller barrier and lagoon at Stasjonsbukta on the north coast. The composition of these barrier beaches is not known but assumed to be composed of mixed sands and gravels transported from the west. Such a large beach complex is unusual for a small mid-oceanic island. However, the full impact of Atlantic westerly waves and swell at Sud-Jan is modified by a large circular submarine reef (probably an eroded volcanic cone), which reaches to within 11 m of the sea surface some 10 km south of Sud-Jan. Although the north coast is subject to variable amounts of sea ice for an average of four months, the south coast is mainly free of ice (Gloerson et al. 1992). Elsewhere, the coastline of Sud-Jan is steep and rocky but the cliffs are not as high as in Nord-Jan and small gravelly cliff-foot beaches fed by rockfall and wave erosion of the adjacent lava occur.

Bermuda and Sable

With two notable exceptions, Bermuda and Sable, all of the Atlantic Ocean Islands have, or have had, a close and recent association with tectonic and volcanic activity. However, Bermuda, lying to the east of the Carolinas coast of the United States is a group of 150 limestone islands located about 1,000 km east of Cape Hatteras (Fig. 1). The islands rest on the southern margin of a 650 km² platform that is presently

Atlantic Ocean Islands, Coastal Geomorphology, Fig. 7

The calcarenite islands of Bermuda are sited atop a lava plateau. The present coastline is mainly composed of low cliffs with long beaches in the south (after Vacher 1973)



submerged to depths of 20 m. However, boreholes at Bermuda have penetrated through a thin capping of calcarenites to meet lava at a variety of depths of 171, 43, 33, and 21–24 m (King 1972). The platform of volcanic rocks is the site for the northernmost coral reefs in the North Atlantic. According to Vacher (1973) the present-day Bermuda platform consists of four geomorphological-ecological provinces: a reef-front terrace at 20 m depth; a main reef composed of 4 m deep coral-algal reefs everywhere except in the south where algal intergrowths occur; a 16 m deep lagoon in the north and the islands themselves forming a northeast trending chain near the southern edge of the platform (Fig. 7). The exposed limestones of the islands are Pleistocene calcarenite, 90% of which is aeolian and the rest is beach and shallow water in origin (Vacher 1973). The aeolianites were built from calcarenite produced mainly offshore and transported to the shore and so the inference is that during an earlier phase of submergence, the coastline was supplied by abundant reef-derived materials, which then led to beach and dune development before becoming lithified. This supply of reef-derived sediment to beaches and dunes continues today and progradation appears to be healthy (Bird 1985a), but patchy. Nevertheless the present supply also appears to be more restricted than the former supply, since only on the island's south shore do extensive beaches backed by dune occur. These extensive sandy beaches are pink because of large numbers of fresh *Homotrema* clasts derived from the reefs 1 km from the shoreline (Mackenzie et al. 1965). However, much of the modern south coast is cliff with only limited sources of offshore-derived sediment. Unlike during the earlier development of the calcarenites, the sediments of the lagoon now appear to play little part in beach supply. As a result, the islands' lagoon-facing shore is erosional with a line of cliffs and few pocket beaches (Vacher 1973).

Sable Island (Fig. 1), on the continental shelf to the east of Nova Scotia, Canada, is a low and wind-swept series of sandy islands famous for its sandy shoals and shipwrecks. The development of the Sable Island Bank, on which the islands sit, is related to the proximity of the continental shelf-break and the maximum ice positions of the Late Wisconsin and later readvances. These ice movements deposited a thick suite of glacial till and superficial sandy material, that was subsequently subject to transgression, modification, and transport (King 2001). The island has spectacular but desolate sandy beaches backed by sand dunes that reach up to 30 m high and cover 40% of the island's surface (Byrne and McCann 1995). The intertidal zones of the long sandy beaches are characterized by prominent shoreface attached ridges with intervening depressions that reach up to 12 m deep (Dalrymple and Hoogendoorn 1997). Strong along-shore currents cause eastward migration of these bars along-shore at rates of up to 50 m a⁻¹ and at angles of up to 50° to the coastal orientation (Dalrymple and Hoogendoorn 1997).

The dunes of Sable consist of primary dunes that have developed *in situ* together with secondary dunes that have migrated across the island. The resultant coastal morphology represents a mix of both natural processes and anthropogenic disturbance. For example, the constantly changing coastal outline has resulted in the relocation of the lighthouse and is thought to be partly due to dune mobilization and reduced vegetation cover under the enhanced grazing pressure of introduced ponies (Owens and Bowen 1977). Sable Island has undergone 14.5 km of eastward migration but, in spite of this, its 30 km² has been maintained over the past 200 years (Cameron 1965). Sable Island thus seems to be subject to a regime of deposition that appears roughly balanced by an equivalent amount of erosion.

Cross-References

- ▶ [Antarctica, Coastal Ecology and Geomorphology](#)
- ▶ [Atlantic Ocean Islands, Coastal Ecology](#)
- ▶ [Boulder Pavements](#)
- ▶ [Cliffs, Erosion Rates](#)
- ▶ [Cliffs, Lithology versus Erosion Rates](#)
- ▶ [Glaciated Coasts](#)
- ▶ [Scour and Burial of Objects in Shallow Water](#)
- ▶ [Shore Platforms](#)
- ▶ [Volcanic Coasts](#)
- ▶ [Wave Power](#)

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Atolls

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Atolls are coral reefs found in the open ocean consisting of an annular rim surrounding a central lagoon. There is a general presumption that atolls have volcanic foundations and some 425 have been recognized around the world (Wiens 1962; McLean and Woodroffe 1994). However, because of similar superficial morphology many shallower water coral reefs have been termed “shelf atolls” (Ladd 1977). In Indonesian waters 55 such reefs have been recognized though many will not have the volcanic foundations of the majority of atolls found in the Pacific and Indian Oceans.

The largest atoll is Kwajalein in the Marshall Islands ($120 \times 32 \text{ km}^2$) followed by Rangiroa in the Tuamotus ($79 \times 34 \text{ km}^2$), though many smaller atolls are only a few kilometers in diameter. In a sample of 99 atolls, Stoddart (1965) indicated a mean area of 272.5 km^2 , and atolls worldwide are considered to have a total area of $115,000 \text{ km}^2$.

The Contribution of Atoll Studies to Marine Geology

Ever since the historic scientific expedition of Charles Darwin in the *Beagle* (1832–1836) atolls have captured the imagination of marine geologists and Darwin’s synthesis of his ideas on the formation of coral reefs (1842) have stimulated 160 years of research. This stimulus has come from the facts that:

- reef-building corals grow in shallow (<100 m) water
- many atolls rise from depths of several thousand meters
- there is a similarity of morphology (annular rim and lagoon) wherever atolls are formed.

Darwin’s hypothesis of atolls evolving from fringing reefs around a volcanic island via a barrier reef stage was the catalyst for subsequent mainstream investigations into glacio-eustatic sea-level change, vertical and horizontal tectonic movements, and the foundations and internal structure of oceanic atolls. Atoll investigations have subsequently played a major part in research into carbonate diagenesis, isostasy, continental drift, plate tectonics, and seafloor spreading. They have generally confirmed the remarkable insight of Darwin regarding the origin of atolls.

The Internal Structure and Origin of Atolls

Darwin recognized that resolution of the problem of atoll development required deep drilling and, after several aborted attempts, T. Edgeworth David of Sydney University in 1896–1898 drilled the atoll of Funafuti to a depth of 339.7 m, all in shallow water coral, but so fragmented that those who opposed subsidence argued that the core passed through a detrital slope deposit. It was to be almost 50 years before drilling totally penetrated the coral of an atoll to the basaltic basement at a depth of more than 1,300 m on Enewetak atoll. Subsequent drilling, much of it associated with atomic tests, at Enewetak and Bikini atolls but also at Midway by the US Geological Survey, and at Muroroa and Fangataufa in French Polynesia, by the French, confirmed both the volcanic foundations of atolls and the depth of in situ coral far below the photic limits of both modern coral growth and growth at glacially lowered sea levels. Volcanic basement was determined at 1,283 and 1,408 m at Enewetak,

415 and 438 m on Mururoa (though seismic results show basalt as shallow as -170 m, Guille et al. 1996), 153 and 503 m at Midway and 270–400 m at Fangataufa. These figures are typical of those from many other atolls, which have now been investigated by drilling or seismic survey.

While shallow water corals have been recognized to depths greater than 1000 m, clearly confirming widespread subsidence, considerable lithification, cementation, and diagenesis has taken place in the lower sections of the atoll foundations, which have been determined as up to 30,000,000 years in age. Alteration of the carbonate rocks results from fluctuations of more than 100 m in sea level with vertical migration of phreatic water table favoring the transformation of the original aragonite of the biogenic structure to calcite. At depths greater than 500 m cold permeating sea water dissolves both aragonite and calcite leaving behind only low-Mg calcite. Dolomitization takes place within thick aquifers of mixed fresh and saline waters and is common in the lowest portions of the atoll foundations.

Like other reefs, atolls were most recently exposed during the last glaciation when karstic processes prevailed on the atoll surface, as in previous low sea-level phases, forming a clearly identifiable “solution unconformity” beneath Holocene reef material which is commonly only 10–20 m thick. Karstic features, including the saucer-like shape of atolls, have long been considered to be the major determinant of the gross morphology of atolls (e.g., see Purdy 1974).

The internal structure and radiometric ages of basalts beneath atolls have been integral parts of the evidence to develop the ideas of plate tectonics and seafloor spreading (e.g., Scott and Rotondo 1983; Guille et al. 1996). Volcanoes develop at oceanic “hot spots” from which they migrate through tectonic plate movement, acquiring coral reefs if they are in warm enough waters. Subsidence takes place due to both volcanic loading and cooling, and reefs and islands pass through the classic Darwinian sequence of fringing, barrier, and atoll formations producing linear chains along the direction of plate movement, such as the Hawaiian and Society Islands. If they are out of the zone of coral growth, subsidence continues and atolls become submerged as guyots, for example, the Emperor Seamounts to the northwest of the Hawaiian chain.

The Surface Morphology of Atolls

The high energy situation of most atolls means that once they reached sea level after their last period of exposure, they have quickly developed strong zonal patterns. Reef slopes commonly display groove and spur structures, reef crests are surmounted by significant algal ridges and reef flats of about 500 m width are a conduit for sediments which form accumulating aprons of sand within the lagoon. Lagoons,

with or without patch reefs are often of similar depth, within the range of 10–30 m with only a thick Holocene sediment veneer over the older Pleistocene foundations. Water circulation within the lagoons is complex, with wave setup raising swell side water levels and strong flow over the rim. Water escapes through leeward passes but in many instances a return bottom current back towards the swell side may be present.

Particularly on the swell and/or windward side of atolls, are linear islands termed “motus” separated by narrow passages or “hoa.” The majority are constructed of coral gravel broken from the reef front and deposited a set distance back on the reef flat as the entraining waves lose their transportational capacity. Spectacular episodes of deposition have been described during hurricanes or tropical cyclones but these are not absolutely necessary for motu formation as equatorial atolls such as the Maldives, not experiencing strong, tropical revolving storms, also have significant island development. These islands may be only a metre or so above high tide and are regarded as the coastal land-forms most susceptible to sea-level rise associated with global climate change.

Cross-References

- [Algal Rims](#)
- [Beachrock](#)
- [Bioconstruction](#)
- [Bioherms and Biostromes](#)
- [Cays](#)
- [Coral Reef Islands](#)
- [Coral Reefs](#)
- [Pacific Ocean Islands, Coastal Geomorphology](#)
- [Small Islands](#)

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Australia, Coastal Ecology

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Australia is a large island continent with over 61,700 km of coastline (including near by islands). The coastline stretches from the tropics (10°S) to cool temperate (44°S), north–south, encompassing 30° of longitude (113°E to 153°E), east–west. Australia's ocean territory is one of the largest marine jurisdictions in the world (16.1 million km²) and extends from the tropical epicenter of global marine biodiversity to the Antarctic (Zann 1995). Australia's continental inshore environment contains a major slice of the marine biodiversity of the Southern Hemisphere, including a large number of unique species (in the temperate south) and species threatened in neighboring regions (in the tropical north).

Several features distinguish Australian coastal environments from other coastal environments in the world: high climate variability; low runoff; and the influence of the El Niño–Southern Oscillation. Australia has the least runoff, the fewest rivers and the lowest rainfall of any inhabited continent. In addition, high evaporation rates and low continental relief (particularly along the southern coast) combine to markedly reduce the rate of runoff to oceans. Australia's high climate variability is partly associated with the effects of the El Niño–Southern Oscillation, which results in drought followed by heavy rains and flooding – particularly in the east (DEST 1996). In southern areas, ecosystems are attuned to drought/flood cycles while in northern areas, ecosystems are attuned to wet/dry cycles.

Australia's unique geological and tectonic history have produced a highly diverse and distinctive marine biota. Over geological timescales, Australia's coastal environment and biota has evolved and been greatly influenced by geological, tectonic, and climatic stability, and along the southern coast, by geographical and climatic isolation (Poore 1995). Australia has had a relatively stable climate due to the northward movement of the Australian tectonic plate (which has compensated for global cooling). The rate of change of relative sea level along the coastlines has been slow and largely controlled by changes in global sea level and eustatic processes, rather than localized tectonic movements. This has allowed the adaptation and migration of the marine biota. In addition, Australia's variable climate, range of coastal environments, and the low nutrient status of its coastal waters, have contributed significantly to the diversity and endemism of Australia's marine environments, particularly among some marine taxon (e.g., fish, mangroves, seagrasses, macroalgae, sponges). Low nutrient regimes generally promote biological diversity

and coevolutionary strategies to rapidly harvest, utilize, and recycle limited nutrient resources.

The position, size, and shape of the Australian continent have also contributed to the high level of marine biodiversity. Australia straddles the Tropic of Capricorn, so over one-third of the continent is in the tropics with the remainder in the temperate latitudes. The large continental landmass of Australia (and extensive continental shelf) and the shape of the continent provide extensive east–west coasts, with few contiguous coasts in the world possessing such extensive east–west coastlines. Australia's northern coastline adjoins tropical water masses and the southern coastline (the longest in the Southern Hemisphere) adjoins the Southern Ocean, resulting in distinct tropical and temperate biotas. The El Niño–Southern Oscillation also has a major influence on water temperatures and marine life.

In southern Australia, endemism among the marine fauna and flora has been enhanced by the isolation of the coastal shelf biota from other southern continents (by ocean basins) since the Paleocene or Eocene and also, from a latitudinal temperate gradient to the north (Poore 1995). Further, geological stability has allowed relatively undisturbed sequences, compared with many other parts of the world, and the survival of many ancient and relict species of Pangaean and Gondwanan origin, particularly for many invertebrate species (Durham 1979; Fleming 1979). The long east–west, ice-free, extent of the southern coastline (i.e., the longest stretch of south facing coastline in the Southern Hemisphere) has also contributed to the diversity among the temperate biota. Together, these factors have contributed to globally significant levels of biodiversity and endemism in the marine biota, particularly among the marine algae (Womersley 1990; Phillips 2001) and some invertebrate taxon (i.e., bryozoans, ascidians, sponges) (Edyvane 1996). In addition, many of the marine species, which inhabit the temperate reefs of Australia, are characterized by short larval periods and localized dispersal. For these reasons, there is a great tendency for local and regional rarity and endemism in temperate waters, with species distributions characterized by small, isolated, localized populations.

Australia's "State of the Environment Report" (Commonwealth of Australia 2001) concluded that Australia's marine environments and habitats are generally in good condition, to the extent that this can be measured. Many of Australia's remote areas are assumed to be in good health because they are thought to face few pressures, but information is lacking to confirm this. Only a few areas, however, can be regarded as pristine, because pressures such as nutrient loading, pollution with persistent chemicals, and fishing, have affected nearly all parts of Australia's marine and estuarine ecosystems. Near many cities and other parts of the coastline the condition of some habitats (particularly seagrasses, saltmarshes, and mangroves), is poor. There is increasing concern at the state of estuaries,

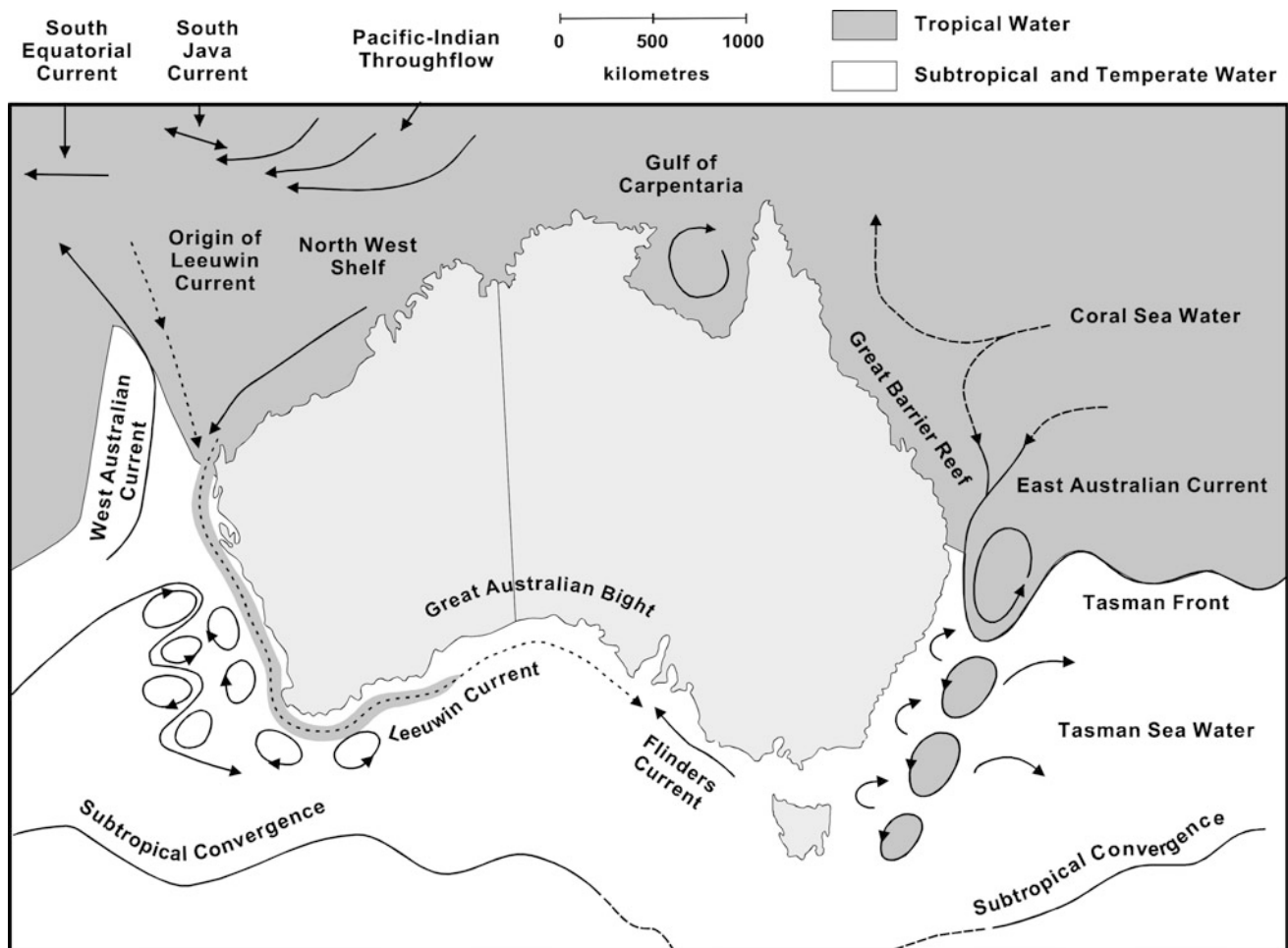
with assessments of declining water quality, threats to estuarine fisheries and other impacts in many areas. Most rivers, estuaries, and coastal waters near Australia's large population centers show signs of eutrophication. (Commonwealth of Australia 2001) Blooms of toxic marine algae, some species of which may have been introduced from other countries in ships' ballast waters, are periodic, serious problems in parts of southern Australia.

Physical Setting

Oceanography

The waters surrounding Australia (and New Zealand) are linked to three of the large ocean basins of the Southern Hemisphere, the Pacific, the Indian, and Southern Oceans. The southwest Pacific Ocean, which borders the eastern margins of Australia, is dominated by the large anticlockwise circulation of the South Pacific. The elements of this gyre consists of the western-flowing South Equatorial Current,

which after crossing the tropical Pacific, deflects southwards to form the East Australian Current, and then turns eastward. South of Australia is the massive Antarctic Circumpolar Current or "West Wind drift". On the western margins of Australia, the Indian Ocean South Equatorial Current, driven by trade winds and westerlies, has a similar anti-clockwise circulation, flowing towards the east coast of Africa and merging with the Agulhas Current, the strongest western boundary current in the Southern Hemisphere. The subtropical convergence and the subpolar convergence zones isolate tropical and subtropical waters of Australasia from the nutrient-rich waters of the Southern Ocean, although some mixing of subantarctic waters does occur at depth in the south Tasman Sea (see Fig. 1). For the most part, the southern coast faces the full force of the Southern Ocean and as such, experience some of the highest wave energies in Australia. The El Nino–Southern Oscillation (and El Nino/La Nina cycles) has a major influence on prevailing current systems around Australia, dramatically affecting water temperatures and marine life, including major pelagic fisheries.



Australia, Coastal Ecology, Fig. 1 Major oceanographic features of Australia

Both sides of the island continent of Australia experience poleward-going warm currents, which act as conduits to bring tropical ocean waters with their entrained plants and animals to southern Australian waters. On the east is the East Australian Current (EAC) (a western boundary current) and on the west is the Leeuwin Current, which flows southward along the outer shelf to the southwest corner of the continent, and then runs eastward along the southern margin (Cresswell 1991). The EAC flows south from the Coral Sea (Church 1987) close to the continental slope, until central New South Wales (NSW) where it runs further offshore (Cresswell et al. 1983). Once or twice a year the EAC turns into loops in the Tasman Sea off the NSW coast. These loops become detached from the current and form EAC eddies, warm disc-shaped water parcels having a diameter of 200–300 km, and extending 1500–2000 m into the ocean (Nilsson and Cresswell 1981).

On the west coast of Australia, the Leeuwin Current is unique. Unlike other eastern boundary currents (such as the California, Benguela, and Peru Currents and the currents off North Africa), the Leeuwin Current flows southward (rather than equatorial) and is associated with warm water incursions and downwelling (rather than cold water and upwelling) (Cresswell and Golding 1980; Bye 1983). The Leeuwin Current is primarily driven by thermosteric effects associated with the flow of extremely warm waters from the Pacific to Indian oceans (via the Indonesian Archipelago). This strongly affects shelf circulation along the entire west coast (and to a lesser extent, south coast) and has profound effects on marine life and also, fisheries (i.e., Southern Bluefin tuna, western rock lobster, mackerel, Horse mackerel, Australian salmon, Australian herring) (Lenanton et al. 1991). The warm low-salinity waters of the Leeuwin Current are thought to be responsible for the distribution, migration, and patterns of dispersal of many pelagic marine organisms from the warm waters of the northwest of Australia to the southern seaboard. This is evidenced by a unique “tropical” or Indo-Pacific element, both in the demersal and pelagic fauna of the southern coast (Maxwell and Cresswell 1981).

Wind-driven continental shelf waves have been identified as a primary source of current variability in practically all Australian continental shelf waters, such as the Great Barrier Reef (Cahill and Middleton 1993), the New South Wales shelf, the Great Australian Bight (Provis and Lennon 1981), Bass Strait (Middleton and Viera 1991) and the North West Shelf (Webster 1985). With topographic scales of several hundred kilometers the flows are strongly controlled by the coastline and somewhat insulated from the open ocean, and can be extremely complex (e.g. Gulf of Carpentaria, Bass Strait). Between November and May cyclones are a common feature of Australian tropical waters, and can generate strong currents and high mean sea levels, as well as high seas.

Isolation from open ocean waters allow some water masses of distinctive properties to form, particularly in embayments

and gulfs. Inverse estuaries (i.e. estuaries which are more saline at the head, rather at the mouth of the estuary) are a characteristic feature of Australia’s coasts, particularly in semi-arid regions, due to Australia’s low rainfall and variable climate. In the upper reaches of Spencer Gulf and Gulf St. Vincent (in South Australia), low runoff coupled with high evaporation rates result in salinities exceeding 48 psu and 42 psu, respectively (Nunes and Lennon 1986; de Silva Samarasinghe and Lennon 1987). Similarly, the limited depth of Bass Strait (approximately 100 m), results in greater cooling of waters in winter (compared with the deeper adjacent ocean), and subsequent cascade down the continental slope into the Tasman Sea (Godfrey et al. 1980; Villanoy and Tomczak 1991).

Nutrients

Australian coastal waters are generally low in nutrients (especially nitrate and phosphate), and hence, are low in phytoplankton and productivity. The oligotrophic nature of Australia’s coastal waters is due to both, continental and oceanographic features. Oceanographic factors that contribute to the low nutrient status include: the isolation of Australia’s coastal waters (i.e., Great Australian Bight) from rich subantarctic waters to the south; the lack of a nutrient-rich eastern boundary current (and associated upwelling) in the Indian Ocean (originating in high latitudes) – but rather, warm water, poleward currents that bring tropical, nutrient-depleted waters along the continent’s east and west coasts (i.e., the East Australian Current and Leeuwin Current); the dominance of large areas by subtropical waters with limited nutrient reserves down to 100–200 m; and the lack of major sources of upwelling of nutrients along the Australian coastline. In addition, wind-induced upwellings, which give enhanced productivity to Equatorial Pacific and Equatorial Atlantic waters, are largely ineffective in the tropical waters of Australia, because of the confused array of closely positioned land masses and islands and their confined seas to the north. Several continental factors also contribute to general nutrient impoverishment. These include: the low nutrient status (particularly in phosphates) of the well-weathered, coastal-drained soils in many areas of Australia; low coastal relief or topography; and low run-off (due to the low and high variable rainfall in Australia).

Australia’s marine and estuarine communities have evolved in a generally low ambient nutrient environment, but with in some places a highly variable, nutrient supply, due to the great variability in the climate – particularly the variability in runoff from the land. This may account for the apparent sensitivity and large-scale decline of some Australian biota (such as seagrasses) to nutrient enrichment of estuarine and coastal waters. Following European settlement, the catchments of Australian rivers are now, poorly vegetated, and low rates of infiltration of rainfall result. Thus, runoff from Australian rivers

is more highly pulsed with long droughts interspersed with major flooding events. This flooding typically leads to turbidity and nutrient plumes into coastal waters.

Much of Australia's continental shelf and coastal marine environment lies within the tropics. At higher latitudes, water temperature, levels of wind mixing and inputs of light necessary for photosynthesis go through pronounced seasonal cycles (Harris et al. 1987, 1991; Clementson et al. 1989). Because of the annual winter replenishment of nutrients, temperate oceanic and continental shelf regions tend to support larger seasonal blooms of algae and greater production at higher trophic levels than do (non-upwelling) tropical ocean and shelf systems. However, there are no large seasonal blooms producing surpluses of organic matter. As a result, Australia lacks the large demersal fisheries that characterize Northern Hemisphere continental shelf systems. Lacking a well-defined seasonal cycle, biological variability in tropical ecosystems is more closely related to disturbance events and oceanographic processes such as upwelling, floods, tidal mixing, and cyclones.

Nutrient enrichment occurs in several isolated areas around Australia. These include persistent upwellings along much of Australia's eastern seaboard (along the New South Wales coast between Port Stephens and the Queensland border) and at sites along the southern coast of Victoria (Gippsland coast) and South Australia (southern Eyre Peninsula and between Port Macdonnell and Robe), bottom coastal enrichment off New South Wales, cyclonic eddies off eastern Tasmania (southwest Tasman Sea), convective overturn (off Tasmania), North West Shelf enrichment, and tidally induced enrichments of the Great Barrier Reef waters (Wenju et al. 1990; Furnas 1995). Inshore southern habitats under the influence of localized nutrient-rich, cold water, seasonal upwellings, have resulted in very high levels of reef biodiversity (particularly among the macroalgae and invertebrates) and productivity. Off southern Eyre Peninsula, the region is also a key area of pilchard abundance (Ward et al. 1998) and one of the most important sites for seabirds and marine mammals in temperate Australia (Edyvane 2000).

Continental Shelf and Shelf Processes

Australia's continental shelf covers approximately 2 million km², an area equal to about 25% of the continent's land surface. It is shallow, generally less than 200 m in water depth, and in vertical profile is a smooth, flat surface, which dips gently seawards, with some parts rimmed by shelf edge barrier reef systems (e.g., the Great Barrier Reef shelf). In a global context, the depth of the shelf break (20–550 m water depth and defined as 200 m by international convention) and the width (2–1500 km) exhibit a wide variability. The shelf break is located about 10 km offshore from Fraser Island (east coast) and North West Cape at Exmouth (west coast), but over 500 km distance offshore on the Arafura Shelf in the north.

The continental shelf is comprised of granitic crustal material – unlike the deep ocean bed which is underlain by basalt. During Pleistocene periods of lower relative sea levels, the shelf was exposed subaerially and the Australian mainland was “joined” by dry land to several of the adjacent large islands such as Tasmania and New Guinea. Such “land bridges” are considered to have facilitated the migration of animals and humans in the late Pleistocene “Ice Age”. The shallow seas comprising the Sahul Shelf, Gulf of Carpentaria, Torres Strait, and Bass Strait were thus subjected to erosion and sedimentation by rivers and wind on several occasions during the past 150,000 years. The basins in Bass Strait, Bonaparte Gulf, and the Gulf of Carpentaria are considered to have been the sites of large fresh to brackish water lakes and lagoons during these periods of emergence.

Coastal sedimentation processes are dominated by: (1) swell waves and storm currents (80% of the world's continental shelves); (2) tidal currents (17%); or (3) intruding ocean currents (3%). Currents produced during storm events – either as tropical cyclones or temperate storms – dominate in the erosion and transportation of sediment over 82% of the Australian shelf surface area. Detrital and organic sediment (i.e. biogenic) sediment is generally dominant on the continental shelf, with carbonate content greatest in offshore sediments (Harris 1995). In southern Australia, the inshore non-calcareous sediments are generally dominated by coarse-grained relict (i.e., sediment deposited in a previous low sea environment) or reworked quartz sand. In contrast, the inner shelf, non-calcareous sediments of tropical regions are commonly terrigenous silts and clays supplied by the large river systems of northern Australia (e.g., Great Barrier Reef region) (Belperio and Searle 1988).

Sediments on Australia's southern continental shelf are arranged in zones parallel to the coast (reflecting the dominance of ocean swell and storm generated currents and long period swell waves) and represent some of the largest modern, cool-water, open shelf accumulations of carbonate sediments in the world (Connolly and von der Borch 1967; Wass et al. 1970; Gostin et al. 1988; James and von der Borch 1991). These include the Otway region of southeastern South Australia (i.e., Lacedpede and Bonney Shelf) (James et al. 1992) and western Victoria (Otway Shelf) and Great Australian Bight (i.e., Eucla Platform) (James and Bone 1992), the Rottneest Shelf, and the New South Wales Shelf (Davies 1979). While hermatypic corals and the calcareous green alga *Halimeda* characterize tropical shelf carbonate sediments (north of 24°S), bryozoa or “lace corals” and coralline algae dominate the cooler water sediments along the southern margins of Australia (south of 38°S) (Gostin et al. 1988).

The sediment loads supplied to the coastal zone by many rivers are trapped in estuaries, hence, little or no sediment may reach the deeper waters of the continental shelf (particularly along the southern shelf). River catchments in the seasonally

wet northern parts of Australia generally have a higher sediment yield (on the order of 100–300 t/km²/yr) than those in the south (on the order of 10–30 t/km²/yr) (Harris 1995). Export of continental terrigenous sediment to the wide southern shelf is low because of the low continental relief and lack of significant drainage to the sea, particularly in the Great Australian Bight (i.e., from southwestern Australia, to the mouth of Murray River in central South Australia). Within sheltered areas of low coastal relief, redistribution of coastal sediments result in shallow coastal embayments, dominated by tidal salt marshes, mangroves, and sea grass (e.g., Port Phillip Bay, Spencer Gulf, Gulf St. Vincent), while in exposed coastal areas of high wave energy (e.g., the Great Australian Bight), these reworked sediments supply sediment for extensive beach and dunal systems which dominate these regions (Gostin et al. 1988).

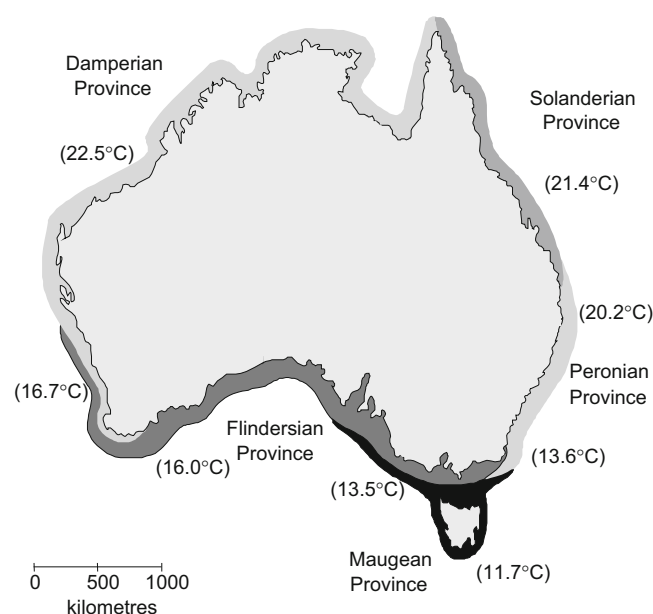
Tropical cyclones, associated with atmospheric low-pressure systems, are the cause of storm events in much of northern Australia. The sections of the Australian shelf most frequently affected by cyclones are the North West Shelf with up to 25 cyclones per decade and the Great Barrier Reef with up to 15 cyclones per decade. Tropical cyclones induce strong currents that erode and transport sediment over a wide area. Modeling studies on the North West Shelf (Hearn and Holloway 1990) have shown that, under the influence of tropical cyclones, strong westward flowing coastal and inner shelf currents are established between the eye of the cyclone and the coast. Such cyclone-induced currents are clearly a significant factor affecting sediment movement on cyclone-dominated shelves, but they may also influence the long-term (net) sediment movement on some otherwise tidally dominated sections of the shelf.

Tidally dominated shelves occur generally where the mean spring tidal range measured along the coast exceeds 4 m (macrotidal). Around Australia, tidal ranges greater than 4 m occur along the northwestern coastline between Port Hedland and Darwin and reach a maximum range of about 9.2 m in King Sound. The southern Great Barrier Reef is also macrotidal, with a tidal range of 8.2 m in Broad Sound. Tidal currents are also an important sand-transporting agent in mesotidal (2–4 m tidal ranges) areas, such as Torres Strait, Moreton Bay (Queensland), and Bass Strait. Tidal currents are able to dominate sand transport in microtidal areas (tidal range less than 2 m) in restricted cases where coastal geometry affords shelter from ocean generated swell and wind-driven currents. Such is the case in many bays (e.g., Shark Bay), the approaches to some major ports (e.g., Port Phillip Bay) and in partly enclosed gulfs (e.g., Spencer Gulf, Gulf St. Vincent, and the Gulf of Carpentaria). Sediment transport and dispersal controlled by intrusive ocean currents affects only a little more than 1% of the Australian continental shelf. The only location where such currents are known to dominate sediment movement around Australia is the shelf offshore from Fraser Island, where the southward-flowing East Australian Current intrudes.

Coastal Marine Ecosystems and Habitats

Marine Ecosystems

The marine ecosystems of Australia has been the subject of several biogeographic studies and classifications in recent decades (Womersley and Edmonds 1958; Womersley 1959, 1981a, b, 1990; Bennett and Pope 1960; Knox 1963; Markina 1976; Edgar 1984). The biogeographical regions or provinces identified are traditionally regarded as coastal regions characterized by a relatively distinct and homogeneous flora and fauna, with only a small percentage of species common to adjacent regions, and usually differing in sea temperatures from adjacent provinces by more than 5 °C. Many of these classifications have, until recently, been based on a knowledge of the distribution of intertidal biota (rather than on assessments of whole floras and faunas). Traditionally, the tropical and subtropical marine environments of Australia, i.e. >20 °C, have been classified into three major biogeographical regions or provinces: the Damperian Province (from the Houtman Abrolhos, 28°50'S, off Western Australia, around northern Australia to Torres Strait); the Solanderian (or Banksian) Province (the mainland coast from Torres Strait, south to about 25°S in southern Queensland; and the Great Barrier Reef Province (comprising the Great Barrier Reef and associated islands) (see Fig. 2). However, the widespread distribution of most pan-tropical Indo-Pacific algae suggests that the whole of the tropical Australian coast is probably best regarded as one province, with possibly sub-provincial regions (Womersley 1990).



Australia, Coastal Ecology, Fig. 2 Marine biogeographical provinces of Australia, and average surface sea temperatures. (From Edyvane 1999; courtesy of SARDI)

Australia, Coastal Ecology, Table 1 Coastline and coastal types as a percentage of length of coastline around Australia. Percentages remaining from the total is coral reefs and estuary mouths. (Adapted from Fairweather and Quinn 1995)

State/territory	Coastline		Coast type				Total
	(km)	Percentage of total	Rock %	Sand %	Mud %	Mud/mangrove %	
Queensland	13,347	22.3	4	36	1	19	59
Northern Territory	10,953	18.3	9	40	6	48	97
Western Australia	20,781	34.8	19	49	2	19	87
South Australia	5067	8.5	33	59	1	8	100
Victoria	2512	4.2	26	64	7	9	99
Tasmania	4882	8.2	29	68	2	2	99
New South Wales ^a	2194	3.7	33	65	1	2	100
Total	59,736	100	18	47	2	19	84

Source: Australian Land Information Group (AUSLIG) 1:100,000 Coastline Database (1993)

^aIncludes Jervis Bay Territory (57 km)

The temperate marine environments of Australia have traditionally been classified into three provinces: the Flindersian Province, which encompasses the warm–cool temperate (i.e., 12–20 °C) waters of southwest Western Australia, along the southern coast of Australia, to southern New South Wales, including the waters of South Australia, Victoria and Tasmania; the Maugean Province (or Subprovince of the Flindersian), which encompasses the cold temperate (i.e., 10–15 °C) biota of Tasmania and Victoria and, to a lesser extent, South Australia (east of Robe) and New South Wales (south of Eden); and the Peronian Province, which encompasses the warm temperate (15–20 °C) biota of the southern coast of Queensland, down the New South Wales coast to the Victorian border (Bennett and Pope 1960; Knox 1963; Womersley 1981a, b; Edgar 1984). The biota of Tasmania comprise elements of all three southern Australian marine biogeographic provinces (i.e., Peronian, Maugean, and Flindersian) (Edgar 1984).

In recent years, biogeographic or regional ecosystem classifications have been developed for many coastal and near-shore marine environments around Australia (Muldoon 1995). In several states, this has involved utilizing analytical multivariate procedures to classify patterns in nearshore ecosystem diversity. Many of these regionalizations have now been integrated as part of a national effort to develop an “*Interim Marine and Coastal Regionalisation of Australia*” (IMCRA) to assist in the development of a representative system of Marine Protected Areas (IMCRA 1998). To-date, a total of two pelagic and nine demersal provinces (and biotones) are recognized for the continental shelf (out to the limit of the Australian EEZ), based on physical oceanography, seafloor topography and the distribution of pelagic and demersal fish (CSIRO 1996a, b; IMCRA 1998). At a finer scale, a total of 60 mesoscale bioregions have been identified for the nearshore marine environments of Australia, based on a wide range of physical and biological descriptors, such as climate, oceanography (water temperature, wave energy), tidal range, coastal geomorphology, and biology (habitats, marine mammals, endemic species) (IMCRA 1998).

Coastal and Marine Habitats

The coastal habitats of Australia are largely influenced by the interplay of physical factors, particularly oceanography, climate, and coastal geology. As such the inshore environments of Australia range from low-energy muddy tidal deltas, which dominate the tropical north, to the swell-dominated rocky coasts and sandy shores of the temperate south. Muddy sediments dominate the tropical environments of Australia, either under mangrove forests or as wide, near-flat shores on macro-tidal coasts. While rocky shores are more widespread in temperate marine environments (particularly in the Great Australian Bight and Tasmania), occupying approximately 30% of the coastline (cf. approximately 11% in tropical areas), sandy beaches dominate the temperate coasts of Australia (see Table 1). While sandy beaches are common all around Australia, they occupy approximately 50–70% of the temperate coastline, particularly along the eastern and western coast of Australia – regions swept by Eastern Australian Current and the Leeuwin Current, respectively (Fairweather and Quinn 1995). Coral reefs occupy 14% of Australia’s coastal habitat and are essentially restricted to tropical regions.

Australian estuaries occur over a wide range of geological and climatic conditions and consequently display a great variety of form (Saenger 1995). Australia has 783 major estuaries; 415 in the tropics, 170 in the subtropics, and 198 in temperate areas (Table 2). Estuaries are most common in the humid tropics. There are few estuaries on the long arid coastline in the southwest and west. In the semi-arid and arid regions of Australia, low rainfall and variable climate result in irregular freshwater inputs and prevalence of “inverse estuaries”, that is estuaries that generally flow (and flood) only after local rains have fallen. Gulf St. Vincent and Spencer Gulf represent the largest temperate inverse estuaries in Australia, while Shark Bay and Exmouth Gulf (in Western Australia) are the largest tropical estuaries. Naturally rich in nutrients, estuaries are ecologically highly productive and are important habitats for adult marine fauna as well as critical nursery habitats for the juveniles of many species.

Australia, Coastal Ecology, Table 2 Coastal habitats around Australia. (Adapted from Galloway 1982; Saenger 1995; Kirkman 1997; Edyvane 1999)

State/ territory	Coastal habitat		Number of estuaries
	Mangrove (km ²)	Seagrass (km ²)	
Queensland	4602 (39.8%)	22,300 ^a (40.1%)	307
Northern Territory	4119 (35.6%)	900 ^b (1.6%)	137
Western Australia	2517 (21.8%)	22,000 (39.6%)	145
South Australia	211 (1.8%)	9620 (17.3%)	15
Victoria	12 (0.1%)	100 (0.2%)	35
Tasmania	0	500 (0.9%)	63
New South Wales	99 (0.9%)	155 (0.3%)	81
Total	11,558	55,575	783

^aIncludes Torres Strait (18,000 km²)

^bComprises Gulf of Carpentaria (900 km²)

Coastal saltmarshes. Coastal saltmarshes in Australia are intertidal plant communities dominated by herbs and low shrubs, which are often associated with estuaries. Although there is a high degree of endemism at the species level in the saltmarsh flora of Australia, at generic and family level there is a strong similarity (in structure and composition) between Australian saltmarshes and those elsewhere in the Southern Hemisphere and also in the Northern Hemisphere (Adam 1990). When viewed in a world context, Australian saltmarshes are not as distinctly “Australian” as are the various terrestrial communities of the continent (Adam 1995).

Australia has 13,595 km² of coastal saltmarshes, with the most extensive areas occurring in the tropics (Bucher and Saenger 1991). On the continental scale, there is a striking trend of increased species richness and community complexity with increasing latitude (Saenger et al. 1977; Specht 1981; Adam 1990). Saltmarshes in northern Australia, although extensive, are very species poor (frequently *S. virginicus* – dominated), with frequently less than 10 vascular plant species in their total flora. Compared to northern Australia, saltmarshes in southern Australia (i.e., Victoria and Tasmania), which cover much smaller areas, frequently support at least three times as many species (Adam 1995). Not only is the flora of individual sites much greater at higher latitudes, but the regional saltmarsh flora is also much higher.

In northern Australia both mangrove and saltmarsh may be restricted to the lower, more frequently flooded intertidal zone while the upper intertidal takes the form of extensive hypersaline flats with only very sparse and localized vegetation. In southern Australia a biogeographic distinction can be drawn between saltmarshes on arid or seasonally arid (Mediterranean climate) coasts and those on temperate coasts with relatively high rainfall (Bridgewater 1982; Adam 1990).

On drier coasts saltmarsh vegetation is characterized by a diversity of succulent, shrubby chenopods with a tendency to more open vegetation towards the upper tidal limit, while on wetter shores vegetation is denser with more grassland and sedgeland communities. Much of the primary production of saltmarshes is utilized in detrital pathways, which include microbially mediated decomposition, and take place either within salt marshes or in adjacent waters. The assumption of high productivity is linked with the “outwelling hypothesis”, by which coastal wetlands act as a source of detritus and nutrients exported to offshore. Data on the linkages of saltmarshes to adjacent waters in Australia are lacking (Adam 1995).

Saltmarshes provide habitat for numerous organisms of both terrestrial and marine origin. The most important fish habitats in intertidal wetlands are seagrasses and mangroves: Australian saltmarshes, unlike those overseas, generally have few permanent creeks and pans and so provide few fish habitats. Nevertheless, studies indicate that at least some saltmarshes may provide habitats utilized by fish (Morton et al. 1987; Connolly et al. 1997). Coastal wetlands are an important habitat for birds. The number of species breeding in saltmarshes is small, but the upper marsh vegetation provides nest sites for some species. A large part of the population of one of the rarest species in Australia, the orange-bellied parrot (*Neophema chrysogaster*), overwinters on salt-marshes in Victoria where it feeds on the seeds of chenopods. Migratory waders feed largely on invertebrates in intertidal sand and mudflats but saltmarshes may provide secure high tide roosts. Little is known about the utilization of saltmarsh by other faunal groups.

Mangrove forests. Australia has the third largest area of mangroves in the world (approximately 11,500 km²) and some of the most diverse communities (Galloway 1982). The mangrove flora of Australasia (the area including New Guinea, New Caledonia, Australia and New Zealand) is one of the richest in the world, having approximately five times the species richness of all other regions excepting the neighboring region of Indo-Malesia (Duke 1992). Generally, the greatest concentration of mangrove species in Australasia is found in southern New Guinea and northeastern Australia, where 45 taxa of mangrove plants are shared (Robertson and Alongi 1995). The most recent review of mangrove species in Australia (Duke 1992) recognized 39 taxa of mangrove plants belonging to 21 genera and 19 families. The taxa include at least four rare hybrids of more common species. Only one species, the newly discovered *Avicennia integra*, appears to be endemic to Australia. All other species are widely distributed on the island of New Guinea or in Southeast Asia.

Mangroves are generally associated with low energy, muddy coastlines, particularly tropical deltas. However, they can grow on a wide variety of substrates including sand, volcanic lava, or carbonate sediments (Chapman

1976). Mangrove environments dominate the tropical coastal habitats of northern Australia, occupying approximately 11,228 km² (cf. 333 km² in southern temperate Australia) (see Table 2). Mangrove forests show their greatest development on humid, tropical coastlines where there are extensive intertidal zones (as found on low gradient or macrotidal coasts), and abundant fine-grained sediment, high rainfall and freshwater runoff from catchments. However, there are large local forests in the subtropics and as far south as Corner Inlet in Victoria at 38°S. The temperate marine environments of Australia are generally characterized by a lack of mud, tidal delta, and mangrove habitats, apart from large sheltered embayments (such as Port Phillip Bay, Spencer Gulf and Gulf St. Vincent). The largest areas of temperate mangrove forest occur in South Australia (approximately 332 km²), largely within the sheltered gulfs of Spencer Gulf and Gulf St. Vincent (Galloway 1982). Similarly, species richness in mangrove communities is greatest in the northeast humid tropics and there is a gradual decrease in the number of species with increase in latitude (Robertson and Alongi 1995). Some estuarine systems on Cape York have up to 35 species of mangrove plants while to the south (Victoria and South Australia) only one species, *Avicennia marina*, occurs.

Variations in air temperature, rainfall, river runoff, sediment type (and deposition rate), tidal amplitude, and geomorphology along the Australian coast produce a variety of regional coastal settings for mangroves to grow (Robertson and Alongi 1995). Mangrove communities vary from tall (up to 40 m), highly productive, closed canopy forests dominated by species of the Family Rhizophoraceae (particularly species of *Rhizophora* and *Bruguiera*) in the high rainfall, humid tropics of northeastern Queensland and the Kimberley coast of Western Australia – to the open woodlands of *A. marina* – which dominate the mangrove habitats at latitudes greater than 30°S. As conditions become more arid within the tropics (such as the Pilbara coast in Western Australia), water and salinity stress on mangroves increase in the intertidal zone and species richness decreases, and open canopy woodlands or short (1–5 m), and low productivity open shrublands develop (Semeniuk et al. 1978).

Understanding of mangrove food webs in Australia currently only refers to tidally driven (non-estuarine) tropical mangrove forests (of north Queensland). These studies suggest that the carbon budget does not follow the classic overseas models of mangrove food chains – with high sediment bacterial biomass and production (sediment bacteria are a significant sink for carbon fixed by mangrove vegetation) greater than the sum of litter fall and dead wood turnover (Alongi 1990). Meiofaunal and macrofaunal densities are low in mangrove forests, possible due to low food quality (i.e., very low nitrogen content) of mangrove detritus. Elements of the macrofauna, particularly decapod crustaceans (such as

grapsid crabs of the subfamily Sesarminae), play major roles in food chains and nutrient transformations (i.e. retention of nutrient litter and hence, total tidal export of materials from forests) in intertidal tropical forests (Robertson 1991). Little work has been done on other mangrove forest types and current models are unlikely to apply to the range of forest types in the different regions of Australia.

Mangrove ecosystems are important habitats for nekton. Fish species of the families Ambassidae, Clupeidae, Engraulididae, Gobiidae, and Leiognathidae are numerically dominant in mangrove waters, while species of Sparidae, Haemulidae, Lutjanidae, Carcharhinidae, Centropomidae, and Carangidae contribute significantly to the biomass of mangrove fish communities. Many of the dominant fish families occur in mangroves only as juveniles and recruitment from offshore waters is dependent on a number of factors, not least of which is the strength of the wet season and thus the prevailing inshore seawater salinities (Robertson and Duke 1990). While recruitment factors are important, the variation in forest types and the availability of food and microhabitats are important factors in controlling spatial variation in fish densities and community structure (Robertson and Blaber 1992).

There can be significant export of detritus from mangrove habitats to nearshore receiving waters. The importance of mangrove detritus in supporting higher consumers (such as fish and prawns) in estuarine and coastal areas of Australia has been confirmed. For instance, the zoea larvae of sesarimid and ocypodid crabs from mangrove forests are a major food source for juvenile fish in north Queensland estuaries. The juveniles of some penaeid prawns feed on mangrove detritus or on meiofauna that is mangrove dependent.

Some of Australia's most important single species commercial and recreational fisheries are directly linked to mangroves during at least part of their life cycles, such as the banana prawn (*Penaeus merguensis*), bream (*Acanthopagrus australis* and *Acanthopagrus berda*), grunter (*Pomadasy kakaan*), barramundi and mangrove jack (*Lutjanus argentimaculatus*) (Robertson and Alongi 1995). Other commercial fisheries have more indirect but equally important connections to mangroves. In the wet tropics of northeastern Australia, the seagrass meadows on which juvenile tiger prawns (*Penaeus esculentus*) depend often occur immediately seaward of mangrove forests (Robertson and Lee Long 1991). Mangroves in Australia are also important nursery grounds for a variety of small non-commercial fish species, which are an important food source for carnivorous estuarine fishes such as barramundi and trevallies and jacks (Carangidae), and offshore billfish and mackerel.

As elsewhere in the world, mangroves play a major role in coastal protection in Australia. Mangroves in northern Australia are likely to play a major role in ameliorating the effects of cyclones and storm surges on nearby urban centers and

low-lying agricultural lands. Mangrove forests are also major sites of trapping and stabilization of sediments derived from river catchments.

Seagrass beds. Seagrasses are productive, widespread, and ecologically significant features of Australia's tropical and temperate nearshore environments (Larkum et al. 1989). Australia has the highest biodiversity of seagrasses in the world, the largest areas of temperate seagrass and one of the largest areas of tropical seagrass. Australian seagrasses are characterized by high speciation and also, endemism, especially in temperate regions. Australia has over 30 species of seagrass, which can be broadly categorized into tropical and temperate groups (Kuo and McComb 1989; Kirkman 1997). On the basis of species composition and distribution a transition zone occurs at approximately 30°S (Moreton Bay) on the east coast and 25°S (Shark Bay) on the west coast (Larkum and den Hartog 1989).

Of the 55,500 km² of seagrass meadows in Australia, the largest meadows occur in tropical Australia, in Torres Strait (17,500 km² or 31.5%) and Shark Bay (13,000 km² or 23.4%). Within temperate waters, seagrass meadows are largest and most diverse along the southern coast of Western Australia (9000 km² or 16.2%) (Poiner and Peterken 1995), and Spencer Gulf (5520 km²) and Gulf St. Vincent (2440 km²) in South Australia (Edyvane 1999). In contrast, seagrass abundance (and diversity) is low in southeastern Australia, where the high-energy coast line restricts seagrass to estuaries and protected bays. For instance, seagrass occupies approximately 500 km² (or 0.9%) in coastal Tasmanian waters, 150 km² (or 0.3%) in the waters of New South Wales and 100 km² (or 0.2%) in Victoria waters (see Table 2).

Australia's seagrasses can be divided into those with temperate and those with tropical distributions, with Shark Bay on the west coast and Moreton Bay on the east coast located at the center of the overlap zones. Temperate species are distributed across the southern half of the continent, extending northwards on both the east and west coasts. These species have been studied most extensively, particularly the large genera *Amphibolis*, *Posidonia*, and *Zostera*, although species remain that have been little studied. The highest biomasses, and highest regional species diversity, occur in southwestern Australia, where seagrasses occur inside fringing coastal limestone reefs, or in semi-enclosed embayments. Large seagrass meadows are present in protected areas across the Great Australian Bight, and into South Australia and Tasmania. Along the New South Wales coastline, seagrasses are confined to estuaries such as Botany Bay, except for the large sheltered embayment of Jervis Bay (Kirkman 1997).

In northern Australia, seagrass species possess tropical affinities, for example, *Thalassia* and *Cymodocea*. Tropical beds can be of high diversity, but generally possess lower biomasses than in temperate zones. While large areas of

seagrasses occupy embayments such as Hervey Bay, Queensland, tropical seagrasses are generally confined to intertidal environments, or to deep water (Lee Long et al. 1993). The genera *Halophila* and *Halodule* extend beyond the tropics into cooler waters. In the tropics these genera are heavily grazed by turtles and dugongs (Lanyon et al. 1989). Grazing of tropical seagrasses by dugongs and sea turtles provides another unique aspect of tropical Australian seagrass meadows. Caribbean and Indo-Pacific seagrass meadows are characterized by the absence or depletion of macrograzers (i.e., manatees, sea turtles, dugongs), due to overfishing. Hence, tropical Australian seagrass meadows are probably more representative of the natural conditions in which tropical seagrasses evolved.

Seagrass growth is confined to estuaries and protected embayments in southeastern Australia, due to the high-energy coastline. Although a generalized north-to-south distribution pattern can be given, the distribution and dominance of the three sea grass genera (*Zostera*, *Halophila*, and *Posidonia*) is dictated by the occurrence and nature of those coastal features, which offer a suitable growth habitat for them.

Unlike temperate seagrass species, tropical seagrasses tend to occur in mixed-species stands. Thirteen species from seven genera are found across northern Australia. The region has a greater diversity of seagrass species and communities than elsewhere in the Indo-Pacific. Torres Strait supports one of the largest seagrass areas in Australia. A total of 17,500 km² of seagrass supporting habitat associated with 295 km of coastline or reef has been identified and mapped. The Gulf of Carpentaria, a shallow marine embayment, supports 906 km² of seagrass habitat along 671 km of coastline. The northwestern coast of Australia is the least well-known areas of northern Australia. Tropical seagrass meadows directly support dugong (*Dugong dugon*) and green sea turtles (*Chelonia mydas*) (Lanyon et al. 1989).

Seagrass communities are of considerable importance in the processes of coastal ecosystems because of their high rates of primary production and their ability to trap (and stabilize) coastal sediments and organic nutrients. Their importance to commercial and recreational fisheries is well documented (Cappo et al. 1998; Butler and Jernakoff 1999). Additionally, seagrasses are important in substrate stabilization, supply and fixation of biogenic calcium carbonate, detrital food chains, nutrient cycling and as substrate for epibiota, and as critical habitats for many species (Larkum et al. 1989).

Rocky reefs. Rocky shores are the main hard substrata of headlands along the open coasts and sometimes within estuaries. Rocky reefs occur along the entire length of the Australian coastline, mostly within shallow ocean waters. However, they are more widespread in temperate marine environments (particularly in the Great Australian Bight and Tasmania), occupying approximately 30% of the coastline (cf. approximately 11% in tropical areas) (Table 1). Prominent areas of reef occur seaward of most headlands and rock

platforms. Significant amounts of subtidal rocky reef are also found in deeper waters offshore, around islands, and in estuaries; among the latter, in drowned river valleys.

Species diversity is particularly high in temperate rocky reef habitats (Keough and Butler 1995). The southern coastline has the world's highest diversity of red and brown algae (around 1155 species), bryozoans (lace corals), crustaceans, and ascidians (sea squirts). Diverse assemblages of brown, red, and green macroalgae, along with sponges, ascidians and other sessile invertebrates enhance habitat complexity and provide many opportunities for specialization (e.g., Jones and Andrew 1990; Lincoln Smith and Jones 1995). Furthermore, the large macroalgae (such as kelp), that partially cover most rocky reefs, enhance overall species diversity by providing patches of shaded habitat favored by distinct assemblages of organisms (Kennelly 1995).

The intertidal ecology of Australian rocky shores (particularly temperate coasts) has been relatively well described (Womersley and Edmonds 1958; Womersley and King 1990). In contrast, very few studies have been conducted on tropical rocky shores. Similarly, the subtidal hard substrata or reefs of Australia are little known. It has only been since the advent of SCUBA apparatus in the 1960s, that detailed subtidal reef studies have been conducted and these have largely been restricted to temperate reefs (e.g., Shepherd and Womersley 1970, 1971, 1976, 1981; Shepherd and Sprigg 1976; Shepherd 1983; Edgar 1984; Jones and Andrew 1990; Underwood and Kennelly 1990).

The ecology of subtidal temperate reefs in Australia is characterized by the structural dominance and diversity of large macroalgae and an abundance of sessile and mobile invertebrate assemblages (i.e., sponges, bryozoans, ascidians, hydroids, echinoderms, molluscs, crustaceans) (for recent views, see Underwood and Chapman 1995; Keough and Butler 1995). In the cold temperate waters of Australia (east of Robe, South Australia), shallow, wave-exposed nearshore reefs are dominated by large canopy-forming species of brown macroalgae, such as the Giant Kelp (*Macrocystis*), *Phyllospora comosa* and the just subtidal, Bull Kelp (*Durvillea potatorum*) (Womersley 1981a, b; Womersley and King 1990; Underwood and Kennelly 1990, 1991). In warm temperate waters (west of Robe), wave-exposed nearshore reefs are dominated by smaller (i.e., up to 2 m high), canopy-forming brown macroalgae, such as *Ecklonia radiata*, and fucoid algae (such as *Seirococcus axillaris*, *Scytothalia dorycarpa*, and species of *Cystophora* and *Sargassum*). In deeper waters, these large algae are replaced by smaller turfing red algae and sessile invertebrates (particularly sponges, bryozoans, ascidians, hydroids, and soft corals).

Urchins are important algal grazers on temperate reefs (Keough and Butler 1995). Dominant genera on open coasts range from *Centrostephanus* and *Heliocidaris* in New South Wales, to *Heliocidaris* alone in South Australia, to a mixture

of *Heliocidaris*, *Tripneustes*, and *Echinometra* in southwestern Western Australia. In the warm temperate waters of eastern Australia, heavy grazing by the common eastern Australian urchin (*Centrostephanus rodgersii*), typically results in distinct "urchin barrens" (Jones and Andrew 1990). While this habitat (and urchin) has, until recently, been rare in southern Australia, it has become more common with the introduction (and increasing dominance) of *Centrostephanus* in southern waters over the past several decades. This is probably due to the increased southern penetration of the warm water, Eastern Australian Current along the eastern seaboard (due possibly to climate change), which has resulted in a significant increase in warm temperate components of the marine biota in southern Australia (Edgar 1999).

Rocky reef provides refuge and feeding opportunities for a wide variety of fish and mobile invertebrates (Jones and Andrew 1990; Lincoln Smith and Jones 1995). Small fish and invertebrates can escape predators by hiding in cracks and crevices. Larger fish, along with octopus and cuttlefish appear to use rocky reef as cover from which they can ambush passing prey. Pelagic fish are also attracted to rocky reef areas by aggregations of small baitfish. Many species of invertebrates and algae live on or within rocky habitat and provide a diverse range of foods. Some fish along with abalone and sea urchins eat drift and/or attached algae associated with rocky reefs (Jones and Andrew 1990). Many commercially and/or recreationally important fish and invertebrates depend on rocky reef habitat for some or all of their life (i.e., rock lobster, abalone). However, very little is known of the habitat linkages of these species, and their trophic interactions with other reefal organisms.

The species and size composition of fish communities using rocky reefs is likely to vary between (superficially) similar reefs, even within a relatively small area (e.g., a particular embayment, estuary or length of coastline), because of factors such as water quality, larval supply and proximity to other habitats such as sea grasses and mangroves (Bell et al. 1988; Hannan and Williams 1998). Thus, many reefs are likely to have a degree of uniqueness in terms of their ecological function. Estuarine reefs, though relatively species-poor, are likely to represent an important "stepping stone" for larger juvenile fishes moving between nursery habitats (such as sea grass and mangroves) and habitats used by adults.

In areas of high wave exposure and localized, cold-water oceanic upwellings, very high levels of invertebrate and macroalgal diversity (i.e., 95–125 dominant species) occur on nearshore reefs, particularly for species of Rhodophyta (i.e., red algae) (Shepherd 1981; Edyvane and Baker 1996). Species of articulate and crustose coralline red algae are particularly prevalent in these mixed-red algal communities, and are known to form key microhabitats for the larvae of economically important invertebrates (such as the abalone, *Haliotis rubra*) (Shepherd and Turner 1985). In areas of very

high current or tidal movement, rocky reefs are dominated by suspension feeders (i.e., sponges, echinoderms, bryozoans), which form diverse sponge gardens, even in relatively shallow waters (<15 m).

Some macroalgal-dominated rocky reefs, by virtue of their latitude and/or offshore location, support distinct communities of tropical species. Diverse assemblages of reef-building corals along with associated tropical fish, algae, and invertebrates can be found as far south as the Solitary Islands (New South Wales) and Rottnest Island (Western Australia). Offshore islands are often host to tropical species well beyond their normal (inshore) ranges as they are often subject to warmer water temperatures than those experienced inshore (Kuitert 1993), due to greater influence by the warm East Australian Current (Kailola et al. 1993).

Sandy shores. Sandy and soft-sediment shores and their biotic assemblages are unlike any other marine benthic habitat. In contrast to rocky shores, sandy and soft-sediment habitats are characterized by their three-dimensional nature (i.e., depth is an important variable); range of grain size, depth and chemistry of the sediments (which exerts a profound influence on the types of organisms living within it); large size range of organisms which include some that ingest the sediment matrix, as well as living on or within it; and the lack of large attached plants and dominance of microscopic primary producers (see review by Fairweather and Quinn 1995). Importantly, soft-sediment habitats provide the contiguity of habitat with other types, such as sea grasses, mangroves, salt marsh and open ocean, which facilitates the movement of organisms among them – both between and within different stages of their life cycles.

As elsewhere in the world, unvegetated, soft-sediment shores remain one of the most under-researched marine benthic habitats in Australia (Fairweather 1990; Fairweather and Quinn 1995). In Australia soft-sediment research has been neglected in favor of studies on coral reefs, mangroves, rocky shores, and vegetated habitats within estuaries (Fairweather and Quinn 1995). In a review of research papers published between 1980 and 1987, Fairweather (1990) reported that only 10% of the 729 papers examined dealt with sandy bottoms, and only 6% with muddy bottoms. This is despite the fact that sandy shores represent on average, approximately 47% (or 36–68%) of Australia's coastline, while muddy shores (particularly adjacent to mangrove forests) comprise approximately 21% (or 1–48%) of the coastline (Fairweather and Quinn 1995).

Soft-sediment research studies in Australia are very limited and have largely been descriptive (i.e., species lists and zonation schemes) with a few quantitative or semi-quantitative studies (i.e., community descriptions linked to depth zonation), and include tidal flat studies (Rainer 1981), subtidal soft-sediment studies on detailed examinations of spatial and temporal variation (Morrissey et al. 1992a, b) and quantitative

surveys of benthos in Port Phillip Bay (Poore and Rainer 1979). Significantly, soft-sediment research in Australia has benefited considerably from foreign scientists who have established research programs in Australia, particularly on tidal flats in Western Australia (i.e., Black and Peterson 1987; Peterson and Black 1987) and sandy beaches in NSW and Western Australia (i.e., Dexter 1983a, b, 1984; MacLachlan and Hesp 1984).

Due to the paucity of experimental studies, there are presently no cohesive theories about any aspect of the ecology of soft-sediment intertidal assemblages in Australia (Fairweather and Quinn 1995). Similarly, understanding of trophic interactions on sandy beaches and hence the flow of energy in soft-sediment ecosystems, is limited.

Coral reefs. The central Indo-Pacific is the world's center of coral reef diversity, with at least 50% of the world's coral reefs, covering an area greater than 300,000 km². Australia has the largest area of coral reefs of any nation and the largest coral reef complex, the Great Barrier Reef (GBR). The GBR is the largest single coral reef in the world. There are also large areas of coral reefs in Torres Strait, the Coral Sea Territories and central and northern Western Australia. The Tasman Sea reefs (Elizabeth and Middleton Reefs and Lord Howe Island fringing reef) are the highest-latitude coral reefs in the world, thriving in conditions that are marginal for coral growth elsewhere. Australia's coral reefs form seven distinctive groups: high-latitude reefs of eastern Australia (Solitary Islands, Lord Howe Island, Elizabeth and Middleton Reefs), the GBR, reefs of the Coral Sea (e.g., Ashmore Reef, south to Cato and Wreck reefs), reefs of northern Australia and shallow, turbid waters of the eastern Arafura Sea, Cocos (Keeling) Atoll and Christmas Island, reefs of the Northwest Shelf, reefs of coastal Western Australia (e.g., Ningaloo and Houtman Abrolhos).

The central Indo-Pacific region of northern Australia and Southeast Asia supports by far the richest invertebrate fauna associated with coral reefs in the world (i.e., more than 100 coral genera and some 500 coral species, compared with only 25 genera in the Western Atlantic, and about 60 species from the Caribbean). Australia, because of its geographic position within the world's center of marine biodiversity, is critical to the conservation of corals because of the level of degradation in the region's equatorial reefs. Approximately 70% of all central Indo-Pacific coral reefs have been significantly degraded (Vernon 1995). This degradation has occurred primarily through over-fishing (which has effectively removed the top of the food pyramid in most Southeast Asian and Japanese areas of reef), eutrophication and increased sedimentation (from urban outfall, deforestation, agricultural runoff and coastal development), and direct intrusive activities (principally through subsistence food gathering – particularly mining practices, shell collecting, and unregulated tourism).

The GBR is not the most diverse in terms of species (Indonesian and Philippine reefs have greater numbers of coral species) but it is extremely diverse in terms of reef types, habitats, and environmental regimes. This is because the GBR is large enough to extend from the low latitude tropics to subtropical zones, to have regions with very different climates (wind patterns and rainfall), tidal regimes, water qualities, bathymetry, island types, substrata, and even geological histories.

Specific threats to Australia's coral reefs include, elevated nutrients in the inner GBR, coral bleaching events, outbreaks of crown-of-thorns (*Acanthaster*) starfish in the outer central and northern GBR and Tasman reefs, damage from the passage of tropical cyclones and outbreaks of coral-eating *Drupella* snails in Ningaloo Reef, Western Australia. Coral reefs are well represented in marine protected areas in Australia. Australia's reefs are less affected by human activities than those in other countries, mainly due to low to moderate levels of use and their remoteness, but elevated nutrients and sediments resulting from inland soil erosion are a threat in non-arid regions.

Research on Australia's coral reefs has largely excluded inter-reefal regions (these comprise more than 90% of the area of the GBR and are heavily trawled for fish and prawns); has inadequately considered biology and processes at a total-systems level; has largely excluded the far northern GBR, the reefs of the Northern Territory and, most importantly, the reefs of Western Australia; and has failed to consider microbial processes (Vernon 1995). Research has also failed to adequately address the needs of long-term conservation, in particular, the effects of sediments and nutrients on reefs, the root cause of *Acanthaster* outbreaks, and the long-term effects of commercial and recreational fishing.

Evolution of Marine Biodiversity

Australia's coastal environments range from tropical to temperate latitudes and consequently, encompass a wide range of coastal habitats, such as coral reefs, algal reefs, seagrass beds, mangroves, saltmarshes and also, non-vegetated habitats (sandy beaches, mudflats). All major groups of marine organisms are represented, and many of the highly diverse groups and species are endemic – particularly in the southern temperate Australian waters. Australia has the world's largest areas and highest species diversity of tropical and temperate seagrasses, largest area of coral reefs, highest mangrove species diversity and third-largest area of mangroves (Zann 1995).

The distribution patterns today are the result of contributions from two different early Tertiary biotas:

1. The pan-Pacific Tethyan biota and its successor, the Indo-West Pacific biota, have dominated the northern coasts of Australia since the beginning of the Tertiary and also

contribute to temperate biotas, especially in the southwest. To the north, barriers to interchange of shelf and coastal biotas with Southeast Asia are only slight. There are, therefore, many widespread tropical elements in the northern Australian biota and a low percentage of endemism. At its southern limit the Tethyan element is limited by the latitudinal temperature gradient.

2. The temperate Paleoaustrian fauna has dominated southeastern Australian coasts also from the early Tertiary and is now the major element of the biota of the entire southern coast.

Australia's marine biological diversity, like that of the land, is notable for its high proportion of endemic species, particularly in southern Australia. While the marine flora and fauna of tropical Australia and the Indo-Pacific mixed some 20 million years ago (when the continental plates of Australia and South East Asia collided), the marine biota of southern temperate Australia has remained geologically and climatically isolated for over 65 million years – resulting in some of the highest levels of endemism in the world (Poore 1995). In southern Australia about 80–90% of the species in most marine groups (e.g., fish, echinoderms, molluscs) are considered to be endemic (Poore 1995) (see Table 3). In northern Australia, only about 10% of the species in most groups are endemic. Approximately 85% of fish species, 95% of molluscs, and 90% of echinoderms are endemic. In contrast, approximately 13%, 10%, and 13% of fish, mollusc and echinoderms, respectively, are endemic in the tropical regions of Australia (Wilson and Allen 1987). The species in northern waters are mostly shared with the Indonesian archipelago, the epicenter of global marine biological diversity, but a region where many marine ecosystems are under threat.

Australia's southern marine platform, one of the largest in the world, is of considerable evolutionary significance. Due to its tectonic stability since the late Eocene (approximately 40 million years ago), sequences deposited have been little disturbed compared with many other parts of the world. As such, it provides a unique glimpse of the direct ancestral lineages for many invertebrate species found there today. These include the bivalve genus *Bassina* (family Veneridae), which has a continual fossil record from the Oligocene, and *Neotrigonia*, the only extant member of the family Trigoniidae, which was widespread and abundant in the Mesozoic (Durham 1979; Fleming 1979). Many of the endemic genera, such as the cowries *Notocypraea* and the volutes *Ternivoluta*, have fossil lineages dating back to the early Tertiary and may be survivors from an ancient paleoaustrian fauna. Other endemic genera, such as the cowries *Zoila*, the bivalves *Miltha* (family Lucinidae) and gastropods of the genus *Diastoma* (family Diastomatidae) are relicts of once widespread Tethyan groups that have become extinct elsewhere (Wilson and Allen 1987).

Australia, Coastal Ecology, Table 3 Marine biodiversity in Australia. (From Edyvane 1999; courtesy of SARDI)

Taxon/group	Temperate		Tropical	
	Number of species	Degree of endemism	Number of species	Degree of endemism
Macroalgae	1155	75% (red algae)	400	Low
Sea grass	22	95%?	15	Low
Fish	600	85%	1900	13%
Echinoderms	220	~90%	High	13%
Asteroids	50	High	?	—
Crinoids	7	High	High	—
Ophiuroids	74	High	?	—
Echinoids	49	High	?	—
Holothurians	40	80%	?	—
Ascidians	210	73%	?	?
Mollusks	?	95%	?	10%
Cnidaria	—	—	—	—
Anthozoa	200	?	High	?
Hydrozoa	200	High	Low	?
Scyphozoa	10	?	High	?
Sponges	~1000	High	?	?
Bryozoa	~500	High	Low	?
Pycnogonids	50	73%	Low	?

Australia has a very long coastline (both latitudinally and longitudinally) but trends in species diversity are poorly documented. While salt-marshes, mangroves, fishes, corals, molluscs, echinoderms, and decapod crustaceans (crabs, hermit crabs, lobsters, shrimps, prawns) decrease in diversity from north to south, the same trends are not apparent in other marine taxa, that is macroalgae, seagrasses, sponges, bryozoans, nudi-branches, ascidians, and peracarid crustaceans (amphipods, isopods, and others) (Table 3). In marine amphipods (Barnard 1991), the reverse is true. In isopod crustaceans a diversity gradient overall is not yet apparent but relative dominance of families certainly changes with latitude as it does in many other taxa. These biogeographic distributions have repercussions in functional ecology – for example, the taxonomic composition of crustacean scavenger guilds. Cypridinid ostracodes, cirolanid isopods, and lysianassoid amphipods play this role but last of these become relatively more important as one moves from tropical to temperate waters.

A noteworthy aspect of the temperate fauna is the occurrence in many genera of eastern and western species pairs that do not interbreed because of geographic isolation (allopatric species). Among the fish fauna there are some 18 of these pairs of sister species, including the wrasse, *Achoerodus viridis* (Eastern Blue Groper) and *Achoerodus gouldii* (Western Blue Groper), *Pelates octolineatus* (western species) and *Pelates sexlineatus* (eastern species) of the grunter family (Terapontidae), and *Vicentia* species (western species) and *Vicentia con-sparsa* (eastern species) of the cardinal fish family (Apogonidae) (Wilson and Allen 1987). As a result, warm temperate fish faunas of the Pacific and Indian Oceans

are distinct from each other. East–west species pairs have also been recorded in brachyuran crabs, mollusks, and asteroids in Bass Strait (Dartnall 1974).

A number of Australia's marine species, although found elsewhere, find the greatest security in Australian waters. These include the green turtle (*C. mydas*), hawksbill turtle (*Eretmochelys imbricata*), and logger-head turtle (*Caretta caretta*), now rare except in Australian waters (Marsh et al. 1995). The dugong (*D. dugon*), which is suffering from population decline in many parts of its range, is found in greater numbers in Australian waters than anywhere else in the world.

Trends in Marine Biota

The seas around Australia are globally significant in terms of many taxonomic groups. However, little is known of the status of most of the marine species. Scientific interest to date has largely centered on the higher vertebrates such as turtles, seabirds, seals, dugongs, and whales. Generally, microscopic organisms (algae, fungi, and bacteria), invertebrates and fish have been neglected. Many marine species remain undescribed and relatively little is known about most of the described species. The conservation status of very few invertebrate species is known. A number of species appear vulnerable because they are rare and have quite restricted habitats. An enormous taxonomic and monitoring effort would be required in Australia to describe all species and to determine their status. While regulations governing many of the fished species have long existed in Australia, marine fish

conservation is a relatively new field and the conservation status of most species is poorly known.

Australia's marine phytoplankton includes representatives of all 13 algal classes, including diatoms (5000 species) and dinoflagellates (2000 species). The marine plant diversity of Australia is of global significance, particularly among the macroalgae of southern Australia, which represents one of the richest and most unique marine flora's in the world (Womersley 1990; Huisman et al. 1998; Phillips 2001). The richness of the temperate macroalgal flora (i.e., 1155 species) is approximately 50–80% greater than for other comparable regions around the world, with an estimated 800 species and over 75% endemism recorded in the Rhodophyta (red algae) alone (Womersley 1990). This is largely due to the length of the southerly facing rocky coastline (i.e., the longest, ice-free, temperate coastline in the world) and the long period of geological isolation (Edyvane 1996). While other regions around the world, such as Japan and the Pacific North America, have recorded a higher number of macroalgal species (i.e., 1452 and 1254 species, respectively), the coastal waters of these regions encompass a wide range of climatic conditions, from arctic to tropical. Similarly, the level of temperate species biodiversity in macroalgae is approximately three times the level recorded in the tropical regions of Australia, where approximately 200–400 species of macroalgae have been described (Womersley 1990).

The angiosperms (flowering plants) are also very well represented. Australia has 11 of the world's 12 genera of seagrasses and over half the total number of species (Larkum and den Hartog 1989). Australia's waters also contain the highest level of species diversity and endemism for seagrasses in the world, with the greatest levels of speciation and endemism in temperate waters, where 22 species have been recorded (cf. 15 species in tropical waters) (Poiner and Peterken 1995). Australia has 16 of the world's 20 families of mangroves and 40 of the world's 55 species of mangroves.

Among the marine fauna, Australia has one of the largest fish faunas in the world, due to the wide latitudinal range of Australian marine habitats, the presence of the Great Barrier Reef (the world's largest coral reef) and the radiation of a number of temperate fish families (Paxton et al. 1989). The warm East Australian Current and Leeuwin Current, flowing down the east and west coasts, respectively, also seasonally bring tropical species into more southerly latitudes. Australia has an estimated 4000–4500 species of fish, of which around 3600 have been described. About a quarter of the species are endemic; most of these are found in southern Australia.

Diversity of several major marine taxa is highest in the tropical regions of Australia. This includes molluscs, echinoderms, and fish. Among the molluscs, most families are represented in the Australian fauna and have more species in the north of the country than in the south. The level of endemism in the north is low (about 10%) with most species

distributed widely in the Indo-West Pacific region. In contrast, endemism of the southern Australian fauna is about 95% and several endemic genera occur. Some endemic genera are relicts of the once widespread Tethyan fauna (as are the endemic families, Trigoniidae, Campanilidae, and Diastomatidae, each with a sole living representative). Other endemic genera are relicts of the ancient Paleoaustrian fauna.

Echinoderms, with only a few hundred Australian species, repeat the biogeographic patterns of the fishes and molluscs with only 13% of species endemic in the tropical region and 90% in the temperate region. Many echinoderms have a long larval life which may explain why 22% of Tasmanian species occur also in New Zealand where they are transported by the West Wind Drift.

Among the fish fauna, approximately 3400 species of marine fishes are pelagic or oceanic and wide-ranging in tropical and temperate seas (Paxton et al. 1989). About three-quarters occur on the shelf and nearshore. The greatest number of species are in the tropics where approximately 120 families, 600 genera and more than half of the species (1900) are found; and most of these are common to the Indo-West Pacific region. Although most species have pelagic eggs and larvae a moderate level of endemism (13%) is maintained with the help of southerly flowing currents on both the east and west Australian coasts.

The fish fauna of temperate Australia comprises about 600 species of which 85% is endemic and 11% is shared with New Zealand (Wilson and Allen 1987; Kuiter 1993). Endemism at the genus level is about 38% (of an estimated 290 genera). One contribution to this high level of endemism is radiation in a few families with low dispersability; viviparous clinids, brooding syngnathids, and nesting gobioides and gobiids. In some 8 families of fish that have more than 10 species, more than half of the total world species occur in Australian waters (e.g., Cheilodactylidae). The five marine fish families Brachionichthyidae, Pataecidae, Enoplosidae, Gnathacanthidae and Dinolestidae are endemic to Australia and a number of these families are monotypic. Several families of fish are noted for their explosive radiations in southern Australia. They include pipefishes and sea horses (Syngnathidae), leatherjackets (Monacanthidae), and anglerfishes (Antennariidae). The Australian pipefish fauna is regarded as uniquely diverse. Nearly 80% of the 47 genera in the Indo-Pacific region are found in Australia and 14 of these genera (38%) are endemic. Ten of these genera are monotypic (Wilson and Allen 1987).

Australia also has an extremely rich chondrichthyan (shark, chimeras or ghost sharks, and rays) fish fauna. It is estimated that at least 296 species, comprising 166 sharks, 117 rays, and 13 chimeras, live in Australian waters. The richness of this fauna is apparent through comparisons with other areas: 182 species have been recorded from southern Africa; 174 from the Japanese Archipelago; and 130 from the eastern North Atlantic and Mediterranean Seas (Last and

Stevens 1994). More than half of the entire Chondrichthyan fauna, some 54%, is endemic to Australia. Of the shark fauna, 48% of the species are endemic, with most of the endemic species being found in tropical or warm temperate areas. More than half of the Australian chimeras appear to be regional endemics, and some 73% of Australian rays are endemic.

Among the reptiles, Australia has about 30 of the world's 50 species of sea snakes, around half of which are endemic (Marsh et al. 1995). The family of aipysurids live in coral reef waters and the family of hydrophiids live in inter-reefal waters of Australia's tropics. Six of the seven world's turtle species are found in Australian waters. One, the flatback turtle, is endemic. Breeding migrations may cover hundreds to thousands of kilometers and many turtles breeding in Australia may live around the island of Papua New Guinea, the southwestern Pacific Islands and Indonesia, making species management difficult. Turtles are slow to reach maturity and breed perhaps only five times in their lives, making them extremely vulnerable to overexploitation.

About 110 species of seabirds belonging to 12 families are found in Australia and its external territories. Of these, 76 species breed and spend their entire lives in the region. The remaining 34 species are regular or occasional visitors. Some 14 species or subspecies of Australia's seabirds (13% of the total) are considered to be threatened, largely because their colonies on oceanic islands are few in number and are vulnerable to harvesting and natural disasters. The wandering albatross on Macquarie Island, Abbot's booby on Christmas Island and the Australian subspecies of the little tern are classified as "Endangered" under IUCN criteria. Lord Howe's Kermadec petrel and white-bellied storm-petrel and Christmas Island's Christmas frigatebird are considered "Vulnerable."

In contrast to temperate biota, reef-building scleractinian corals are essentially tropical and their distribution is substantially different from that of echinoderms, molluscs, and fishes. The Australian fauna is a subset of Indo-West Pacific corals and is largely confined to the Great Barrier Reef and smaller reefs on the west coast (Vernon 1995). There is not a gradual reduction in the number of species with increasing latitude but rather an abrupt depletion of species at the termination of the Great Barrier Reef and at the Houtman Abrolhos. Coral reefs do occur further south – for example, at Lord Howe Island and Solitary Islands – but the number of species is few. Few species of coral occur on the southern coast of Australia and reefs are not formed there.

The tropical dugong is the only fully herbivorous marine mammal and the only Sirenian (sea cow) to occur in Australia (Marsh et al. 1995). It is extinct or near extinct in most of its former range which extended from East Africa to South East Asia and the Western Pacific. Northern Australia has the last

significant populations (estimated to be over 80,000) in the world. Large, long-lived mammals, dugongs become sexually mature at around 10 years and calve every 3–5 years, making them vulnerable to over-hunting. Major concerns are possible overhunting of Torres Strait populations, mortalities in fish gillnets and shark nets and loss of sea grass habitat.

Three species of eared seals or pinnipeds breed in mainland Australian waters: the rare and endemic Australian sea lion (*Neophoca cinerea*), the Australian fur seal (*Arctocephalus pusillus doriferus*) and New Zealand fur seal (*Arctocephalus forsteri*) (Shaughnessy 1999). The Australian populations of the New Zealand Fur Seal are limited in their distribution to southern Tasmania and the Great Australian Bight, and are found on the islands of Recherche Archipelago (WA), eastwards to Kangaroo Island (SA). The Australian Sea Lion is one Australia's most endangered marine mammals and one of the rarest and most endangered pinnipeds in the world and is endemic to Australia (Gales et al. 1994). Prior to seal-hunting, this species occurred along the whole of the southern coastline, but is now limited to the offshore islands of Western and South Australia, from the Houtman Abrolhos to the islands of Recherche Archipelago (WA), and from Nuyts Archipelago to Kangaroo Island (SA). Breeding populations of the Australian Fur Seal, which also breeds in South Australia, are confined to southeastern Australia, including Tasmania (Shaughnessy 1999). Major human threats include entanglement in fishing nets and ocean litter, incidental killing, oil pollution, and disturbances by visitors.

Eight species of baleen whales and 35 species of toothed whales, porpoises, and dolphins are found in Australian waters, although none is endemic (Bannister et al. 1996). This is almost 60% of the world's total cetacean species. Australian breeding populations of southern right whales were seriously depleted by hunting as early as 1845. Their population has slowly increased from small remnants totaling a hundred or so, to between 500 and 800. Australian breeding populations of humpback whales were depleted by 1963. Numbers are recovering and there are now thought to be up to 4000 breeding in Australian waters. Gillnets, shark nets set off bathing beaches, discarded fishing nets and ingestion of plastic litter are considered threats to cetaceans in Australian waters.

Marine and Coastal Research

While marine and coastal research in Australia is relatively young, Australia has gained preeminence in a number of research areas, including coral reef/tropical marine studies, marine geosciences, physical oceanography, fisheries research, coastal hydrodynamics and numerical modeling, and remote sensing techniques (DITAC 1989; Zann 1996).

While some subjects and geographic areas are well known, there are serious gaps in knowledge and understanding of Australia's marine environment, particularly in relation to coastal and reef ecosystems, ocean/atmosphere processes, marine chemistry, chemical oceanography and importantly, baseline information on coastal and marine ecosystems, habitats and processes (to facilitate sustainable use).

Research on Australia's tropical marine environments and habitats (such as mangroves and coral reefs), particularly in the Great Barrier Reef region, is relatively advanced. This is largely due to the location and research missions of national marine research and management agencies (Fairweather and Quinn 1995; Edyvane 1996). In 1991, around 40% of marine research projects in Australia were undertaken in the tropics. Most research was undertaken in the Great Barrier Reef region (31% of total), followed by the New South Wales coast (16%), Bass Strait (12%), the southwest coast (5%) and the northwest (3%) (Zann 1996).

Despite the high concentration of Australia's population (approximately 80% of the total population) in the coastal temperate areas of the continent, the habitats and biodiversity of temperate marine ecosystems (particularly subtidal habitats), have largely been under-researched and hence, are poorly documented and understood (Fairweather and Quinn 1995). This has resulted in the lack of habitat conservation (and integrated management) in many areas of southern Australia, despite the high level of threats and the high level of biodiversity and endemism among a range of marine faunal and floral groups.

Temperate habitat studies in Australia have largely focused on isolated studies of intertidal rocky shores and estuaries and sea grass areas – habitats recognized as important in the management of nearshore fisheries (Fairweather and Quinn 1995). Research on temperate rocky shores has principally focused on intertidal habitats or the biology of individual reef species or assemblages and particularly, population studies of commercially important species (Fairweather and Quinn 1995). Regional habitat studies are generally rare and descriptions of macrofloral (or faunal) assemblages of subtidal habitats are geographically very patchy. For some regions (e.g., South Australia, Tasmania) detailed information is known based on isolated subtidal field surveys conducted in the 1970s and 1980s (i.e., Shepherd and Womersley 1970, 1971, 1976, 1981; Shepherd and Sprigg 1976; Shepherd 1983; Edgar 1984). It is only recently that studies involving broadscale coastal mapping and documentation of temperate subtidal communities, including, reef, sea grass, and soft-bottom communities, have commenced (Edyvane 1999; Edgar et al. 1997). Elsewhere (such as in the Great Australian Bight and offshore waters), there are almost no published inventories or maps on subtidal benthic habitats (Edyvane 1996, 2000). This reductionist approach to the study of

coastal nearshore ecosystems has principally hindered understanding of these ecosystems and the ecological linkages between reefal and non-reefal habitats (such as sea grass, mud, sandy coasts).

Habitat inventories are particularly required for rocky coasts and reefs and the continental shelf and large gulfs – the two habitats supporting the most valuable commercial fisheries (Cappo et al. 1998). The need for a better understanding of habitat processes and dynamics was highlighted for all major habitat types – as testable paradigms about fisheries production exist only for temperate seagrass. Relative to their fisheries value, overall gaps in knowledge are particularly outstanding for the continental shelf (and slope), followed by rocky coasts, and reefs. Perhaps least studied in Australia are the nursery role and biological processes of sandy shores.

Much of the marine ecological research conducted in temperate Australia has focused on the study of ecological processes, particularly on rocky intertidal shores, rather than a description of the patterns of biodiversity and the distribution and abundance of marine assemblages. These studies have particularly focused on the role of intra- and inter-specific competition and predation and their influence on the structure of rocky intertidal assemblages (Underwood and Kennelly 1990), and also, patterns and variability in recruitment of marine invertebrates, both on small and large spatial scales (see review by Fairweather and Quinn 1995; Keough and Butler 1995).

Marine ecological research in Australia requires more effort into baseline mapping of marine biodiversity and the impacts of human use (at a range of temporal and spatial scales), long time-series datasets for trend analysis, and the modeling of coastal processes, and also a greater focus on the linkages between: catchment practices and effects on nearshore ecosystems; reefal habitats and non-reefal habitats; offshore and nearshore ecosystems; and the relationship between coastal physiography, physical processes and scaled patterns of biodiversity, that is ecosystem geography (Edyvane 1996).

In recent years, a national audit of catchments, rivers and estuaries has been undertaken (Commonwealth of Australia 2002). Several large, ecosystem-scale modeling studies of Australia's marine environments have also been undertaken, primarily for pollution management. These include Port Phillip Bay in Victoria (CSIRO 1996b), and Cockburn Sound and Albany Harbor (Simpson and Masini 1990) in Western Australia, and the Huon River estuary in Tasmania, (CSIRO 2000) and Hervey Bay in Queensland (Tibbetts et al. 1998, Dennison and Abal 1999). A major ecosystem study of the North West Shelf (in Western Australia) has been undertaken for tropical fisheries and multiple-use, ecosystem based management (NWSJES 2002).

Summary

Australia's marine environment extends from the coast to the boundary of its 200 nautical mile exclusive economic zone and covers million square kilometers of seas (an area 16% larger than the land). As the world's largest island, Australia has wide range of coastal and marine environments, which stretch approximately 61,700 km, from the tropical northern regions to temperate southern latitudes. Along this extensive coastline occur a wide range of habitats and biological communities including rocky shores, sandy beaches, algal reefs, kelp forests, which dominate the temperate south, and coral reefs, estuaries, bays, seagrasses beds, mangrove forests and coastal salt marshes, which dominate the tropical north and also, the less understood mid-water, outer-shelf, and deepwater habitats. Australia's marine environments also include external territories in the Indian Ocean, South Pacific, Southern Ocean, and Antarctica.

The extent and diversity of Australia's marine and coastal environments has resulted in some of most diverse, unique and spectacular marine life in the world – supporting some of the greatest number of marine species in the world. For instance, Australia has the world's largest areas and highest species diversity of tropical and temperate sea-grasses, the highest diversity of marine macroalgae, the largest area of coral reefs, the highest mangrove species diversity and global levels of biodiversity for a range of marine invertebrates (e.g., bryozoans, ascidians, nudibranchs) (Zann 1995). Approximately 4000 species of fish, 43 species of whales and dolphins, and 6 of the 7 world species of marine turtles are recorded from Australia.

All major groups of marine organisms are represented in Australian waters, and many of the highly diverse species are endemic or unique to Australia's waters. The nature of the temperate Australian marine biota, particularly the high level of endemism, is principally a result of the unique geological and tectonic history of Australia, particularly the long period of geological and climatic isolation. In temperate southern Australian waters, which have been geographically and climatically isolated for around 65 million years, most known species (i.e., 90–95%) are endemic or restricted to the area (Poore 1995). In contrast, in the waters of tropical northern Australia, which are connected by currents to the Indian and Pacific Ocean tropics, most species (i.e., 85–90%), are shared with the Asia-Pacific region.

Cross-References

- [Australia, Coastal Geomorphology](#)
- [Coral Reefs](#)
- [History, Coastal Ecology](#)

- [Indian Ocean Coasts, Coastal Ecology](#)
- [Mangroves, Ecology](#)
- [New Zealand, Coastal Ecology](#)
- [Rock Coast Processes](#)
- [Salt Marsh](#)
- [Sandy Coasts](#)
- [Wetlands](#)

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Australia, Coastal Geomorphology

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The Australian coastline is about 20,000 km long (Fig. 1) and has many long gently curving sandy beaches interspersed with sectors of cliffs and steep coast. The beaches on the western and southern coasts, from Broome in the northwest to Wilsons Promontory in Victoria and the western coasts of the Bass Strait islands and Tasmania, are predominantly calcareous, whereas those of the eastern and northern coasts are generally quartzose, except in the vicinity of fringing coral reefs. Many of the beaches form the seaward margins of depositional sand barriers bearing multiple beach ridge or dune topography (Bird 1973; Thom 1984), and on the western and southern coasts, the calcareous sands of the older (Pleistocene) dunes have been partially lithified by secondary carbonate precipitation to form dune calcarenites (calcareous aeolianites). These are extensive between Broome and Cape Leeuwin on the Western Australian coast, between Streaky Bay in South Australia and Cape Otway in Victoria, and locally around the shores of Bass Strait, including the west coasts of King Island and Flinders Island and the northwest coast of Tasmania. In places these Pleistocene dune calcarenites have been exposed to marine erosion and cut back to form cliffs and shore platforms (Fig. 2).

Cliffs and rocky shores have also developed where older geological formations reach the coast: in Tertiary formations near Cape Cuvier, in granites and older shield formations between Cape Leeuwin and Albany, and in Tertiary formations at the head of the Great Australian Bight. Precambrian rocks reach the coast on the Eyre and Yorke Peninsulas, on the Fleurieu Peninsula, and on Kangaroo Island in South

Australia. In Victoria there are cliffs cut in various formations, including Cenozoic basalts, Tertiary and Mesozoic sediments, Paleozoic rocks, and intruded granites. Tasmania has extensive rocky and steep coasts developed on pre-Carboniferous rocks and intruded granites, with Mesozoic formations, including Jurassic dolerites, on its southeastern coast. In New South Wales, there are cliffed and rocky headlands on Mesozoic and older rock formations, and in Queensland steep coasts, with limited cliffing, occur on similar formations inshore from the Great Barrier Reef (Fig. 3). On the north coast of Australia, cliffed and rocky sectors are extensive in Arnhem Land and where the Kimberley Ranges reach the sea, both areas of Paleozoic and older rock formations.

In southwestern and southeastern Australia, most of the rivers flow into estuarine lagoons with outlets constricted by sandy barrier formations. These lagoons show stages of infilling, largely by fluvial sediments, and some have been reclaimed by river deposition as broad swampy alluvial plains. In Northern Australia, river mouths are less encumbered by coastal sand deposition and form estuarine gulfs and deltaic regions. The shores of estuaries, lagoons, and sheltered embayments are often bordered by marshes and swamps, which have advanced on to tidal sandflats or mudflats. The latter becomes more extensive on sectors of the northern coast where tide range is large.

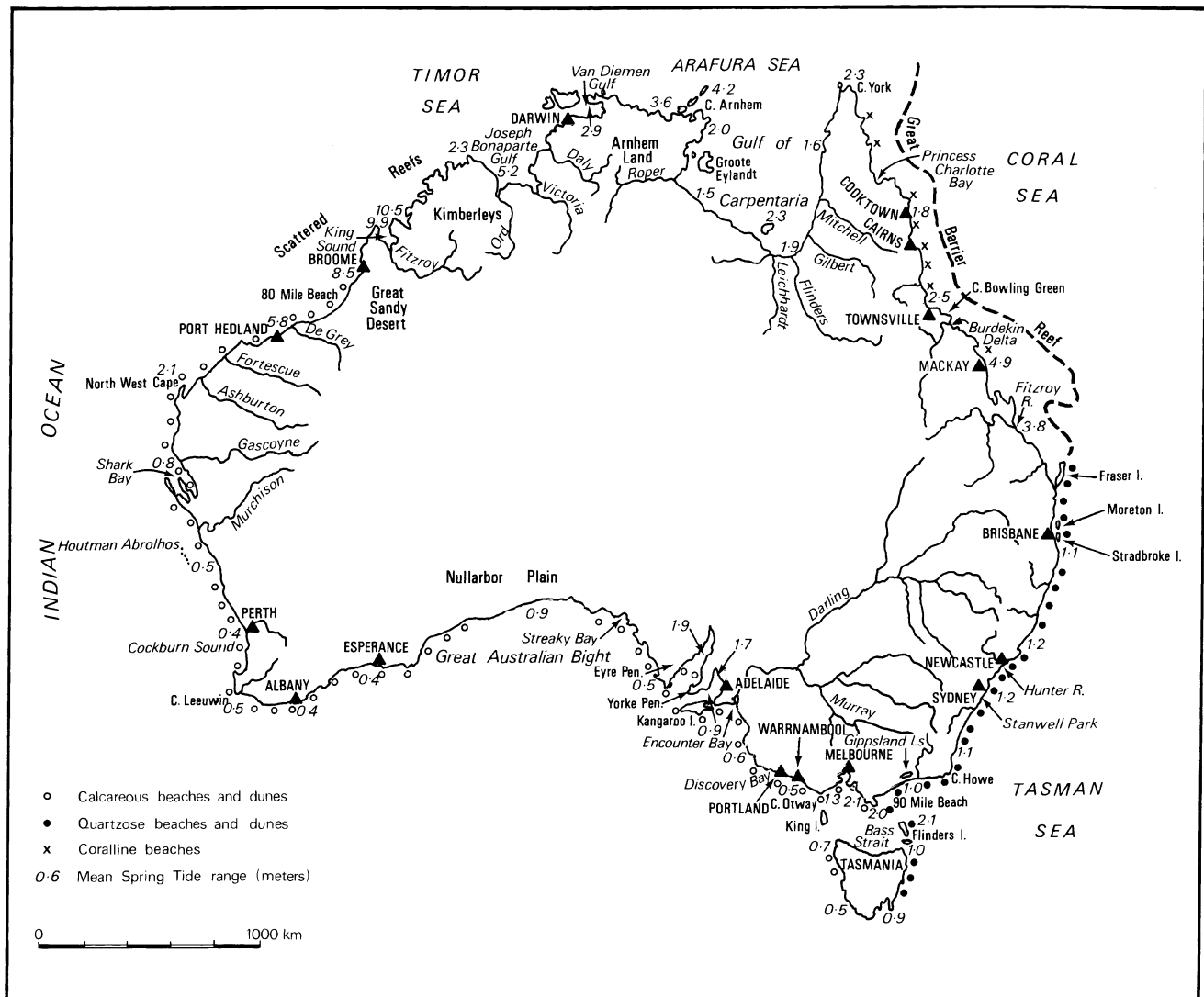
Coralline reef formations are extensive off the northern half of Australia, particularly in the Great Barrier Reef off Queensland. Numerous outlying patch reefs and coastal fringing reefs occur off the north and northwest coasts, around to Houtman Abrolhos off Western Australia (Fairbridge 1967).

Variations in morphology around the Australian coast are related to geological, climatic, oceanic, and biological factors that vary regionally around the margins of the continent (Davies 1980), taking account of changes in climate and sea level that have occurred on and around the Australian land mass (7.8 million km²) and its bordering continental shelves (2.5 million km²).

Geological Factors

The western half of the Australian continent is a shield region dominated by Paleozoic and older formations that produce the hard, rocky coasts of the Kimberley region and southwest Western Australia. However, on much of the western coast, these formations are obscured by Pleistocene dune calcarenites and associated Holocene dunes. On the Eyre Peninsula in South Australia, the dune calcarenite mantle has been cut back by marine erosion to expose underlying harder Precambrian rocks, which become promontories as bordering cliffs in the less resistant dune calcarenites recede.

The eastern margin of Australia consists of folded Paleozoic rock formations with structural trends roughly parallel with the



Australia, Coastal Geomorphology, Fig. 1 Location and feature map

eastern coastline. These formations have been uplifted and dissected to form the Eastern Highlands, with a watershed varying from 20 to 400 km inland and a succession of relatively small drainage basins feeding rivers that flow down to the Pacific coast. The structural pattern is complicated by the presence of sedimentary basins occupied by Mesozoic formations: the Sydney Basin, for example, contains the major sandstone formations prominent in cliffs along the New South Wales coast between Jervis Bay and Newcastle.

Between the Western Shield and the Eastern Highlands, the continent is dominated by Cenozoic rocks, which occupy a number of basins, several of which extend to and beyond the present coastline. In eastern Victoria the Latrobe Valley syncline passes beneath a depression occupied by the Gippsland Lakes region, where complex sand barriers of Pleistocene and Holocene age are fringed by the Ninety Mile Beach. In western Victoria Tertiary rocks dip gently off the uplifted

Mesozoic formations of the Otway Ranges and pass beneath an extensive volcanic province, locally active into the early Holocene: the sequence is clearly displayed in cliff sections westward from Cape Otway to Discovery Bay. Bass Strait is underlain by another sedimentary basin, and on a smaller scale, Port Phillip and Western Port Bays in Victoria occupy fault-bounded sunkenlands (Bird 1993). Western Tasmania is dominated by lower Paleozoic rock formations, but in the southeast, these are overlain by Mesozoic rocks in an area where Cenozoic faulting has influenced the evolution of ridges and valleys now partly drowned as rias.

In South Australia, a sedimentary basin underlies the Lower Murray valley, and the river flows across Tertiary formations to enter coastal lagoons (Lakes Albert and Alexandrina) behind the Pleistocene dune calcarenite barriers of Encounter Bay. At the head of the Great Australian Bight, the Eucla Basin contains Tertiary formations that outcrop on a



Australia, Coastal Geomorphology, Fig. 2 Cliffs and shore platforms cut in Pleistocene dune calcarenite on the Nepean Ocean coast near Melbourne, Victoria (Photo: E.C.F. Bird, Copyright-Geostudies)

Australia, Coastal Geomorphology, Fig. 3 Steep coast of Macalister Range, North of Cairns, Queensland, with an absence of cliffing on a moderate wave energy coast inshore from the Great Barrier Reef (Photo: E.C.F. Bird, Copyright-Geostudies)



long stretch of even-crested cliffs bordering the riverless Nullarbor Plain. Similar cliffed formations mark the Carnarvon Basin at Cape Cuvier, but the Tertiary rocks in the Canning Basin, in northwest Australia, vanish beneath the

Great Sandy Desert and the dune calcarenites of the Eighty Mile Beach, and the Joseph Bonaparte Gulf Basin is largely a shelf structure. The Carpentaria Basin runs northward under the Gulf of Carpentaria, a graben with Cenozoic formations

declining from the western side of Cape York Peninsula, and the much smaller Laura Basin occupies the hinterland of Princess Charlotte Bay on the eastern side.

Despite the existence of these several Cenozoic basins, Australia has been generally a rather stable continent, and there are extensive planation surfaces, Cenozoic and older, that show little tectonic deformation. In consequence the continent is notably flat and low-lying, with 80% of the land area below the 500 m contour. Many of the rivers (e.g., the Murray) have low channel gradients as they approach the sea and carry mainly fine-grained sediments (silt and clay) to their mouths. Steeper gradients exist in rivers draining the Pacific slopes of the Eastern Highlands, and some of these carry sand and gravel in the loads delivered to the sea. Fluvial sediment yields are also influenced by outcrop lithology and past as well as present weathering regimes. Most of the granite outcrops in the Eastern Highlands have been deeply weathered, and rivers draining them derive quartz-rich sands; weathered basalt supplies silt and clay, and river gravels are produced by the disintegration of Paleozoic metasediments. On the coast marine erosion has generally removed weathered mantles, so that the granites, for example, usually protrude as rocky headlands, as at Wilsons Promontory in Victoria.

Although the general picture is one of prolonged stability in Australia, marginal neotectonic activity has been recorded locally, especially in Victoria, and the possibility that slight coastal warping has continued into Holocene times cannot be ruled out.

Climatic Factors

The prevailing aridity of the interior and much of western and southern Australia has resulted in few rivers draining to the coast: those few flow intermittently, with occasional brief episodes of flooding. In the southwest of Western Australia and the southeast of South Australia, winter rainfall nourishes seasonal runoff, but Australia's largest river system, the Murray-Darling, draining a catchment of about a million square kilometers, has only a limited discharge at its mouth compared with those of similar scale in more humid regions. In consequence of hinterland aridity, terrigenous sediment supply to the coasts of much of western and southern Australia has been meager, and coastal deposition nourished primarily from seafloor sources is dominated by carbonate sands of biogenic origin. The presence of such material in shelf sediments is a further consequence of the lack of terrigenous supply, hinterland aridity having evidently prevailed also during Pleistocene phases of lowered sea level.

By contrast, the Eastern Highlands have a higher and more regular rainfall, and the steeper catchments draining to the Pacific coast nourish river systems that have delivered a more

substantial terrigenous sediment supply. Coastal sediments are dominated by quartzose sands, some of which are supplied directly by such rivers as the Shoalhaven, the Hunter, the Burdekin, and the Don (Fig. 4), particularly during episodes of flooding. Much of the sand deposited in beaches and barriers along the Victorian and New South Wales coasts has, however, been swept in from the seafloor during successive marine transgressions in Pleistocene and Holocene times by the reworking of shelf deposits of quartzose sand derived partly from fluvial supply during low sea-level phases and partly from weathered rock outcrops now submerged. The contrast between the carbonate sands and dune calcarenites of southern and western Australia and the quartzose sands of the southeastern and eastern coast is thus related to a contrast in shelf sedimentation, due in turn to a contrast in hinterland topography and climate through Quaternary times.

Along the Queensland coast, terrigenous sediment supplied by rivers includes quartzose sands and lithic gravels derived from relatively steep catchments as well as muddy sediment derived from tropical weathering mantles on catchment outcrops. In addition, coastal sedimentation is locally complicated by the presence of fringing and nearshore reefs built by corals and associated organisms in tropical waters. Reef clastics derived from these are added to adjacent beaches, but despite the proximity of a rich coralline province, the beaches of Queensland are generally poor in carbonates (Bird 1978).

In Northern Australia, summer rainfall (some from occasional cyclones) results in substantial runoff from river systems, but catchments are generally larger and less steep than those of the east coast, and fluvial loads are dominated by fine-grained sediment derived from prolonged weathering of surface outcrops and sandy yields being limited and localized. Muddy sedimentation is extensive on the shores of the Gulf of Carpentaria and in northern embayments and inlets, particularly around Van Diemen Gulf, Joseph Bonaparte Gulf, and King Sound, each of which receives major inputs of river water and sediment.

Other features related to climate include the nature and rate of rock weathering in the shore zone, most active under humid tropical conditions in Northern Australia, and the vigor of wind action, which is much stronger on temperate coasts than in the humid tropical north, a contrast that helps to explain the more extensive development and greater mobility of coastal dunes in western, southern, and southeastern Australia (Fig. 5) than in the north and northeast. Davies (1980) elucidated such contrasts as these within a climatically based sector classification, clockwise around the continent, from warm temperate humid (Fraser Island to Portland), warm temperate arid (Portland to North West Cape), tropical arid (North West Cape to Broome), and tropical humid (Broome to Fraser Island) coastal sectors.

Australia, Coastal Geomorphology, Fig. 4 The River Don, in north Queensland, carries sand and gravel down to the coast, where it is spread along the shore by wave action (Photo: E.C.F. Bird, Copyright-Geostudies)



Oceanic Factors

Swell generated by storms in the Southern Ocean moves in to the western, southern, and southeastern coasts of Australia, and its refracted patterns are effective in shaping the long, curved outlines typical of sandy beaches. In New South Wales, for example, refraction of the dominant southeasterly swell is responsible for the asymmetrically curved outlines of beaches between bedrock headlands (Chapman et al. 1982).

In the southern half of Australia, coasts are frequently exposed to storm wave action generated in coastal waters as cyclonic depressions move through from west to east. With relatively deep water close inshore, these are typically high

wave energy coasts with bold cliffs and headlands and smooth depositional outlines. Extensive sand deposition has constricted the valley mouth inlets formed by marine submergence during the Late Quaternary marine transgression and formed estuarine thresholds of inwashed sand at their entrances (Roy 1984).

On the east coast north of Fraser Island, ocean swell is interrupted by reef structures, and in the lee of the Great Barrier Reef, it is largely excluded from coastal waters (Hopley 1982). Within these waters the prevailing southeasterly winds generate only moderate wave energy, and consequently the coast has but limited sectors of cliffing, poor development of shore platforms, and irregular depositional



Australia, Coastal Geomorphology, Fig. 5 Parabolic dunes spilling inland from the sandy beach on the coast of Encounter Bay, South Australia (Photo: E.C.F. Bird, Copyright-Geostudies)

outlines, with many spits, cusate forelands, and some protruding deltas, notably the Burdekin; sectors of mudflat and mangrove swamp occur on shores in the lee of headlands and islands. Sandy barriers and dune systems have been deposited on a much smaller scale than on the oceanic coasts and take the form of beach-ridge plains, often built as cheniers on emergent mudflats. River mouths are less encumbered by sand deposition, and coastal lagoons do not occur north of Fraser Island. In the absence of ocean swell, shoreward drifting of sea floor sand has made only a limited contribution to coastal sedimentation on the Queensland coast.

On the north coast of Australia, swell from the Indian Ocean is attenuated by the exceptionally broad continental shelf, and its effects are further diminished by the large tide ranges. In the Timor and Arafura Seas and within the Gulf of Carpentaria, local wind action produces generally low to moderate wave energy on bordering coasts. Steep sectors show little cliffing and retain the intricate outlines of submerged landscapes of fluvial dissection, while depositional sectors are typically irregular and varied in detail. Storm surges accompanying tropical cyclones accomplish rapid short-term changes and may be responsible for the emplacement of sandy cheniers on coastal plains, for example, on the depositional lowland south of Van Diemen Gulf and on the low-lying coasts of the Gulf of Carpentaria, where surges of up to 7 m have been reported.

Although Pacific tsunamis have generated only minor surges on the Australian coast in the past two centuries, it is possible that they occurred at earlier stages on a larger scale and may have shaped some apparently emerged features (Bryant et al. 1996).

The broad patterns of longshore drifting on the Australian coast are northward and eastward from Cape Leeuwin and northward along the east coast, but where coastal outlines are adjusted to the dominant swell patterns, little net drifting occurs (Davies 1980).

In addition to variations in wave energy, there are marked contrasts in tide range around the Australian coast (Fig. 1). In the southern part of the continent, from North West Cape around to Fraser Island, mean spring tide ranges are generally microtidal (<2 m), rising to mesotidal (2–4 m) in Spencer Gulf and Gulf St. Vincent in South Australia and in Western Port Bay and Corner Inlet in Victoria. In southwestern Australia the spring tide range is less than 0.5 m, and it can be difficult to detect tidal oscillations because wind stress and barometric pressure variations raise and lower the sea surface by larger amounts. On the Queensland coast and in the southeastern part of the Gulf of Carpentaria, tide ranges are typically mesotidal, but the northwestern coast between Port Hedland and Darwin is macrotidal (>4 m), with a maximum of 10.5 m in Collier Bay, near Yampi Sound.



Australia, Coastal Geomorphology, Fig. 6 Mangroves advancing seaward on the coast of Cairns Bay, north Queensland. An outer zone of *Avicennia* is backed by *Rhizophora* and then a varied mangrove community (Photo: E.C.F. Bird, Copyright-Geostudies)

Where tide range is small, wave energy is concentrated within a narrow vertical range, and where it is large, wave energy is more dispersed. Thus, the tidal variation emphasizes the contrast between high wave energy conditions in the southern part of Australia and low wave energy conditions along the northern coast. Large tide ranges also generate strong ebb and flow currents, which impede the deposition of wave-shaped structures, such as spits and barriers, and lead to the formation of wide funnel-shaped estuaries, such as that of the Fitzroy, opening into King Sound. Broad sandflats and mudflats are exposed at low tide, and, with diminished and intermittent wave action above mid-tide level, salt marshes and mangrove swamps become more extensive than on microtidal coasts, where the small tide range compresses the biological zonations.

Biological Factors

The effects of organisms on coastal processes are limited by their geographical range, conditioned mainly by climatic and oceanic factors. Reef-building corals live in Australian coastal waters from Houtman Abrolhos in the west around to the Capricorn Group at the southern end of the Great Barrier Reef. They are most prolific off the northeast coast, while off the northwest they give place to bryozoan and oyster

banks, and off the west coast, the reefs are submerged dune calcarenite ridges with only a veneer of corals (Fairbridge 1967).

Mangroves extend as far south as Corner Inlet ($38^{\circ}51'45''\text{S}$) adjacent to Wilsons Promontory in Victoria. In the humid tropical sector of northeast Queensland, there are up to 27 mangrove species in wide swamps backed by rain forest vegetation (Fig. 6), but in drier sectors there are fewer species, and the seaward mangrove fringe is backed by salt marshes and hypersaline mudflats. The effects of mangroves on sedimentation vary from one species to another, but the white mangrove (*Avicennia marina*), the most widespread of the Australian mangrove species and the only one extending as far south as Victoria, has a pneumatophore network that acts as a sediment filter and is consequently an agent of land building alongside estuaries and embayments sheltered from strong wave action (Bird 1993). Introduced *Spartina anglica* has re-shaped the intertidal profiles of Andersons Inlet, a Victorian estuary, and the Tamar estuary in Tasmania, by developing a depositional marshland terrace, but it does not grow in warmer environments farther north.

Shelly beaches (as distinct from beaches of shelf-derived biogenic carbonate sand) are confined to coastal sectors where shell-bearing organisms live abundantly, notably on rocky shores and reefs and within estuaries and lagoons. These are

extensive toward the heads of the South Australian Gulfs and on the western shores of Port Phillip Bay, where their dominance reflects the local absence of other types of beach material on a coast formed by the marginal submergence of a basaltic plain (Bird 1993).

Inherited Features

Australian coastal landforms include various features inherited from past phases of contrasted environments. Some of these are related to changes of sea level during Quaternary times: the submergence of valley mouths and the shoreward drifting of sand from the sea floor to form

beaches, barriers, and coastal dunes have already been noted. In addition, there are coastal features formed during past phases of higher sea level, some of which have persisted from Pleistocene times, while others may be referred to a higher Holocene sea-level episode. These features include emerged shore platforms, some with associated fossil beach deposits backed by former cliffs now degraded to bluffs; emerged coralline reefs; emerged sandy thresholds near the mouths of estuaries and lagoons; and beach ridge and barrier features standing higher, in relation to sea level, than they did at the time of their formation.

The interpretation of some of these features has been questioned. It is necessary to distinguish between emerged shelly beaches and the shelly aboriginal kitchen middens

Australia, Coastal Geomorphology, Fig. 7 The intertidal zone in King Sound, Western Australia, has branching tidal creeks fringed by mangroves, desiccated sandy mudflats, and a serrated high tide shoreline where desert dune ridges and swales have been submerged by the sea (Photo: E.C.F. Bird, Copyright-Geostudies)



frequently encountered above high tide level and to be sure that apparently emerged features were not formed when storm surges or tsunamis carried wave action briefly to higher levels. Coastal features that have undoubtedly emerged may be the outcome of tectonic uplift of the land margin, a fall in sea level, or some combination of the two. Emerged features dating from Pleistocene times often show evidence of dissection by fluvial incision or wind scour extending below present sea level during the last phase of lowered sea level, as on the inner barrier of the Gippsland Lakes (Bird 1993), whereas emerged features of Holocene age do not show such dissection.

On parts of the Australian coast, the landforms bear the imprint of weathering under past phases of climate that were warmer, cooler, wetter, or drier than at present. These include lateritic weathering profiles in cliff exposures of Tertiary rock in Victoria, inherited from a Pliocene phase of warmer and wetter climate, and the presence of periglacial deposits on cliffs in southernmost Tasmania, inherited from a Pleistocene phase of colder climate. On the east coast of the Cape York Peninsula near Cape Flattery, extensive systems of elongated parabolic dunes, now largely covered by scrub and forest, originated under drier and perhaps windier conditions in the past. Patterns of coastal rock weathering northward and southward of the wettest sector of humid tropical coast in northeast Queensland indicate that the warm and wet environment has been at times more extensive than it is now. The

high parabolic dunes on Fraser Island must also have formed under different conditions in the past, for they now carry a cover of rain forest. It is possible that sand mobilization here was also partly related to phases of rising sea level, when vegetated dunes deposited during earlier phases of stable or falling sea level were trimmed back by marine erosion and exposed to the effects of onshore wind action.

The shores of King Sound include narrow sandy peninsula ridges and intervening corridor embayments that are features inherited from an arid landscape of sand ridges and swales that has been partly submerged by the Late Quaternary marine transgression (Fig. 7).

Shore Platforms

Shore platforms on the Australian coast show a variety of forms, some sloping seaward and others almost horizontal. Some are related to coastal structures, especially where they coincide with exhumed bedding planes on resistant strata, but many transgress shore structures. Seaward sloping platforms are essentially abrasion ramps, but subhorizontal platforms have been influenced by shore weathering processes which have tended to flatten and maintain platforms originally cut by abrasion. Subhorizontal platforms are well developed at about high tide level on cliffed coasts cut in basalt and fine-grained



Australia, Coastal Geomorphology, Fig. 8 Cliffs cut in soft Miocene limestone on the Port Campbell coast in Victoria, with some of the outlying stacks known as the Twelve Apostles (Photo: E.C.F. Bird, Copyright-Geostudies)

sedimentary formations, such as sandstones and shales, which are subject to disintegration by wetting and drying processes (and possibly also by salt crystallization) down to a level of saturation by sea water. These are well developed on sandstones and mudstones on the New South Wales coast south of Sydney.

On dune calcarenite coasts, similar platforms have developed at a lower level, equivalent to the level attained by fringing reefs built upward by corals. Their flatness is evidently the outcome of solution processes effective down to the level at which precipitation of carbonates becomes dominant, indurating the platform surface, a process that has been demonstrated on the Nepean Peninsula in Victoria (Fig. 2). Where waves are supplied with sand or gravel derived from cliff erosion, fluvial outflow, or sea floor sources, abrasion becomes more effective, and the subhorizontal benches give place to platforms that slope seaward.

Modern Dynamics

Erosion is predominant on Australian beaches, even on the shores of barriers and beach-ridge plains that prograded earlier in Holocene times, and most beaches are backed by cliffed dunes. Exceptions are found on growing sectors of spits and cusped forelands, on deltas still receiving fluvial sediment (notably in Northern Australia), and on sectors where beach sand has accumulated alongside protruding structures, such as harbor breakwaters. There are also some sectors, notably on the western coast of Western Australia, in Streaky Bay in South Australia, off northwest Tasmania, and near Corner Inlet in Victoria where sand is still moving in from nearshore shoals to maintain or prograde beaches (Sanderson et al. 1998).

The prevalence of beach erosion could be due to a slight rise in sea level in recent decades or to increased storminess in coastal waters, but the most likely explanation is a general reduction in the supply of sand from the sea floor, which had previously nourished beach and barrier progradation. Other factors have contributed locally, such as the reduction of fluvial sand supply following dam construction in the Barron and Burdekin Rivers in North Queensland and the diminution of sand supply where the longshore sand supply has been intercepted by breakwaters, as at Sandringham on the coast of Port Phillip Bay (Bird 1996).

Cliffed coasts have continued to recede, especially on sectors where soft rock formations confront stormy seas, as on the Tertiary sands and clays of the Port Campbell district in Victoria (Fig. 8). Minor progradation has been detected on some mangrove-fringed coasts, as in Cairns Bay, Queensland, and on salt marshes, particularly where *Spartina* grass has been introduced. Elsewhere there is evidence of reduction of

mangrove and salt marsh fringes, as in Western Port Bay, Victoria, where there has been drainage of swampland and construction of boat harbors.

Human Impacts

Coastal morphology in Australia has been modified in various ways by human activities, particularly in the vicinity of urban centers, where ports have been constructed, seawalls built to halt coastline recession, cliffs stabilized, and groins inserted in an attempt to retain a beach fringe. A substantial part of the formerly cliffed northeast coast of Port Phillip Bay, on Melbourne's Bayside, now consists of artificial bluffs lined by such structures, and depletion of the natural beach fringe has been offset by artificial beach nourishment. Similar works have been introduced to the Adelaide coast, where the beach is now maintained by periodic renourishment using sand dredged from a prograded beach alongside a breakwater at the Outer Harbor.

Many coastal dune areas show evidence of vegetation loss and initiation or acceleration of blowouts as a consequence of clearing, burning, trampling, or vehicle damage to the dune vegetation. Attempts have been made to restore stability by fencing dune areas and planting vegetation, including the introduced European marram grass, *Ammophila arenaria*. Such activities as beach nourishment and dune stabilization mark a trend toward management of coastal morphology in Australia, with the aims of conserving many natural systems and their biota and providing recreational opportunities in the coastal environment.

Cross-References

- [Changing Sea Levels](#)
- [Cliffs, Erosion Rates](#)
- [Coastal Changes, Gradual](#)
- [Coastal Changes, Rapid](#)
- [Coastal Subsidence](#)
- [Deltas](#)
- [Glaciated Coasts](#)
- [Shoreline and Coastal Terrain Mapping](#)
- [Volcanic Coasts](#)

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B

Barrier

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Definition

A barrier (coastal barrier) is an elongated coastal ridge of deposited sediment built up by wave action above high tide level offshore or across the mouths of inlets or embayments.

Types of Barriers

A coastal barrier is usually backed by a lagoon or swamp, which separates it from the mainland or from earlier barriers. A barrier, thus defined, is distinct from a bar, which is submerged at least at high tide (Shepard 1952), and from reefs of biogenic origin (see ► [Coral Reefs](#)). Most barriers consist of sand, but some contain gravel as well as sand, and others consist entirely of gravel (shingle: see ► [Gravel Barriers](#)); marine shells (mainly mollusks) are major constituents of the Arabat barrier on the west coast of the Sea of Azov. Chesil Beach, on the south coast of England (Fig. 1), is a well-known shingle barrier, and similar features are seen on the southeast coast of Iceland and on the east and south coasts of South Island, New Zealand. Commonly the gravel has been derived from glacial or periglacial deposits, as on the north coast of Alaska and the southern shores of the Baltic Sea, whereas Chesil Beach is an accumulation of flint and chert gravel derived from the Cretaceous rocks of southern England.

Barriers are said to occupy about 13% of the world's coastline (Leontyev and Nikiforov 1965). They are most extensive on the Gulf and Atlantic coasts of North America and the ocean coasts of Australia, South Africa, and eastern South America, but they also occur on a smaller scale elsewhere, notably in Sri Lanka and New Zealand. Some barriers

are transgressive, migrating landward across lagoon and swamp deposits; others remain in position or are widened seaward by progradation, usually indicated by successively formed beach or dune ridges. Transgressive barriers occur where sediment is washed or blown over into backing lagoons or swamps, particularly during storms. Low sectors of a barrier through which storm waves or surges flow are called swashways, and sediment swept over a barrier through these is deposited as a washover fan on the inner shore. Shingle barriers may be permeable, so that storm surges drive water through them, to be deposited on their inner shore as lobate fans, known as cans on Chesil Beach, where they protrude into the backing lagoon, the Fleet.

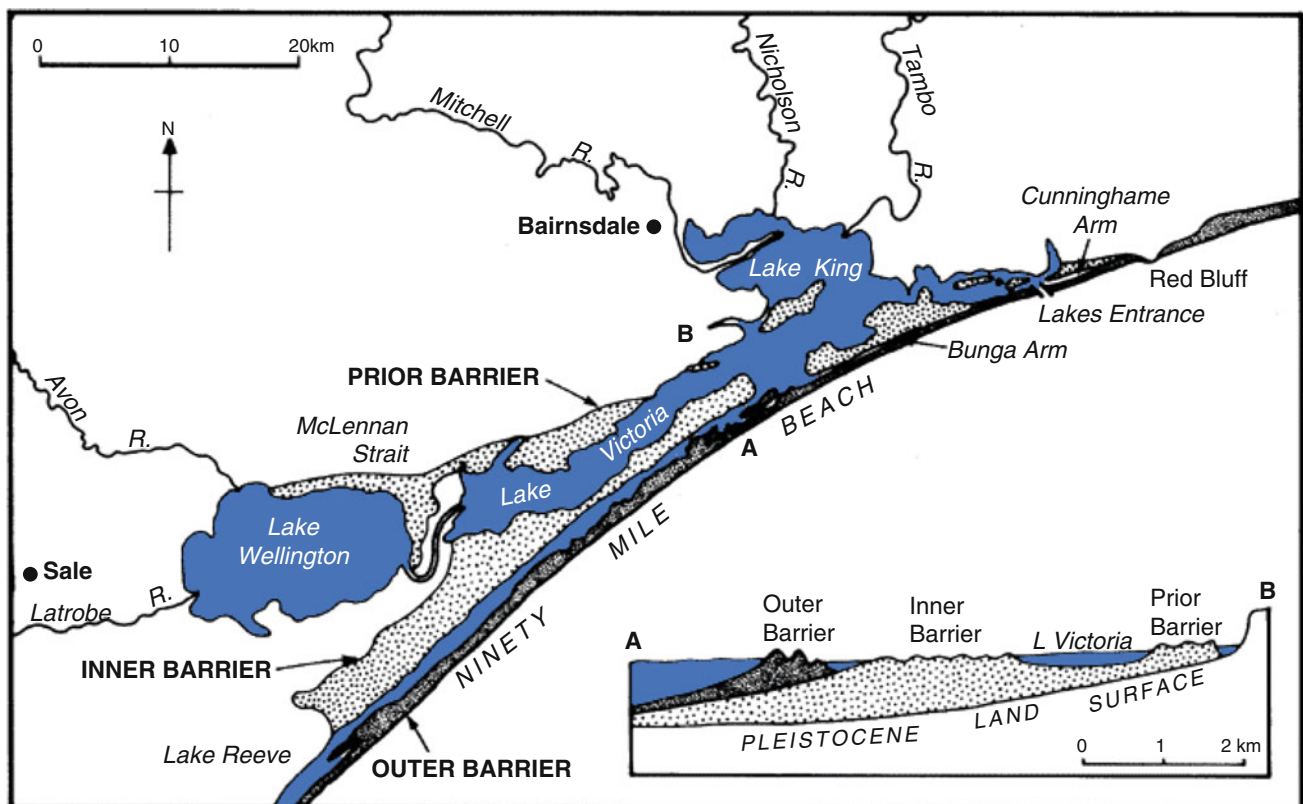
Multiple Barriers

On some coasts, there are multiple barriers, the inner and older barriers being of Pleistocene age bordered (and often partly overlain) by outer and younger barriers of Holocene age. Thus the Gippsland Lakes in southeastern Australia (Fig. 2) are enclosed by an inner barrier and an outer barrier (the Ninety Mile Beach), separated by lagoons and swamps, and with relics of an earlier, prior barrier that predates the enclosure of the existing lakes (Bird 1973). In this case, barriers have developed seaward of earlier marine coastlines, but evidence of preceding exposure to the open sea is not always present, particularly in lagoons where the enclosing barriers have been transgressing landward during a phase of rising sea level, as on the Siberian coast (Zenkovich 1967).

The term barrier beach indicates a single, narrow elongated ridge (usually less than 200 m wide) built parallel to the coast, without surmounting dunes. A barrier island is bordered by transverse gaps (tidal inlets, lagoon entrances, river outlets), which may be migratory and subject to closure; it usually bears beach ridges, dunes and associated swamps, and minor lagoons and may incorporate recurves (Schwartz 1973). Examples include Scoll Head Island on the east coast of



Barrier, Fig. 1 Shingle barrier, Chesil Beach, Dorset, England, with the Fleet Lagoon on the left. Copyright-Geostudies



Barrier, Fig. 2 The coastal barriers in the Gippsland Lakes region, with a cross section (A-B) showing the relationship between the Late Pleistocene prior and inner barriers and the Holocene outer barrier, resting upon a Pleistocene land surface. Copyright-Geostudies

England, the Frisian Islands on the Danish, German and Dutch coasts of the North Sea, and several along the Atlantic coast of the United States. A detailed study of the evolution of Gualan Island, a barrier island in the Outer Hebrides, included the effects of storm overwash (Dawson et al. 2012). A barrier with many interruptions becomes a barrier island chain. The term barrier island is inaccurate where there are no bordering transverse gaps. A barrier that is attached to the mainland at one end can be termed a barrier spit, as on the coast north of the Columbia River in Washington State (where Long Beach is a barrier spit partly enclosing Willapa Bay); one built across the mouth of an embayment to completely enclose a lagoon is a bay (baymouth) barrier (the Loe Bar in southwest England is an example). There are also mid-bay barriers and bayhead barriers, defined by their position.

In general, barriers are found on coasts where the tide range is small (as on the southern Baltic Coast and the Gulf of Mexico coast in Texas) and become chains of barrier islands where currents produced by larger tides maintain transverse gaps (as on the Danish, German, and Dutch North Sea coasts). Most barriers are straight or gently curved, but some bay barriers are short and sharply curved, as in the Orkneys, where ayres are shingle barriers built across the mouths of small bays enclosing lagoons (termed oyces) (Bird et al. 2003).

Fetch-Limited Barriers

Pilkey et al. (2009) drew attention to fetch-limited barrier islands, which have formed in sheltered waters such as estuaries or bays, in contrast with open-ocean barrier islands exposed to strong wave action. Many are actively evolving, subject to erosion or accretion, or migrating; others are inactive, surrounded by salt marsh or mangroves. They are smaller than open-ocean barrier islands, averaging roughly 1 km long and 50 m wide and 1–2 m maximum elevation, generally subject to overwash, with little dune development and often stabilized by enclosing salt marsh or mangroves.

Evolution of Barriers

Barriers can form in various ways (Schwartz 1971): multiple causalities being related to the nature and supply of sediment, the transverse profile of the coast, tide range, wave conditions, and relative sea level change. Some barriers may have formed by the emergence of nearshore swash bars as sea level fell (e.g., Knotten, on the Danish island of Laesø), but many developed during and after the Late Quaternary marine transgression by submergence of preexisting sand ridges and the shoreward drifting of seafloor sediment. Of these, some are still transgressive (as on parts of the Atlantic coast of the

United States), while others have become anchored and widened by progradation (as on parts of the southeast Australian coast). The landward transgression of barriers has been demonstrated in recent years on parts of the coast of the Caspian Sea, which has risen about 2 m since 1977 (Kaplin and Selivanov 1995).

Barrier spits may show features indicative of longshore growth, such as former recurved terminations on the landward side (as on Orford Ness in England or the Langue de Barbarie in West Africa), but others have been built and widened by wave-deposited sediment from the seafloor (as on Clatsop Spit in Oregon), and many result from combinations of onshore and longshore sediment drifting. In New Zealand, there are shingle barriers that have prograded intermittently, forming multiple parallel beach ridges as on the Tiwai barrier in the far south of South Island, bordering Awarua Bay in Foveaux Strait.

The shaping of barriers can be traced with reference to patterns of beach and dune ridges indicating stages in their growth and from their stratigraphy, which may indicate phases of upward growth, landward movement, and seaward progradation, as in Van Straaten's (1965) classic study of barriers on the Netherlands coast and Thom's (1984) study of sand barriers in eastern Australia. Barriers of unconsolidated sand are readily reshaped by wave and wind action, but where barrier sediments have become lithified (e.g., the Pleistocene dune calcarenites in Australia and elsewhere), they are more durable and may show cliffing (as in the inner barriers of the Coorong in South Australia). Some barriers incorporate segments of preexisting terrain, such as the glacial moraines on Long Island in New York, Walney Island in northwest England, and Sylt on the German North Sea coast.

Summary

Barriers are coastal ridges of deposited sand or shingle built up by wave action above high tide level offshore or across the mouths of inlets or embayments. They typically enclose coastal lagoons. Some are transgressive (migrating landward); others have been widened by progradation, often which successively formed beach or dune ridges. Multiple barriers may include inner and older barriers of Pleistocene age bordered by outer and younger barriers of Holocene age. A barrier beach is a single, narrow elongated ridge built parallel to the coast, while a barrier island is bordered by transverse gaps which may be migratory and subject to closure and usually bears beach ridges, dunes, and associated swamps. Barrier spits are attached to the mainland at one end; bay barriers have been built across the mouth of an embayment to completely enclose a lagoon, and mid-bay and bayhead barriers are defined by their position.

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Barrier Island Formation and Development Modes

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Barrier Island Definition

Barrier islands are shore-parallel, wave-constructed, usually elongated nearshore landforms, surrounded at the time of their formation on all sides by *subtidal* water bodies and often capped by eolian beds. Passes (inlets) flank barriers and link lagoonal backbarrier basins to the open sea. However, even if their lagoon is drained during low tide stages (e.g., Frisian Islands, North Sea), the barrier designation of mesotidal islands is justified (Otvos 2018). Facing high wave energy and salinity condition on the ocean side, the islands act as *barriers* that protect low-energy brackish lagoons and bays associated with nutrient-rich muddy deposits. Wetlands may subsequently fill the shallow lagoons.

Shifts in remotely-sensed vegetation communities prove to be sensitive indicators of even subtle topographic and geomorphic changes in the island interiors (Carter et al. 2018).

Traditional and proper barrier island definitions (e.g., Price 1951; Oertel 1985) exclude strandplains that directly front wetland-covered delta plains, at sites where subtidal lagoonal-bay basins never existed between beach ridge plains and marshy delta plains landward. Elongated, so-called “fetch-limited” islets located inside semi-enclosed estuarine bodies (lagoons and bays) cannot be considered barriers either, even if developed by ridge progradation (Otvos 2010). Refuting widespread earlier views, elongated-narrow, short-stubby, and other island shape categories do not statistically correlate with supposedly associated tide-dominated, respectively wave-dominated, or mixed energy regimes (Mulhern et al. 2017).

Genetic Categories of Barrier Island Formation

Mainland Foredune Detachment (“Submergence”)

Contrary to the coastal literature, McGee (1890, p.443) did not suggest island formation by coastal ridge engulfment during transgression - only the widening of preexisting lagoons behind islands that underwent slower landward shift than the mainland shoreline during the late Holocene. The idea of Frisian Coast barrier island formation from partially drowned dunes was first advocated between 1899–1901 by Haage and Wahnschaffe (see: Ehlers 1988, p. 23). In North America, Ganong (1908), an eminent Canadian botanist and naturalist may have suggested this idea first (Johnson 1919). The “barrier debate” of the late 1960s–early 1970s in part was related to the proposition that barrier islands form when the transgressing sea encroaches on, surrounds, and partially submerges tall dune ridges in the backshore zone (Hoyt 1967). Without microfossil-based documentation of the coastal depositional facies obtained from drill cores, Hoyt alleged the absence of preexisting marine deposits under the backbarrier lagoons. Lagoons landward of the North Carolina “Outer Banks” barriers (Davis 1994), or the Chandeleur chain in Louisiana (Otvos 1986), illustrate among other examples, that open marine deposits are indeed absent in the rear of landward retreating transgressive islands. The contrast with aggradational island formation is instructive. Foraminifer-rich euhaline marine deposits along the gulfward margin of the Mississippi Sound, occur at numerous sites directly beneath and immediately landward of regressive, laterally, and seaward-prograding barrier islands. During emergence and westward growth of the Mississippi-Alabama island chain the Sound was separated from the Gulf. (q.v., ► “Gulf Shorelines, Last Eustatic Cycle” entry; Otvos 2018). The island chain directly overlies a shallow subtidal sand platform belt that runs generally parallel with the mainland shoreline.

The shallow subtidal platform from which the islands have emerged, in turn, is directly underlain by such muddy open marine Gulf sediments. Subrecent sediments deposited under full- or near full marine salinity conditions may extend for some distance landward under the more recently deposited sandy lagoon muds (Otvos 1981).

Following Hoyt's suggestion (1967), island formation by encroachment of the sea was accepted as the underlying cause for formation of the Last Islands (Isles Dernieres) chain. Delta plain subsidence, erosion of the shore, expansion and merger of delta interior lakes gradually detached a large strip of marshland from the delta hinterland. According to the Hoyt hypothesis, when encroached and surrounded by the transgressing ocean, high dune ridges in the backshore zone become permanent barrier islands that undergo morphological changes and landward shift. After engulfed by the transgressing sea the dunes somehow avoid wave destruction and drowning. This hypothesis was incorporated also into a model of barrier island development along disintegrating delta plains (Penland et al. 1985). Despite obvious and basic contradictions (Otvos 1970a, b; 1986), both models and suppositions are still widely cited in the literature, including Davis and FitzGerald (2004, Fig. 8–15).

Isle Derniere, the original delta-fringe marsh island, evolved differently than portrayed. As typical of sand-starved delta shores, a narrow, low, thin Gulf beach-fronted the detached marsh island (US Army 1962). The Gulf beach of the island was a mere shelly sand veneer over muddy delta marsh deposits. The insignificant eolian sand accumulation runs counter to Hoyt's own requirement for tall eolian dune ridges, capable of surviving wave erosion during partial drowning. Furthermore, certain members in the chain of small islands that emerged at the marsh island site originally displayed a distinctly progradational strandplain morphology. Several islets, therefore, are not eroded fragments of a disintegrating delta plain but thoroughly reconstructed sandy shore landforms. They arose from reworked deposits of deltaic lithosomes in the Isles Dernieres area. Small sandy islands with developing ridge-and-swale (strandplain) topography emerged from the Gulf floor, then prograded obliquely toward (Otvos 2018) the delta-mainland. Consequently, no genetic or morphological links exist between the erosion-detached, fragmented Isles Dernieres delta plain marsh islands and the thin sand veneer along their Gulf beach, on one hand, and the miniature strand plain islands that emerged offshore, on the other.

Interestingly, the search for islands with transgressive encroachment history did turn up an unusual example for the subsidence-inundation mode of barrier island and barrier spit formation. Thermokarst-induced coastal plain subsidence and subsequent wave action without any dune-mediation did result in small ephemeral barrier islands and miniature barrier spits along a permafrost-dominated Arctic seashore (Ruz et al. 1992).

Spit Detachment and Segmentation Model

Although frequently cited in the coastal literature, Gilbert (1885, p 87-88) did *not* suggest barrier island formation by erosional barrier spit segmentation. Gripp (1944) advocated a similar but somewhat obsolete hypothesis for the North Sea coast (see: Ehlers 1988). Without citing field examples, Fisher (1968) proposed spit detachment and segmentation as a common mode of island genesis. However, only very few small, highly vulnerable examples were documented for island formation. These landforms include low, narrow barriers in New England (e.g., Monomoy Island-Nauset Spit, Giese 1988; Sandy Point-Napatree Beach, Fisher 1981), the melting permafrost-ridden Arctic Ocean shore of Canada (Ruz et al. 1992), and Po Delta spits, Italy (Simeoni et al. 2007). Guided by preexisting topography, barrier spits were associated with and/or anchored in eroding Pleistocene headlands during retreat of the Holocene shoreline in the course of the late Quaternary transgression. High headlands represented the antecedent topography on the Atlantic and other trailing-edge passive margin coasts. While barrier islands of this origin are not known from the Gulf, paradoxically, the reverse is true. Pelican-Sand Barrier Spit on the Mobile Bay Ebb-delta and St. Joseph Barrier Spit on the Apalachicola Coast, did form by barrier island attachment to large Dauphin Island, respectively, to the mainland shore (Otvos 2018).

A Controversial Model for Deltaic Barrier Formation: Combining Marine Encroachment with Spit Growth and Segmentation

Following the scheme proposed by Penland et al. (1985), Davis and FitzGerald (2004, p. 165), Pilkey and Frasier (2003), Davidson-Arnott (2010; Fig. 10.27), and several other workers assumed that relict delta headlands in the disintegrating Mississippi River at St. Bernard and Lafourche delta plain complexes of Louisiana provided anchoring and abundant sand source for shore-parallel barrier spit growth. After erosional segmentation, the barrier spits became island chains. This hypothesis, despite its speculative nature and lack of foundations in fact is still very popular and widely cited in the coastal literature.

Headland-tied, storm-segmented barrier spits were considered intermediate stages in the formation of the Chandeleur and other delta plain barrier islands. However, no sizable barrier spits tied to delta headlands occur anywhere along the shores of the disintegrating Mississippi Delta Complex. Intensive, deep ravinement erosion during the latest Holocene transgression over the relict delta lobes would have eliminated all such traces of former headlands and barrier spits east of the Chandeleurs a long time ago (Otvos 1986, 2018). In the south of the Holocene Mississippi delta complex, the minimum age of the severely truncated relict Moreau-Caminada strandplain (<460 yr B.P.; q.v., ► “Gulf Shorelines, Last Eustatic Cycle” entry) indicates that the Mississippi-Lafourche

subdelta headland shore, from which an “island-producing” sand spit may have prograded between ca. 350–150 yr B.P. was located several km offshore from the current shoreline.

Enclosing an estimated 20–40 million m³ of subtidal-to-supratidal sand at its full development, the Chandeleur chain was ~74–80 km long and 1–2 km wide by the mid-nineteenth century. Had a large sand-enriched delta headland sprouted shore-parallel barrier spits capable of surviving storm and ravinement destruction, such are absent from the present island chain and relict delta plain areas. In an entirely different setting, however, driven by cross-shelf and northward-directed island-parallel littoral transport of large volumes of reworked delta and inner shelf deposits, a sizable sand lithosome did accumulate rapidly at Hewes Point, in the north of the Chandeleur chain (Miner et al. 2009). Utilized as the sand source in the construction of a protective “engineered berm” structure (FitzGerald et al. 2015), the formation conditions of this large shoreface sand body were unrelated to island emergence processes.

Neither do the two Timbalier Islands (Fig. 1a), spit relicts, or recently formed short and ephemeral West Belle Pass barrier spit west of the Moreau-Caminada Lafourche Delta (headland in south Louisiana (Kulp et al. 2015) provide valid examples for segmentation of headland-attached spits. Map documentation (Williams et al. 1992; Penland et al. 1994, 2003; Otvos 2018) is lacking to indicate that East Timbalier Island, downdrift from the eroding, fast retreating sandy ridgeplain headland, evolved from an ancient barrier spit. Rapid delta island destruction and dispersal of their relatively scant sand content renders the suggested formation of large, massive offshore sand bodies (e.g., Ship Shoal) in the “submergence” stage of the Penland island-development cycle even less plausible.

Island Emergence by Shoal Aggradation; Breaker-Erosion Avoided

Wave erosion in the breaker zone has long been thought as preventing island emergence from aggrading sand shoals. However, this obstacle is eliminated when island-ward the destructive breaker zone sand transport by constructive swash currents and solitary (translation) waves may reach the low-energy shoal interior. *It is suggested that this is the process, reenforced by occasional non-erosive washover from the open sea, that in the absence of erosive storms builds up shallow shoal centers to slightly above highest tide level.* Subsequent lower tides surround emergent embryonic islands with a potential of further horizontal and vertical expansion.

In advocating island emergence by shoal aggradation, deBeaumont (1845, was the first to propose a theory for barrier island formation; specifically, by shoal aggradation. Gilbert (1885, p. 87) assumed that while forming lagoons

landward, islands may rise above lake level from breaker zone sand bars. Ordemann (1912) suggested a similar mechanism for North Sea islands. Two Frisian barriers, Memmert and Trischen are among the earliest published examples for island aggradation from shoals (Keilhack 1924; see: Ehlers 1988, 1990). Numerous authors subsequently questioned the importance of this formation mode (e.g., Hoyt 1967; Fisher 1968; Davidson-Arnott 2010). They maintained that this origin is valid only for a handful of small, ephemeral islands, an insignificant fraction of barrier islands worldwide. However, while field proof for island genesis was generally eliminated during ravinement erosion, enough evidence remained from the late Holocene highstand stage to question such a sweeping assumption. At the same time, only rather limited proof exists for other modes of island genesis.

Following episodes of relatively rapid landward overstep during the Holocene transgression, newly established barrier chains did form over open marine sediments along the seaward boundaries of newly established lagoons (q.v., ► “[Gulf Shorelines, Last Eustatic Cycle](#)” entry.) Unless traces of antecedent headlands and linked barrier spits are found in the geological record to prove island origins by barrier spit segmentation, aggradational island growth still appears as the chief mechanism of island formation. Only along the landward flanks of barrier islands would storm overwash deposits rest on lagoonal sediments. For example, *Galveston Island*, Texas, directly overlies mostly full-salinity *open marine* deposits (Bernard et al. 1970). *Matagorda Island* started as an intermittently emergent offshore sand shoal. It has migrated landward with the late Holocene transgression and became stabilized as the Gulf of Mexico reached relative stillstand after ~4 ka B.P. (Wilkinson 1975). Based on microfossil data, stratigraphic sections across the Mississippi Sound and the Mississippi barrier islands indicate gradual aggradational emergence over open marine muddy-sandy and higher, sandy *open marine* shoal-platform deposits (Otvos 2018). Numerous small and sizable islands emerged repeatedly during prolonged fair-weather intervals in the past century off the central peninsular Florida Gulf coast (Davis et al. 2003); between 1869–1917, 1969–2005, and on a massive scale during the relatively storm-free decade since 2008 along the relict Mississippi-St. Bernard delta shores; at various times in the Alabama-Mississippi island chain, in the Mobile Bay ebb-delta, and worldwide sites (e.g., Otvos 1970a, b; 1981; Otvos and Giardino 2004; Davis et al. 2003). Pelican-Sand Island, located on the ebb delta had a varied emergence-submergence history. The most recent, minuscule berm basin- islands have emerged in 2015 (Fig. 30 in: USGS 2017; Otvos 2018). Islands in the original long Louisiana-Mississippi chain were also of aggradational origin (Otvos 1981, and q.v., ► “[Gulf Shorelines, Last Eustatic Cycle](#)”, ► “[Coastal Barrier Preservation and Destruction](#)” and ► “[Beach Ridges](#)” entries).

Composite Barrier Islands

Composite islands representing late Holocene barrier deposits, each surrounding a pre-Holocene ridge core, were designated as “erosion remnant islands” in the “sea island” chain of coastal Georgia and South Carolina (Zeigler 1959). Unlike in the encroachment hypothesis that involves partial drowning of high Recent shore dune ridges, the old relict ridges in the composite style of island development mode are semiconsolidated Pleistocene or older landforms. They don’t occur as a long, continuous island chain. After rising ocean waters enclosed remnants of Pleistocene interglacial barriers, prograding Holocene strandplains became attached to them. The pre-Holocene topographic lows behind the relict Pleistocene islands were gradually filled by Holocene wetlands, populated by a dense network of narrow meandering tidal creeks. Holocene barrier islands, anchored at shallow depths in Neogene bedrock, dot the Gulf coast of the Florida Peninsula (Evans et al. 1985; Davis et al. 2003). Dauphin Island, Alabama, formed as Holocene transgression surrounded a Last Interglacial Pleistocene barrier sector that became the shunting focus for the westward-directed littoral drift. This sediment flow has maintained the Alabama-Mississippi island chain ever since (q.v., ► “Gulf Shorelines, Last Eustatic Cycle” ► “Coastal Barrier Preservation and Destruction” and ► “Beach Ridges” entries).

Barrier Dynamics of Transgressive and Highstand Stages During the Last Eustatic Cycle (Table 1)

Transgressive (“retrogradational”) islands are low, narrow, thin barriers that contain relatively small volumes of sand and/or gravel volumes. In the face of rising sea level, insufficient longshore and cross-shelf onshore sediment supply prevents their *in situ* stabilization. Landward storm overwash episodes that result in cross-island washover channeling, lowering, and destruction of dune fields drive islands by the “rollover” mechanism over lagoon-side estuarine-backbarrier deposits. Dillenburg and Hesp (2009) claim that Tramandai Barrier, SE Brazil, experienced uninterrupted shoreward shift since 10 ka B.P. This island category is also very common on the glaciated, sediment-starved New England shores and along the Outer Banks barriers of North Carolina. Transgressive islands represent 17% of the Georgia Bight shoreline; in certain shore compartments, this value reaches 68% (Hayes 1994). On the Texas shore, 80% flanked by barriers, Matagorda Peninsula and South Padre Island belong in this category (Table 1; Morton 1994; Anderson et al. 2014).

Although formed mostly by progradation, because of their downdrift-directed shift, several *prograded* strand plain barrier islands of the Mississippi Delta (e.g., Timbalier, East Timbalier islands (US Geological Survey 2016); Breton Island Fig. 1) do perform as *regressive* islands in their steady

Barrier Island Formation and Development Modes, Table 1 Barrier island development modes (Revised from Otvos (1981), Galloway and Hobday (1983), Morton (1994), Anderson et al. (2014), and other sources)

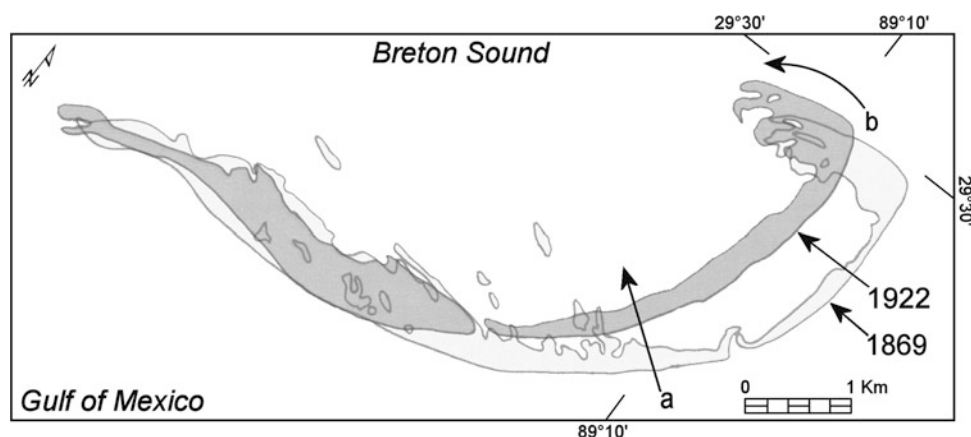
1. **Stable.** Active islands without significant long-term lagoonal/seaward or lateral (updrift and downdrift) shore changes. Balanced in terms of long-term erosion-accretion budgets. Lacks vertical aggradation mode. Only occasional overwash occurs with temporary shore erosion, followed by fair-weather restoration of the lagoonal and ocean shores (e.g., Santa Rosa Island, northwest FL).
2. **Regressive.** Seaward- and/or downdrift-directed ridge plain progradation, occasional vertical aggradation, and localized overwash episodes (e.g., Galveston Island; North Padre Island, TX; and St. Vincent Island, FL.) Locally and/or periodically may alternate with transgressive island development phases.
3. **Transgressive.** Landward advance over backbarrier-lagoonal or bay deposits
 - (a) Landward-shifting island shore parallels mainland shore. This process represents landward rollover-style retreat of seaward island shore (e.g., most recent development mode of Follets Island, TX, pre-twentieth century Chandeleur Islands, LA, numerous Georgia Bight and North Carolina “Outer Banks” islands and island sectors).
 - (b) In part regressive- in part transgressive. Advance by downdrift progradation toward mainland; island migration parallel or at oblique angle to mainland shore. Updrift island sectors erode. Represented by early history of Mississippi-Louisiana island chain and currently by Ship, Horn, Petit Bois islands, MS, Timbalier, and E. Timbalier, and other formerly prograded, during the past half century particularly heavily disintegrating strandplain barriers in the “Last Islands” (Isles Dernieres) island chain, LA.
4. **Aggradational.** Almost exclusively upward growth (Mustang Island, TX). Transitional (e.g., Central Padre Island) toward aggradational-regressive mode.
5. **Inactive.** Drift- and shoaling-sheltered, gradually eroding relict barrier islands without access to significant nourishment by outside sand supply (Cat Island, MS).

advance toward the mainland delta plain shoreline (Fig. 3 in Otvos 1970a and Fig. 9 in Otvos and Carter 2013). Strand plain islands of the prograded Louisiana-Alabama barrier chain have been also shifting westward toward Louisiana’s mainland shore.

Regressive islands and barrier spits tend to be wide and high landforms. They receive abundant sediment nourishment from offshore and longshore sources and generally display relatively high topographic profiles. These barriers experience substantial downdrift-directed and seaward progradation. Numerous northern Gulf islands, including 35% of Texas barriers, formed by this mode (Morton 1994).

Aggradational Growth Mode

Essentially vertical aggradational barrier island develops, without substantial seaward or lagoonward island progradation (Galloway and Hobday 1983). Influenced by fluctuations in sediment supply and hydrologic conditions, spatially and temporarily this mode often is transitional toward the aggradational-regressive island category.



Barrier Island Formation and Development Modes, Fig. 1 Transgressive landward shift of progradational Breton Island, SE Louisiana, between 1869 and 1922. (a) Rollover island sector; (b) recurved prograded sector (Otvos and Carter 2013). Concurrently with their transgressive landward shift, several prograding, regressive

barriers have advanced in a transgressive fashion landward; at an oblique angle or perpendicular to the mainland shore (e.g., previously the two Timbalier islands Last Islands, LA, St. Joseph Barrier, NW Florida, Sand-Pelican Island on Mobile Bay ebb delta, AL)

Transitional and Alternating Barrier Development Modes

Changes in sand starvation or abundance, and major storm episodes, may play a major role in fundamental barrier changes. Timmons et al. (2010) described a case of back-barrier erosion that transformed wide and high *regressive* Bogue Banks Island into a *transgressive* barrier. Horseneck Beach Island in New England, on the other hand, switched from transgressive to a regressive barrier (Ibrahim 1986). Reassessed by Anderson et al. (2014) as transgressive, Follets Island's development in Texas previously has been considered transitional between regressive and transgressive modes (Morton 1994; q.v., ► “Gulf Shorelines, Last Eustatic Cycle” entry). From an extensive mainland strand plain on the southeastern coast of Brazil, Hein et al. (2013) report on a ridge plain “switch interlude” from regressive to transgressive, followed by return to a regressive ridge development mode.

Summary

Of several proposed modes of barrier island initiation, vertical shoal aggradation appears to be by far the most common cause of island formation. Rarely influenced by localized interaction with antecedent headland topography, barrier spit formation and segmentation appear to have played at best a minor role during the transgressive phase. At the start of the present highstand stage, the antecedent topography led to formation of composite islands, anchored in Pleistocene or pre-Quaternary deposits. Of the transgressive, regressive, aggrading, essentially stable, and inactive modes of island development, the two progradation types represent transgressive, respectively, regressive barrier categories.

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Barrier Island Landforms

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Introduction

Coastal barriers are among the most dynamic landforms on Earth and the sites of some of the most beautiful beaches. These coastal accumulation forms are constructed by the combined action of waves, tides, and longshore currents as thin strips of land that build above sea level. They generally parallel the coast and gain vertical elevation through aeolian activity. They are called barriers because they protect the mainland coast and water bodies (backbarrier) from the direct forces of the sea, particularly during storms. They dampen the effects of storm waves, heightened tides, and salt spray. The bays, lagoons, marshes, and tidal creeks that form behind barriers provide safe harborages, nursery grounds for many marine organisms, and important sources of nutrients for coastal waters. Barriers occur in a variety of geomorphic settings but are most common where sediment supplies are abundant and hydrographic regime (wave and tidal forces) is conducive for onshore sand accumulation. This chapter describes the global distribution of coastal barriers, key landform types, morphology, and stratigraphy, as well as the main theories of barrier origin and the effects of hydrographic regime on the development of barrier coasts.

Physical Description

Coastal barriers (barrier islands, barrier spits, baymouth/bayhead barriers) are predominantly wave-built accumulations of sediment (sand and gravel) that accrete vertically due to wave action and subsequent modification by wind processes (aeolian action). Most are linear features that tend to parallel the coast, generally occurring in small groups or long chains. Isolated barriers are common along formerly glaciated (paraglacial) coasts (e.g., New England, Alaska, eastern Canada, northern Europe, Siberia, southern New Zealand) and along high-relief coasts such as those associated with geologically young and/or active continental margins (e.g., parts of Pacific and Mediterranean coasts and some oceanic islands). Barriers are separated from the mainland by a *backbarrier* consisting of tidal flats, shallow bays, lagoons, and/or marsh systems. Barriers may be less than 50 m wide or more than several kilometers in width. In general, barriers are

wide where the supply of sediment has been abundant and relatively narrow where erosion rates are high or in areas of chronic sediment deficit. Barrier length can range from small pocket barriers of a few 10s of meters to open-coast systems that stretch for more than 100 km. This is partly a function of sediment supply but is also strongly influenced by the wave/tidal balance of the region (Hayes 1979).

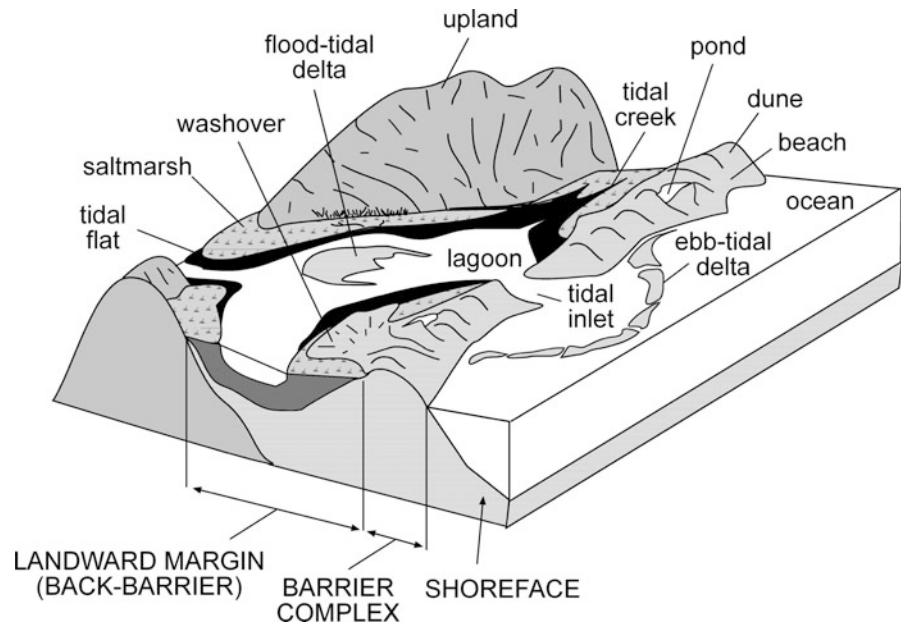
Depending on the regional geological and climatic setting, coastal barriers may have a suite of available sediment types. Sand, which is the most common constituent, comes from a variety of sources including rivers, deltaic, and glacial deposits; eroding cliffs; volcanic sources; and biogenic material. The major components of land-derived sand are the quartz and feldspar, both durable minerals that are a product of physical and chemical decomposition of granitic bedrock, the primary building block of continents. In high latitudes where glaciers have shaped the landscape, gravel is a common constituent of barriers (FitzGerald and Rosen 1987), whereas, in lower latitudes carbonate material, including shells and coral debris, may comprise a major portion of the barrier volume, also called the lithosome (Davis 1994). Along the southeast coast of Iceland, as well as parts of Alaska and Hawai'i, many barriers are composed of black volcanic sands derived from upland volcanic rocks or otherwise are dominated by skeletal remains of local organisms (bioclastic components). Along the Alaska Peninsula, Katmai Bay is fronted by barrier spits composed chiefly of pumice derived from the 1912 eruption of Novarupta.

Individual barriers comprise a suite of subenvironments and their arrangement differs from location to location reflecting the type of barrier and the regional setting (Reinson 1984). Generally, most barriers can be divided in three zones: the beach, the barrier interior, and the landward margin (Fig. 1).

Beach – The beach is the most dynamic part of the barrier consisting of the sediment shaped by waves. The pattern of sediment reworking shifts throughout the tidal cycle and the supratidal area (exposed section above mean high tide), when dry, undergoes aeolian reworking. Beaches exhibit a wide range of morphologies depending upon a number of factors, such as sediment grain size and abundance, influence of storms, and exposure to wind. Sediment is eroded from the beach during storms (moving offshore, landward, or alongshore), and the bulk is returned during more tranquil (fair-weather) wave conditions. During these storm and post-storm phases, the shape and volume of the beach evolve in a predictable fashion (see “► Beach Processes”).

Barrier Interior – Along sand-dominated barriers, the beach is backed by a frontal dune ridge (foredune), which may extend nearly uninterruptedly along the length of the barrier, provided the supply of sediment is adequate and storm incision is localized. This ridge is the first line of defense in protecting the interior of the barrier from the effects of storms. Landward of this region are secondary dunes that

Barrier Island Landforms,
Fig. 1 Coastal barrier sub-
 environments



have a variety of morphologies depending on the historical development of the barrier and subsequent modification by aeolian processes (Leatherman 1979). For example, devegetation of the Provincelands Spit at the tip of Cape Cod, Massachusetts, by early settlers led to the formation of large parabolic dunes that are up to 500 m long and more than 35 m high (FitzGerald et al. 1994). In other locations where barriers have built seaward (prograded) through time (e.g., North Beach Peninsula in Washington or Kiawah Island in South Carolina), former shoreline positions are marked by a series of semi-parallel vegetated beach/dune ridges reaching 3–7 m in height. In other parts of the world, these ridge plains (strandplains) may contain many tens of ridges separated by swales (Tanner 1995). The low-lying swales may extend below the water table and are the sites of fresh and brackish wetlands (bogs, ponds) or saltmarshes.

Landward Margin – Along the landward margin of many barriers, dunes diminish in volume and/or height and the low relief of the barrier gradually changes to an open-water area of a lagoon, bay, or tidal creek, and intertidal sand/mudflat or a coastal saltmarsh. Along retrograding coasts, where the barrier has been migrating onshore, its landward margin may be dominated by aprons of sand (washovers) formed by cross-barrier transport during storms. In time, these sandy deposits may be colonized by salt-tolerant vegetation producing saltmarshes that mimic the shape of washovers or relict flood-tidal deltas. In still other regions, a sandy beach may front the backside of a barrier. In this setting, often characterized by wave-generated sediment transport across the lagoon, wind-blown sand from the beach may subsequently form a rear-dune ridge outlining the landward margin of the barrier.

In alongshore direction, barriers may terminate in an embayment or at a headland. Barriers are separated by tidal

inlets along island chains. Inlets are the waterways that allow for the exchange of water between the coastal ocean and the backbarrier environments. The end of a barrier abutting a tidal inlet is usually the most unstable portion of the barrier due to the effects of inlet migration, erosion, and associated sediment transport patterns – this barrier/inlet system is considered one of the most dynamic of all landforms on Earth.

Global Distribution and Tectonic Setting

General – Barriers comprise approximately 10% of the global coastline. They are found along every continent (except Antarctica), in nearly every type of geological setting and in every kind of climate (Davies 1980; Reinson 1984; Davis 1994; McBride et al. 2013). Tectonically, they are most common along Amero-trailing Edge coasts (Inman and Nordstrom 1971; see below) where low-gradient continental margins provide ideal settings for sediment accumulation. The disappearance of barriers toward the “zero energy” northwest (“Big Bend”) coast of Florida in the Gulf of Mexico attests to the requirement of incident wave energy for their formation. Barrier coasts are best developed in areas of low-to-moderate tidal range (Tidal range is the vertical difference between high and low tide. A classification of coasts based on tidal range consists of microtidal coasts ($TR < 2.0$ m), mesotidal coasts ($2.0 \leq TR \leq 4.0$ m), and macrotidal coasts ($TR > 4.0$ m).) (microtidal to mesotidal range) and in mid-to-low latitudes (Davis 1994). Climatic conditions control the vegetation on the barriers and in back-barrier regions, the type of sediment on beaches, and in some regions such as the Arctic ice-push barriers, the formation and modification of landforms themselves (Davies 1980).

Geomorphic effects of large animals has latitudinal dependency, with sea turtles and pinnipeds dominating (especially in pre-historic times) the seasonal reworking at a number of beaches in low and high latitudes, respectively.

Tectonic Controls – The tectonic setting of the continental and insular margin dictate, to a large extent, the sediment contribution to the coast, the width of the continental shelf, and the overall distribution of topographic elements (hypsoetry) of the coast. The tectonic coastal classification of Inman and Nordstrom (1971) remains useful for illustrating these relationships and explaining the worldwide distribution of barriers. *Amero-trailing edge* coastlines tend to have abundant sediment supplies due to extensive continental drainage. Their low relief coastal plains and continental shelves provide a platform upon which barriers can form and migrate landward during periods of eustatic sea-level rise. The longest barrier chains in the world coincide with Amero-trailing edges and include the East Coast of the United States (3,100 km) and the Gulf of Mexico coast (1,600). There are also sizable barrier chains along the East Coast of South America (960 km), East Coast of Indian coast (680 km), North Sea coast of Europe (560 km), Eastern Siberia (300 km), and the North Slope of Alaska (900 km).

The unparalleled extent of barriers along the East Coast of the United States is undoubtedly a product of the erosion and denudation of the Appalachian Mountains. The massive volume of sediment derived over long geologic periods through multiple phases of glacial and fluvial transport produced the wide coastal plain and continental shelf region, much of which is veneered by layers of unconsolidated sediment or easily eroded sedimentary rocks. It is the ongoing reworking and redistribution of these surface materials by rivers, tides, waves, and wind that are responsible for extensive barrier construction along this region. Likewise, the nearly continuous chain of barriers in the Gulf of Mexico is related to the wide low-gradient coastal plain and continental shelf of this margin, with sediment delivery, particularly along Texas coast, being greater during the past phases of wetter climate. The barrier chain along The Netherlands and German North Sea are another example of barriers forming along a sand-rich coast.

Marginal Sea coasts contain a relatively small percentage of world's barriers even though these regions boast some of the largest fluvial sediment discharge in the world, particularly along the Asian margins. Whereas it would appear that many of these coasts have ample material for barrier construction, the irregular coastal topography leads to much of the sediment infilling submarine valleys rather than forming barriers. In addition, the discharge from many Asian rivers has substantial suspended sediment component that may inhibit the concentration of sand-sized material. In fact, many of the marginal sea coastlines containing barriers, such as the short chains along the Caribbean coast of Nicaragua and Honduras,

have no large riverine sources. The origin of these barriers is probably tied to the reworking of previously deposited shelf sediments, similar to many sections of the US Atlantic Coast.

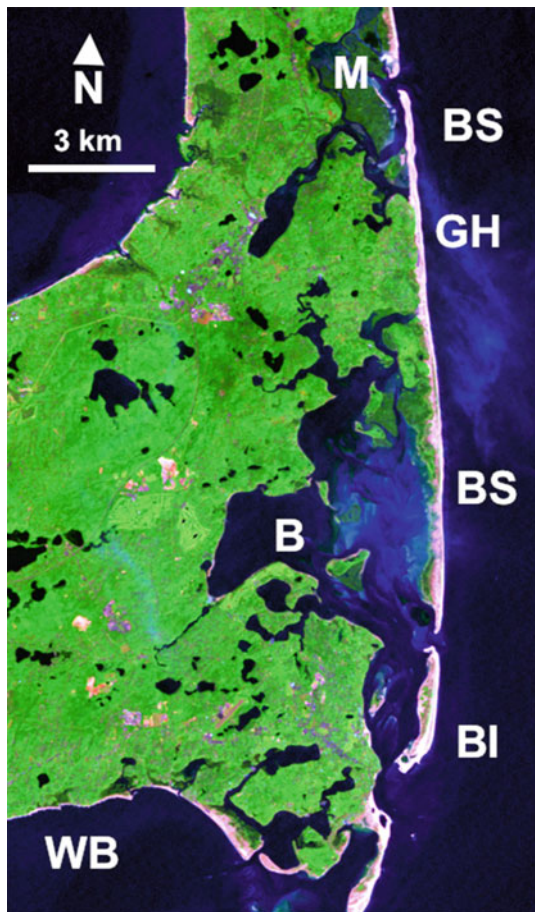
Leading-edge (Collision) coasts also have few barriers chains, but in this case, it is due to an overall lack of sediment. Most of the rivers discharging along these shores have small drainage areas and contribute little sediment for barrier construction. Additionally, the narrow (and thus steep) continental shelves result in high wave energy and rapid sediment dispersal along the coast. Along parts of the California coast some of the material that would otherwise accumulate along the shore is syphoned from the beaches to the deep ocean basins through submarine canyons, some having their heads within 10s of meters of the shoreline. Two of the longest barrier chains along the US Pacific Coast are located adjacent to eroding cliffs or major river mouths, including the southern Washington barriers near the Columbia River (Long Beach Spit of the Willapa Bay) and the Gulf of Alaska barrier chain fronting the Copper River delta.

Only a small percentage of the world's barriers are found along *Neo-trailing* and *Afro-trailing* and *edge coasts* due to an overall lack of sediment along these margins. There is little sediment delivery to the Neo-trailing edge margins of the Red Sea, Gulf of Aden, and Gulf of California due to the geological immaturity in the development of river drainage along their coasts and very low precipitation. Similarly, much of the northern coast of Africa has little river discharge of sediment due to the very arid conditions of the interior region and the recent damming of the Nile. The extensive barrier system (300 km) that exists along the Ivory, Gold, and Slave Coasts on the west-central coast of Africa is related to the numerous moderately sized rivers of this area, including the Congo River (Anthony 1995).

Climatic Controls – By controlling the amount of precipitation and evaporation of a continent, regional, and local climate exerts a strong influence on the size and number of rivers, as well as on the overall volume of sediment delivered to the coast. The drainage area is another important factor governing sediment discharge. Typically, larger drainage areas produce greater sediment delivery to the coast; however, there are exceptions. The Eel River in northern California releases almost twice the sediment load to the coast as the much larger Columbia River to the north, despite having a drainage area two orders of magnitude smaller (9,000 km² vs. 661,000 km²). The high sediment discharge of the Eel River is a product of the mountainous terrain it drains and a bedrock landscape that weathers easily producing abundant sediment.

Barrier Types

Coastal barriers are grouped into three major classes based on their connection to the mainland (bedrock, glacial landforms,



Barrier Island Landforms, Fig. 2 Examples of barrier morphotypes: BI – barrier island; BS – barrier spit; WB – welded barrier. Other coastal features include backbarrier bays (B) and saltmarshes (M). Note barrier attachment to a glacial headland (GH; outer Cape Cod, Massachusetts, USA; USGS Earthshots image)

etc.; Fig. 2). *Barrier spits* are attached to the mainland at one end and the opposite end terminates in a bay or the open ocean. *Welded barriers* are attached to the mainland at both ends and can be assigned to *bayhead*, *mid-bay*, or *baymouth barriers*, depending on their relative position within a bay. *Barrier islands* are isolated from the mainland by inlets at their ends and water bodies or wetlands in the backbarrier.

Barrier Spits

These landforms are most common along irregular coasts where angular wave approach and an abundant supply of sediment result in high rates of longshore sediment transport. These conditions promote spit building across embayments and a general straightening of the coast (Zenkovich 1967; Schwartz 1972). In some instances, spit construction partially closes off a bay forming a tidal inlet between the spit end and the adjacent headland. In these instances, increasingly strong tidal currents flowing through the narrowing inlet prevent spit accretion from sealing off the bay.

Spit Initiation – Spit formation is due to the deflection of sediment that is moving along a beach in the surf and breaker zones into a region of deeper water commonly associated with a bay or tidal inlet. This deflection may be initiated at small protrusions of the shoreline or at major headlands. Similarly, the process may occur at the edge of embayments along an otherwise smooth coastline. The reason the longshore movement of sand forms spits rather than following a pathway along the irregular shoreline is due to a decrease in wave energy in embayments which reduces the capacity of waves to transport sand. Spits exhibit a variety of simple and compound forms, with basic plan-view morphologies described as recurved, cusped, or flying spits (King 1972).

Welded Barriers

These barriers occur along irregular coasts where the supply of sediment is adequate for barrier construction. Depending on their position within a bay, especially if its long axis is normal to the coastline trend, the welded barriers can be described as bayhead, mid-bay, or baymouth barriers. They are common features along rocky coasts such as parts of the West Coast of the United States and along glaciated coasts including New England, Eastern Canada, and Alaska. Most often, this type of barrier is backed by a shallow bay or lagoon; however, freshwater and brackish water marshes also exist. Welded barriers are most common along microtidal coasts where bay areas are diminutive in size. In this type of setting, there is insufficient tidal energy to keep a tidal inlet open. If a welded barrier is breached during a storm, the inlet normally closes quickly because the exchange of tidal waters between the bay and ocean cannot remove the sand dumped into the inlet by waves. Thus, environments with moderate to high wave energy promote the development of welded barriers. The formation of a welded barrier can be a simple process whereby a spit builds across a bay and attaches to the opposite shoreline. However, in many instances, their evolution represents a long-term response to rising sea level. This process is evident along the southern shore of Martha's Vineyard offshore of Cape Cod where low-welded barriers front elongated bays (FitzGerald et al. 1994; Buynevich 2013). Historical evidence shows that some of the barriers were cut by semi-permanent tidal inlets, but these inlets have closed with time as barrier rollover has decreased bay area at a faster rate than rising sea level has inundated the bay shoreline. Some gravel-dominated welded barriers may allow for substantial seawater inflow during flooding tide and thus can be backed by brackish or saltwater wetlands (Forbes and Syvitski 1994).

Barrier Islands

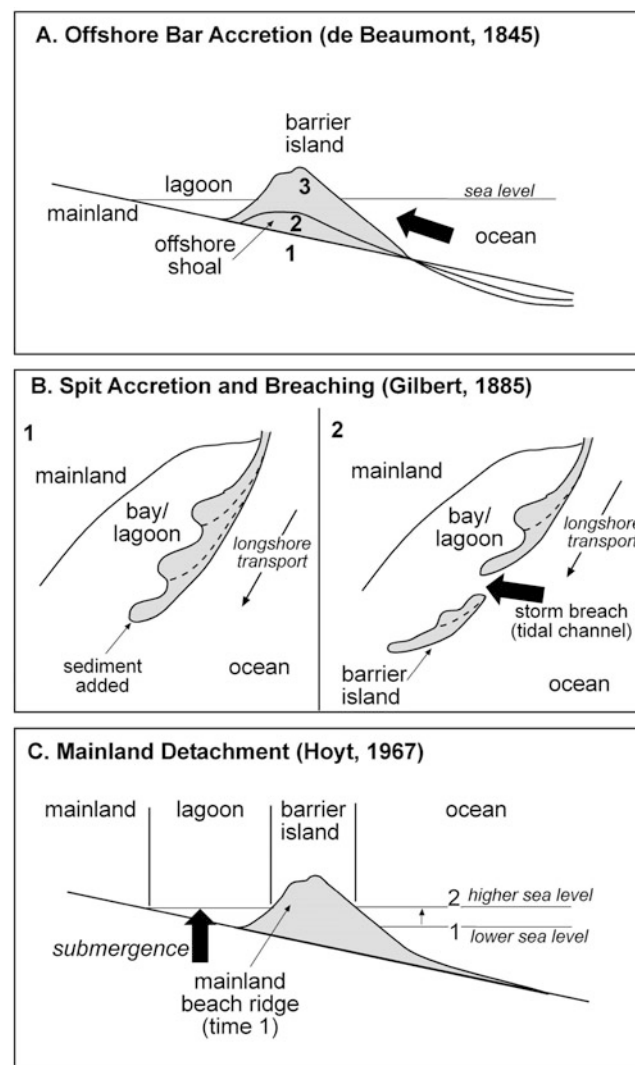
Barrier islands are separated from the mainland or adjacent barriers by inlets at their termini. Backbarrier wetlands or open water bodies separate these landforms from the mainland behind them. Barrier islands may occur in chains that

extend for hundreds of kilometers, interrupted by active (stable or migrating), ephemeral, or man-made inlets. For example, Padre Island along the Texas coast is >200 km long. The length and width of barriers and overall morphology of barrier coasts are related to several parameters including tidal range, wave energy, sediment supply, sea level trends, and basement controls (Hayes 1979; Moslow and Colquhoun 1981; Belknap and Kraft 1985; Roy et al. 1994; McBride et al. 1995, 2013; Hein et al. 2016). The fact they exist along most of the East and Gulf Coasts of the United States with different geologic and oceanographic conditions suggests that barriers can form and be maintained in a variety of settings.

Origin of Barrier Islands – The widespread distribution of barrier islands along the world's coastlines and their occurrence in many different environmental settings have led numerous scientists to speculate on their origin for more than 150 years (Schwartz 1973). The different explanations of barrier island formation can be grouped into three major theories (Fig. 3): (1) *offshore bar theory* (de Beaumont and Johnson), (2) *spit accretion theory* (Gilbert–Fisher), and (3) *submergence theory* (McGee–Hoyt). At present, no one theory explains the development of all barriers and it is sometimes difficult or impossible to prove how an individual barrier originally formed because many lithosomes have been drastically modified since their development due to rising sea level, tidal inlet migration, and related processes.

Recent Barrier Studies – Since the late-1960s, there have been many studies of barrier islands aimed at determining the sedimentary layers making up the barrier and deciphering the manner in which these layers were deposited. This research has been aided by the radiometric dating of organic material contained within the barriers sediments, such as shells, peat, and wood, thereby providing a chronology (timing) of barrier construction. In addition to traditional stratigraphic techniques, ground-penetrating radar (GPR) is also employed to study barriers (Jol et al. 1996; van Heteren et al. 1996, 1998; Hein et al. 2012). Sediment cores are taken in conjunction with the GPR surveys to determine the composition of the sediment layers and the environment in which they were deposited. Additional information concerning former barriers and their associated tidal inlets and lagoons has been gathered from the inner continental shelf using high-resolution shallow seismic surveys, a technology similar in principle to GPR. These advancements have provided new insights about the formation of coastal barrier systems.

In considering the formation of barrier islands, it is important to recognize that almost all the world's barriers are less than 6,500–7,000 years old and most are younger than 4,000 years old. Most barriers formed in a regime of rising sea level but during a time when the rate of rise began to slow. Sedimentological data from the inner continental shelves off the East Coast of the United States, in the North Sea, and in



Barrier Island Landforms, Fig. 3 Theories of barrier island formation. (Modified from Penland et al. 1988)

southeast Australia (Roy et al. 1994; McBride et al. 2013) suggest that barriers once existed offshore and have migrated (retrograded) to their present positions.

Hydrographic Regime and Barrier Coast Morphology

Many factors determine the location and size of individual features along a coast, such as the slope of the land dictating the size of a lagoon or a former stream valley controlling the position of a tidal inlet. The overall distribution of barriers, tidal inlets, and various backbarrier environments is primarily related to the relative contribution of wave and tidal energy. In a simplification of their respective roles, waves are responsible for the transport of sediment along the shore, which tends to elongate barriers and onshore-offshore during storms.

The rise and fall of the tides cause a filling and emptying of backbarrier areas. Tidal inlets through which this exchange of water occurs are the sites of strong tidal flow. Tidal currents transport sand in an onshore and offshore direction.

Hydrographic Regime and Types of Barrier Coast

In a conceptual model conceived by Miles Hayes (1979), coastlines are separated into three classes based on the wave height and tidal range of the region. The three major divisions include wave-dominated, mixed energy, and tide-dominated settings. Barriers are found almost exclusively in the wave dominated and mixed-energy environments. It should be noted that it is not the absolute value of either wave height or tidal range that is significant, but rather the ratio of the two parameters that dictate the presence and distribution of barriers. Because barriers are predominantly wave-built accumulations of sediment, barriers will not form in regions where wave energy is insufficient. The importance of wave energy is illustrated along the West Coast of the United States where barriers exist despite spring tidal ranges approaching 4 m. Here, an abundant sand supply and strong wave energy overcome the effects of tides to produce very long barriers. The key hydrographic types of barrier coasts are described below.

Wave-Dominated Coast – These coasts are largely influenced by wave-generated longshore sediment transport, with tides playing a secondary role. They are characterized by long linear barrier islands and thus are separated by few tidal inlets. The backbarriers are composed of mostly open-water lagoons or bays. Marshes fringe the edge of the mainland and the backside of the barriers, commonly on former washover deposits. Large sand shoals are typically found on the landward side of the tidal inlets. Sand shoals on the seaward side (flood-tidal deltas) are diminutive in size. Coastlines fitting into this class include the coasts of Texas, Panhandle of Florida, Outer Banks of North Carolina, Maryland, northern New Jersey, barrier coast along the Nile River delta, southeast Iceland, parts of Australia and southern Brazil, southeast Baltic Sea, and the northern Black Sea, among others.

Mixed-Energy Coast – In this model, both wave and tidal processes are equally important in shaping coastal morphology. Barriers on these coasts tend to be short and stubby and tidal inlets are more numerous than along wave-dominated coasts, reflecting the greater role of the tides compared to wave-dominated settings. The backbarriers of these regions are mostly filled with sediment and covered by expansive marshes incised by tidal creeks, which convey the flooding and ebbing currents. Open-water areas commonly increase in extent near the inlets. In at least two locations of the world, the backbarriers of this coastal type (East Friesian Islands along Germany; Copper River Delta barriers in Alaska; Davis 1994) consist of extensive unvegetated tidal flats. It is likely that the change from open water lagoons of wave-dominated coasts to the intertidal environments of mixed-energy coast is the result

of greater sediment volumes being transported into the backbarrier by tidal currents. In addition, an increase in tidal range produces larger intertidal areas that promote the saltmarsh development and stabilization of sediment. Mixed-energy coasts include barrier coasts of northern New England, southern New Jersey, Virginia, South Carolina, Georgia, Copper River delta barrier system in Alaska, Friesian Islands in the North Sea, parts of Colombia, west Madagascar, and ice-free segments of the Siberian coast.

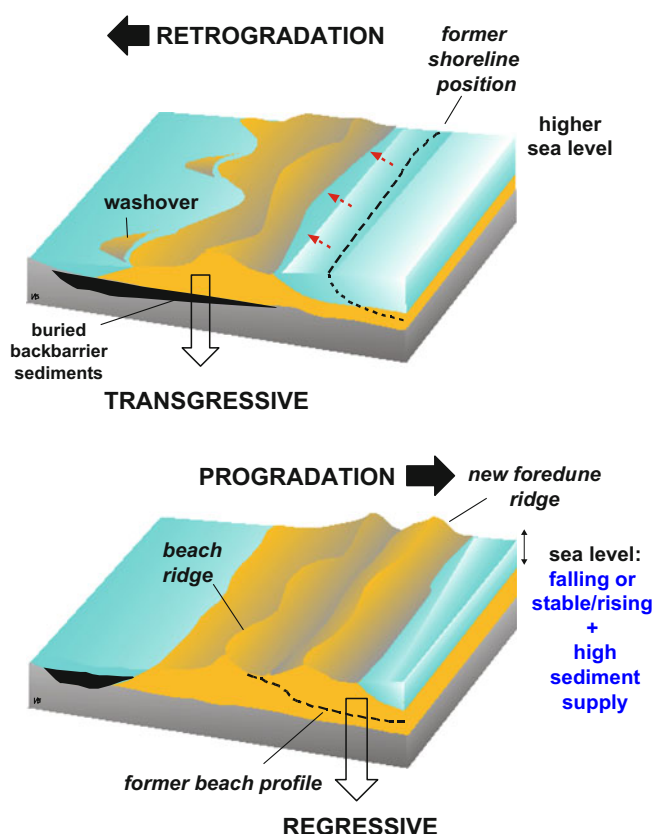
Tide-Dominated Coasts – Many tide-dominated coasts coincide with very large funnel-shaped embayments (>100 km across). This coastline configuration enhances the tidal wave, which produces large tidal ranges, strong tidal currents, and a sedimentation regime that is dominated by onshore-offshore transport. Due to the low wave energy along such coastlines, barriers and tidal inlets are absent. Rivers commonly discharge sediment at the heads of these embayments. The coarse-grained sediment is transported offshore and deposited on large subtidal sand ridges that parallel the tidal flow. The fine-grained sediment tends to accumulate onshore forming expansive tidal flats. Landward of the tidal flats are wide marshes. In equatorial regions, marshes are replaced by mangroves. The Bay of Fundy, Nova Scotia; Gulf of Cambay, India; Head of the Bay of Bengal, Bangladesh; mouth of the Amazon River (although high wave energy results in small welded barriers), and Bristol Bay, Alaska, are examples of tide-dominated coasts.

Prograding, Retrograding, and Aggrading Barriers

The overall form of barriers, their stability, and future erosional, depositional, or migration trends (morphodynamics) are related to the supply of sediment, the rate of sea-level rise, storm cycles, and the topography of the mainland (Fig. 4; McBride et al. 1995). When a barrier builds in a seaward direction, it is said to prograde and is called a *prograding* or *prograded* barrier. The opposite of this condition occurs when the entire landform migrates landward or retreats, called a *retrograding* barrier. A vertical accretion by waves and wind, which is maintained in a regime of sea-level rise, is labeled an *aggrading* barrier. These first two morphodynamic states and types are most common and are described below.

Prograding Barriers

These landforms form in a regime of abundant sand supply during a period of stable or slow rise in sea level in a process similar to mainland-attached strandplains or delta plains. The bulk of sediment may come from alongshore or offshore sources. These conditions were met along much of the East Coast of the United States and many other regions of the world about 5,000–4,000 years ago when the rate of



Barrier Island Landforms, Fig. 4 Styles of barrier evolution and generalized stratigraphy: (a) retrogradation producing transgressive stratigraphy; (b) progradation resulting in regressive stratigraphy

sea-level rise decreased and sediment was contributed to the shore from *eroding headlands* (examples: Provinceland Spit in northern Cape Cod, Massachusetts; Lawrencetown barrier along the Northeast Shore of Nova Scotia; FitzGerald and van Heteren 1999; FitzGerald and Rosen 1987), from the *inner continental shelf* (examples: barrier system along Bogue Banks, North Carolina; Algarve barrier chain, in southern Portugal; Tuncurry barrier in southeast Australia; East Friesian Islands along the German North Sea coast), and *directly from rivers* (examples: most barriers systems in northern New England; barrier chain situated north of the Columbia River along southern coast of Washington).

Prograding barriers commonly build in spurts when the supply of sand is plentiful. Increases in the rate of sediment contribution to the downdrift coast can occur during storms when unconsolidated cliffs are eroded releasing large quantities of sand to the littoral system or during floods when rivers transport high sediment loads to the coast. In other instances, sediment is moved onshore from the inner shelf during long-term accretionary wave conditions. These processes cause beaches to build in a seaward direction, widening the berm and separating the foredune ridge from the ocean by an enlarging expanse of sand.

A common end product of shoreline progradation is the development of *beach ridges*. In contrast to true beach ridges build exclusively by waves, most are compound beach-dune landforms that may become vegetated first by grasses, then shrubbery, and finally by trees. This change in the maturity of the vegetative cover occurs as the ridges are displaced progressively farther from the shoreline and become more protected from salt spray and the effects of storms. Each beach ridge marks a former shoreline position and their overall pattern indicates how the barrier grew and evolved through time. The interval of time between the formation of successive beach ridges or ridge sets may represent decades or centuries. Beach ridges may develop as storm-eroded scarps, wrack-line accumulations, onshore bar migration, or a combination of the above processes (Tanner 1995). A barrier composed of beach ridges is easily identified as one having evolved through progradation; however, this morphology does not necessarily indicate that the barrier is still prograding. Conversely, sedimentation conditions along the barrier may have changed such that it entered an erosional phase, with its width and/or height precluding landward migration (retrogradation). Prograded barriers are characterized by *regressive stratigraphy*, with beach sands overlying nearshore sediments in a generally coarsening-upward sequence.

Retrograding Barriers

This type of barrier forms when the supply of sand is inadequate to keep pace with relative sea-level rise and/or with sediment losses. Stated in terms of a sediment budget, a barrier undergoes *retrogradation* when the amount of sand contributed to the barrier is less than the volume transported away from the barrier. The sand may be lost offshore during storms, moved along shore in the littoral system, or transported across the barrier by overwash. The end result is erosion to the front of the barrier causing a decrease in width of the beach and ultimately a destruction of the foredune ridge. Eventually, a narrowing and lowering of the barrier profile produce a retreat of the barrier across the adjacent bay, lagoon, or marsh system. This landward migration of the barrier, termed *barrier rollover*, is accomplished primarily during storms by *overwash* is a cannibalistic process whereby storm waves transport sand from the beach through the dunes, depositing it along the landward margin of the barrier. Where shallow backbarrier platform results in low accommodation space; the storm-transported and wind-modified barrier lithosome is largely preserved during the retreat and rollover.

Retrograding barriers are identified by their overall narrow width, single or nonexistent foredune ridge, and overwash aprons or fans (washovers). Because these barriers have been migrating over various types of backbarrier settings, including lagoons and marshes, the sedimentary components comprising these environments are buried beneath beach sands.

This produces *transgressive stratigraphy* and the coarsening-upward system similar to a prograded system, but with fine-grained basal sequences of backbarrier rather than nearshore genesis. With barrier rollover, these facies become exposed along the seaward side, usually in the intertidal zone. This explains the appearance of stiff muds and the remnants of peat deposits along the lower beach of many retrograding barriers. In some instances, tree stumps may extend through the sands of the intertidal zone, indicating that the barrier has migrated onshore a distance great enough such that trees once growing on the mainland have resurfaced as stumps along the ocean coast. This exposure of backbarrier sediments must be considered in predictive models of barrier coast evolution, as reworking of fine-grained materials will not contribute to rebuilding the barrier lithosome.

Cross-References

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- [Antarctica, Coastal Ecology and Geomorphology](#)
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- [Asia, Eastern, Coastal Geomorphology](#)
- [Asia, Middle East, Coastal Ecology and Geomorphology](#)
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BARS

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Sedimentary ridges, both symmetric and asymmetric, and generally larger than bedforms that characterize the upper shoreface of coastal zones dominated by waves are called *wave-formed bars*. They were recognized as early as 1845 on the marine coasts of Europe (Elie de Beaumont), by 1851 in the Great Lakes of North America (Desor), and subsequently on marine and lacustrine coasts worldwide (see Schwartz, 1982, pp. 135–139). However, confusion still surrounds this term because of its use for ridges with a wide range of size, morphology, location, and orientation relative to the shoreline. Also, the term *bar* has been used in a variety of environments, from subaerial to those dominated by tidal currents or river currents. Furthermore, the present understanding of the origin (s) and dynamics of *wave-formed bars* is still incomplete.

Shepard (1950) called shore-parallel ridges and troughs *longshore bars and troughs*, equating them with the terms *ball*

and *low* of Evans (1940), and associated them with plunging breakers. He emphasized the *seasonality* of such bars on the west coast of the United States, and subsequently terms such as *winter* and *summer*, *storm* and *normal*, and *storm* and *swell* have been applied to denote the presence or absence of bars. Although a correlation between profile form and storm waves or season may exist in some localities (e.g., Inman et al. 1993), it is not universal. Both *barred* and *non-barred* profiles occur at times in some areas, while in others only one profile type may persist throughout the year. There is usually a distinct *relaxation time* between the forcing conditions and bar adjustment; thus in the short-term, bars are generally in a transient state. In the longer term (years to decades), wave-formed bars represent the *equilibrium morphology* for many coastal environments.

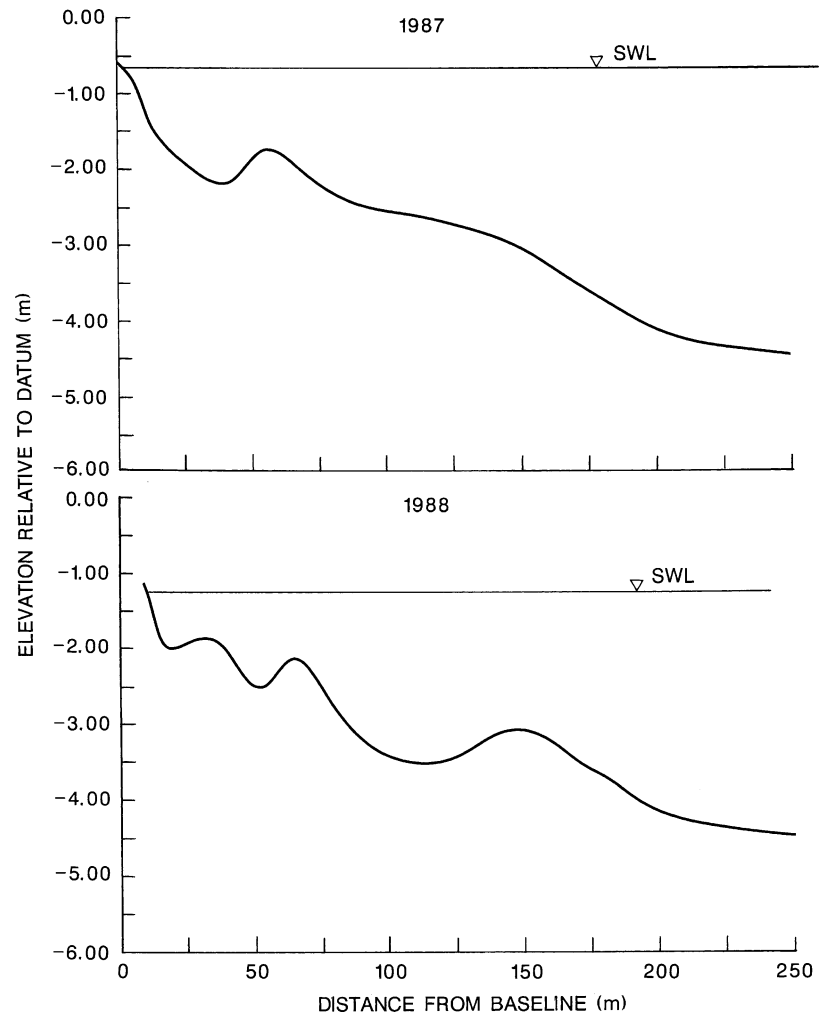
Bar Morphology

Wave-formed bars are most clearly identified as near-symmetrical or asymmetrical undulations in the upper shoreface profile (Fig. 1). They occur intertidally and subtidally, and may range in number from one to more than thirty, this number often varying through time. Short and Aagaard (1993) introduced a bar parameter, $B^* = x_s/gT^2 \tan \beta$, to identify the number of bars on a linear sloping shoreface ($\tan \beta$) terminating at a constant depth at a distance offshore, x_s . When $B^* < 20$, the profile is non-barred, for $B^* = 20$ –50, 1 bar occurs; for $B^* = 50$ –100, 2 bars, for $B^* = 100$ –400, 3 bars; and for $B^* > 400$ there are 4 bars. Crest heights above the adjacent trough can range from less than a decimeter (Carter 1978) to more than 4.75 m (Greenwood and Mittler 1979). In plan view, they form continuous or compartmentalized, linear, sinuous, or crescentic patterns, and range from shore-parallel to shore-normal in orientation, often producing periodic or *rhythmic* topography both alongshore and cross-shore. The morphometry of bars has been studied by Greenwood and Davidson-Arnott (1975), Hands (1976), and Ruessink et al. (2000) in order to define the equilibrium form and dynamics induced by a specific set of environmental constraints.

Bar Classification

A *universal* classification of *wave-formed bars* does not yet exist, and indeed it may never be possible to define perfectly mutually exclusive classes. A simple descriptive classification based on morphology and the associated environmental constraints is illustrated in Table 1 (Greenwood and Davidson-Arnott 1979). The group names are those in common use, and the definitive paper describing each type is cited. Other classifications are based on the concept that bars are part of a temporal sequence of beach profile evolution

BARS, Fig. 1 Typical barred profiles from a sandy nearshore environment in the Canadian Great Lakes. The profiles were surveyed in successive years along the same transect. Note the differing number and position of the wave-formed bars at the same location, even though the mean beach slope is the same



and that they are scaled to that of the controlling wave process. The morphological sequence is controlled by incident wave energy (high and low frequency) and was identified either through aerial photographs or more recently through video-imagery (e.g., Short 1979; Lippmann and Holman 1990; see Fig. 2). Many coastal environments do not experience such sequential behavior.

Ridge and runnel topography (*Type I*) is found on low-angle, macro-to meso-tidal foreshore slopes dominated by surf action and foreshore drainage during the tidal cycle. Although low in amplitude these bars are usually stable in form and position or migrate only slowly. In contrast, the *cusp- or bar-type sand waves* (*Type II*) are extremely dynamic, often destroyed during storms and regenerated as the storm wanes and smaller amplitude, longer period waves propagate shoreward. These bars result from surf bores and swash action near the toe of the swash slope (an alternative name is *swash bar*). Furthermore, they may develop from *Type VI* bars as they migrate relatively rapidly both alongshore and onshore, and in the latter case may *weld* to the foreshore (Davis et al. 1972; Aagaard et al. 1998). Note that there is confusion with

respect to the term *ridge and runnel* as used in northwest Europe and North America (Orford and Wright 1978; Orme and Orme 1988). Here, the term *ridge and runnel* is restricted to its initial definition by King and Williams (1949); the forms described by Hayes and subsequent workers (Hayes and Boothroyd 1969) are classified here as *Type II* bars.

Type III multiple parallel bars (e.g., Nilsson 1973; Exon 1975) and *Type IV transverse bars* (e.g., Niedoroda and Tanner 1970; Carter 1978; Dolan and Dean 1985) tend to be limited to low-angle shorefaces and small to moderate wave heights, coupled with limited water level shifts. However, they have been identified on more energetic shorelines (Konicki and Holman 2000).

The number of bars increases with decreasing beach slope (Davidson-Arnott 1988). The height and spacing of the multiple bars increases in the offshore direction, and bar form is near symmetrical in contrast to the *Type II* group. Transverse bars run normal or obliquely to the shoreline and can range in length from 3 m up to 4 km, with heights from less than 0.05 m up to 2 m and alongshore spacing of the order of 10^0 – 10^2 m (Carter 1978;

BARS, Table 1 Bar morphologies and environmental constraints

Class	Name	Definitive description	Morphology			Environmental constraints				
			Size (m)	Planform	Profile	Number seaward	Location	Wave energy	Breaking processes	Tidal range
Type I	Ridge and runnel	Kang and Williams (1949)	$h \sim 0.2-1.5$ $l \sim 10^3$	Straight, shore parallel	Asymmetric landward	1-4	Intertidal	L-M	Surf-swash, beach drainage	Ma-Me
Type II	Cusp-or bar-type sand wave	Sonu (1973)	$h \sim 0.2-1.5$ $l \sim 10^2$	Straight to spit shaped, shore parallel	Asymmetric landward	1-2	Intertidal and low tide terrace	M	Breakers, surf-swash	Me-Mi
Type III	Multiple parallel	Zenkovich (1967)	$h \sim 0.2-0.75$ $l \sim 10^3$	Straight to sinuous, shore parallel	Near symmetric	4-30 or more	Nearshore & intertidal	L-M	Spilling	Mi-Me
Type IV	Transverse	Niederoda and Tanner (1970)	$h \sim 0.2-0.75$ $l \sim 10^2$	Straight, shore normal	Symmetric and asymmetric landward	1	Nearshore and intertidal	L	Surf-swash.	Mi spilling
Type V	Nearshore I	Shepard (1950)	$h \sim 0.15-1.0$ $l \sim 10^2$ (?)	Straight, shore parallel	Asymmetric landward	1-2	Nearshore	M-H	Plunging	Mi-Me
Type VI	Nearshore II	Evans (1940)	$h \sim 0.25-3.0$ $l \sim 10^3$	Straight, sinuous to crescentic, shore parallel	Asymmetric landward	1-4	Nearshore	M	Spilling	Mi

Note: h = bar height; l = bar length; Ma = macro; Me = meso; Mi = micro; L = low; M = moderate; H = high

Source: Modified from Greenwood and Davidson-Arnott (1979)

WRIGHT & SHORT (1984)		LIPPMANN & HOLMAN (1990)
Beach State 6 <i>Non-barred, dissipative beach</i>	=====	Type H <i>Non-barred dissipative; Infragravity wave scaled surf zone</i>
		Type G <i>2-D, infragravity wave scaled bar</i>
Beach State 5 <i>Longshore Bar & Trough</i>		Type F <i>3-D, non-rhythmic, infragravity wave scaled bar</i>
Beach State 4 <i>Rhythmic Bar & Beach</i>	=====	Type E <i>Offshore rhythmic, infragravity wave scaled bar</i>
		Type D <i>Attached rhythmic, infragravity wave scaled bar</i>
Beach State 3 <i>Transverse Bar & Rip</i>		Type C <i>Attached non-rhythmic, infragravity wave scaled bar</i>
Beach State 2 <i>Swash Bar-Shore Terrace</i>	=====	Type B <i>Incident wave scaled, 2-D bar</i>
Beach State 1 <i>Non-barred reflective beach</i>	=====	Type A <i>Non-barred, reflective, incident wave scale</i>

BARS, Fig. 2 Classification and scaling of sequential upper shoreface morphologies. The equivalence between the contrasting sequences of Wright and Short and Lippmann and Holman is indicated. (Modified after Lippmann and Holman 1990)

Gelfenbaum and Brooks 1997). The larger forms may migrate alongshore at rates up to 8 m a^{-1} . Usually, transverse bars are anchored to the shoreline (indeed they appear as an extension of a shoreline protuberance), but Konicki and Holman (2000) recorded the unusual case of transverse bars running offshore from a *Type VI* bar.

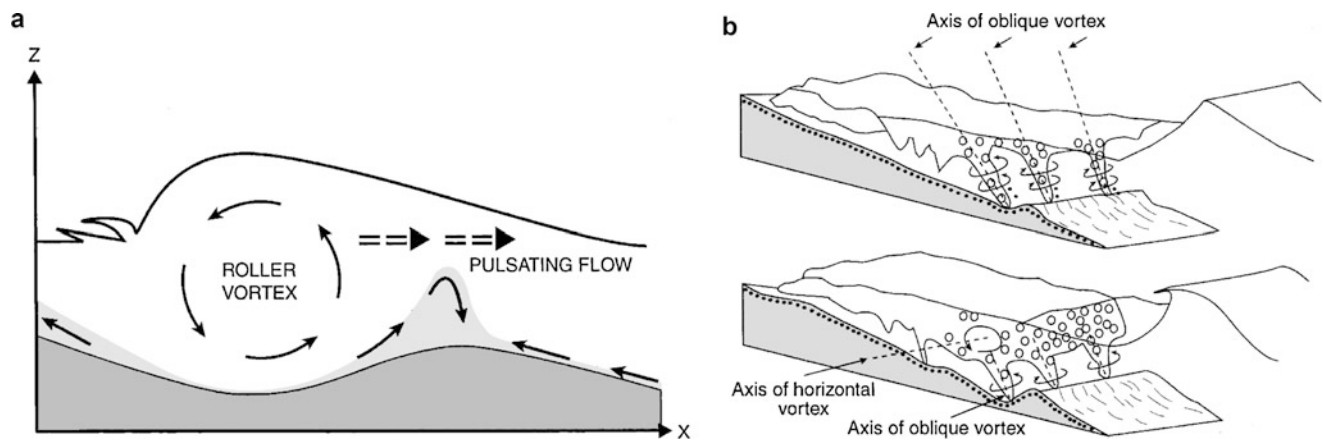
The division of *nearshore bars* into two groups is based upon size, stability, and the controlling waveform. *Type V bars* are associated with large plunging breakers, which produce narrow, low amplitude ridges on relatively steep slopes: they lack a well-defined asymmetry and are essentially unstable modifications of non-barred nearshore profiles. *Type VI bars*, in contrast, are relatively large configurations formed seaward of the low water level. Where there is more than one bar, the distance offshore, depth-of-water over the crest, and bar height all usually increase offshore in a regular manner, although in some cases the height decreases after some offshore distance (Lippmann et al. 1993; Ruessink and Kroon 1994). The volume of sediment in each bar form usually increases consistently offshore. *Type VI bars* may be three-dimensional, sinuous-to-crescentic, and the alongshore length scales may range from 10^2 to 10^3 m (Greenwood and Davidson-Arnott 1975; Bowman and Goldsmith 1983). Where more than one bar is rhythmic, the alongshore wavelength decreases shoreward.

Bar Genesis

The boundary conditions necessary for bar formation depend upon the longer term evolution of the coast, which dictates the nature of the bed materials (grain mineralogy, size, sorting, etc.), the bathymetric setting (slope, exposure, etc.), and the geographic location (wave climate, tidal regime, etc.). Local forcing conditions for bars have been studied both theoretically and empirically, and by experiments in the laboratory and the field (see van Rijn 1998). In general, barred profiles are associated with large values of both wave steepness and wave height-to-grain size ratios, and are associated with the final stages of shoaling and dissipation of wave energy through breaking, and the complex hydrodynamics, which accompany these processes (Wright et al. 1979). Furthermore, the size of wave-formed bars induces a very strong feedback to the shoaling and breaking process. Although cause and effect are far from clear, it is evident that equilibrium bar profiles can exist only where the time-averaged sediment transport (suspended and bedload) is zero everywhere on that profile.

A large number of specific hypotheses have been proposed for bar formation over the last 50 years and all involve mechanisms for convergence of sediment transport; these hypotheses were primarily related to *Type V and VI* bars and fall into three major groups:

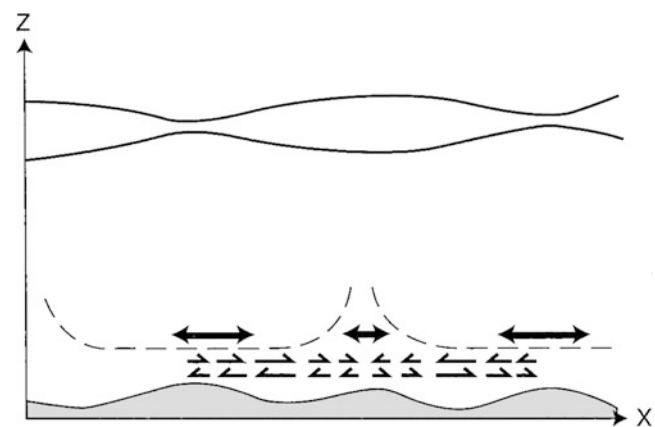
1. *break point hypotheses* relate bars directly to wave breaking and result from: (i) a seaward transport of sediment entrained by roller or helical vortices under plunging or spilling breakers, respectively (Miller 1976; Zhang 1994; Fig. 3); (ii) convergence of sediment at the breakpoint through onshore transport associated with increasing asymmetry and skewness of the high-frequency incident waves and offshore transport through set up induced undertow (Dally and Dean 1984; Dally 1987; Thornton et al. 1996). However, Sallenger and Howd (1989) concluded that bars are not necessarily coupled to the breakpoint, but can grow and migrate, while within the inner surf zone, landward to the point of initial breaking.
2. *Infragravity wave hypotheses* propose that low frequency waves generated within the surf zone (surf beat) or offshore and reflected produce a convergent pattern of drift velocities, which interact with the large incident short wave oscillatory velocities to induce a range of bar forms from two-to three-dimensional crescentic forms (e.g., Bowen and Inman 1971; Short 1975; Bowen 1980; Holman and Bowen 1982; Bowen and Huntley 1984). These waves can be standing or progressive and can be produced in a number of different ways as a result of energy dissipation during breaking and are frequently related to amplitude modulation of the incident wave field (*groupiness*; Roelvink and Broker 1993; Ruessink 1998). Alternating scour and deposition by



BARS, Fig. 3 Bar formation by breaking waves: (a) trough scouring by a roller vortex under plunging breakers and offshore sediment transport converging with sediment driven onshore by shoaling waves (modified

after Miller 1976); (b) trough scouring by oblique vortices generated under spilling breakers (modified after Zhang 1994)

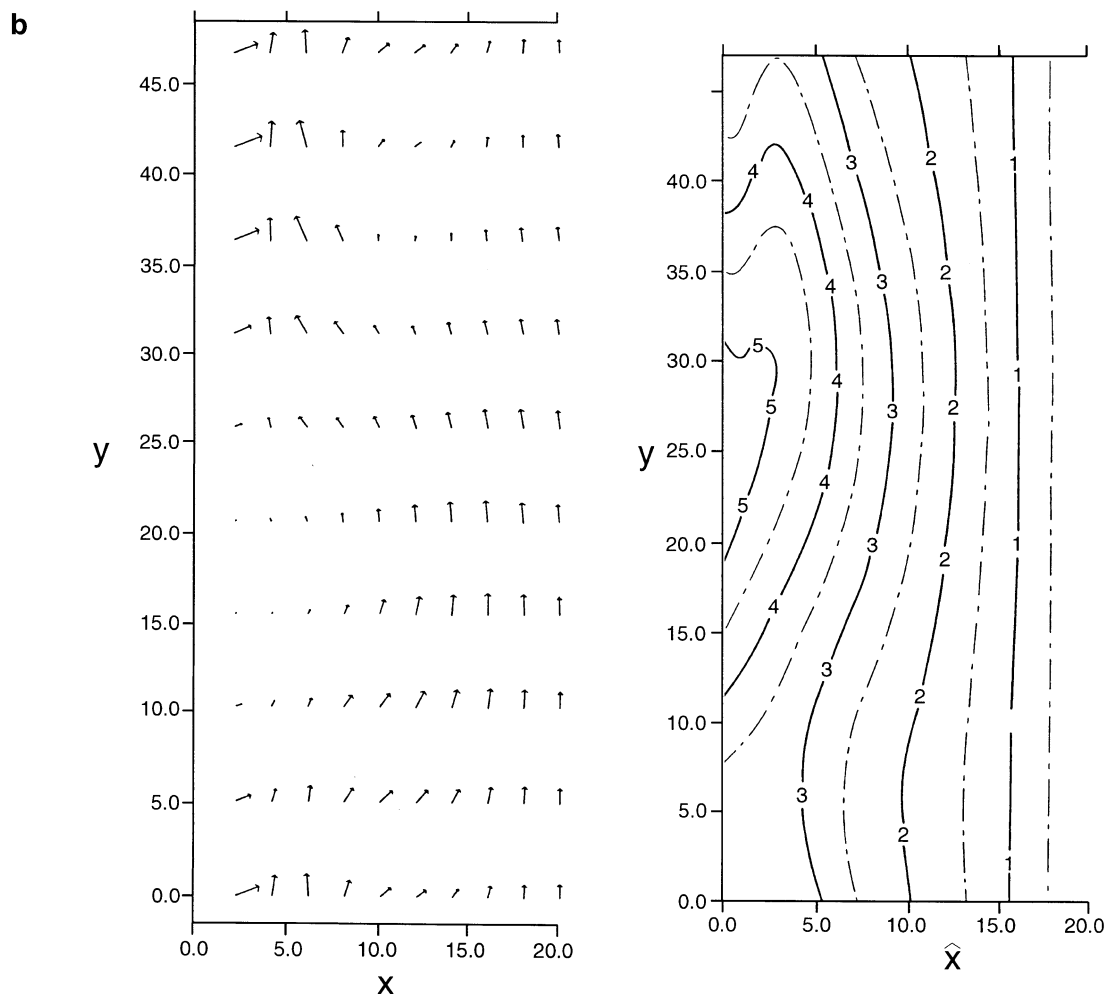
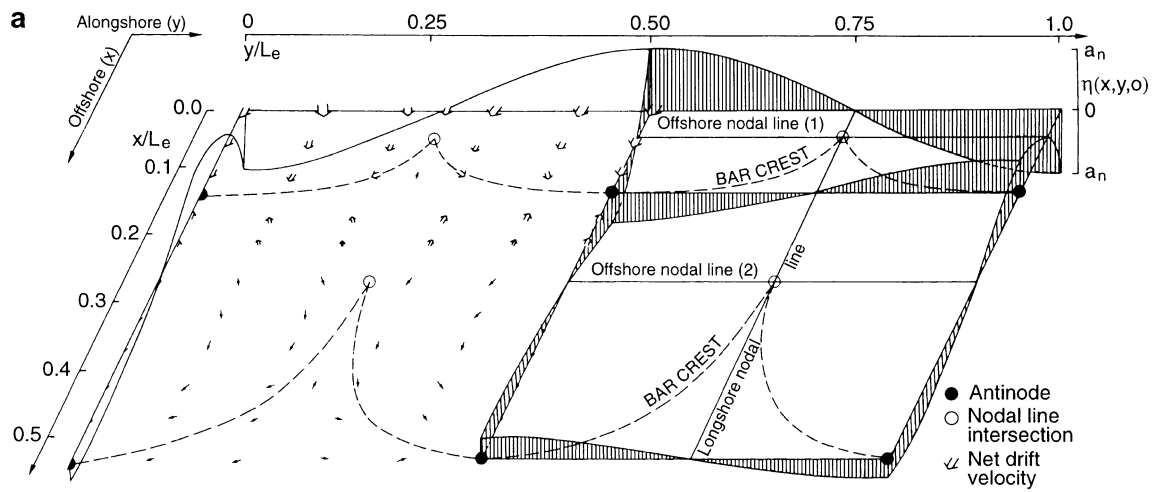
mass transport velocities in the bottom boundary layer generated by standing waves, resulting from the interaction of reflected and incident waves, was shown to occur in the laboratory by Carter et al. (1973; Fig. 4). The boundary layer was actually segregated. At the bed, drift velocities converge at nodes, while at some distance above the bed drift velocities converge at antinodes. Under large waves when bars are most active, suspension transport is dominant and therefore sediment will converge and bars will form at the antinodal position of standing waves (e.g., Bowen 1980). Reflection of waves in the infragravity range was clearly demonstrated by Suhayda (1974) and shown to relate to bar forms by Short (1975) and Katoh (1984). Sediment moves to null positions in the drift velocity field of low-frequency standing (Fig. 5a) or progressive edge waves (Fig. 5b), which are periodic both alongshore and offshore. Recent field measurements have clearly shown the importance of group-bound long waves to suspended sediment transport in barred surf zones (e.g., Osborne and Greenwood 1992), but isolating the drift velocities associated with these secondary waves is difficult. This second-order drift velocity hypothesis requires one dominant wave frequency, which is not common (see Bauer and Greenwood 1990 for an exception). However, there are a number of suggestions to overcome this inadequacy of the edge wave hypothesis. Aagaard (1990) has argued for the excitation of cutoff mode edge waves (limited by the beach slope) and selection of the dominant mode as that mode which is closest to the wave group period. A phase coupling between the primary orbital motion of a partially standing long wave and groupy short waves was also proposed by O'Hare (1994) to avoid this requirement of narrow bandedness in the infragravity spectrum. Other mechanisms producing a limited number of edge wave frequencies and modes are topographic control



BARS, Fig. 4 Bar formation as a result of mass transport in the boundary layer of a strongly reflected incident wave. The surface wave envelope is shown as well as the circulation within the bottom boundary layer. Bed load will converge at nodes of the surface elevation and suspended load at antinodes. Note: the boundary layer flow is indicated by single-headed arrows; the mean flow is indicated by double-headed arrows

(Kirby et al. 1981; Bryan and Bowen 1996) and interaction of edge waves and the longshore current (Howd et al. 1992). O'Hare and Huntley (1994) propose a leaky wave origin for an inner surf zone bar, which is relatively insensitive to the group period, incident wave height, and the width of the infragravity spectrum.

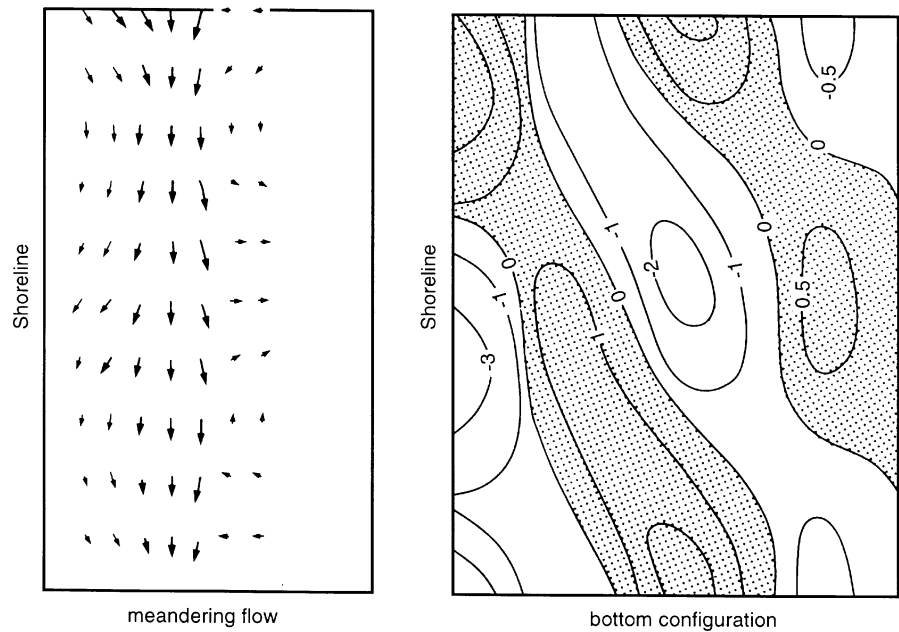
self-organization hypotheses propose that processes associated with the complex, nonlinear feedback between the sand bed and the hydrodynamics give rise to a range of topographic forms. For example, alongshore and offshore sediment movement was proposed under meandering or cellular nearshore circulations produced by (i) instability of longshore flows (Fig. 6; Barcilon and Lau 1973; Hino 1974; Falques 1991; Damgaard Christiansen et al. 1994);



BARS, Fig. 5 Bar formation by infragravity waves: (a) net drift velocities associated with standing edge waves and the creation of crescentic bars (modified after van Beek 1974). (b) dimensionless drift velocities and equilibrium nearshore bathymetry associated with the propagation

of two edge wave modes (1 and 2) of the same frequency in the same direction (modified after Holman and Bowen 1982). Note: y represents the alongshore direction and x the across-shore direction

BARS, Fig. 6 Bar formation due to hydrodynamic instability between longshore currents and the sand bed (modified after Hino 1974). Note the meandering nature of the longshore flow and the sinuous bar topography that is produced



(ii) coupling between morphodynamic instability and mean flows (Deigaard et al. 1999; Vittorio et al. 1999; Falques et al. 2000); and (iii) Bragg scattering from periodic topography (Heathershaw and Davies 1985; O'Hare and Davies 1993; Rey et al. 1995; Yu and Mei 2000). These mechanisms cannot produce bars directly, but require some initial perturbation of the profile. However, it has been shown that some bar characteristics are not well predicted by these models (e.g., the crossshore/ alongshore spacing—see Konicki and Holman 2000). The nonlinear action between shoaling waves and the bed (Boczar-Karakiewicz and Davidson-Arnott 1987) was also proposed as a mechanism for generating periodic patterns of sediment transport which matched the spacing and general shape of multi-barred shorelines.

The horizontal roller vortex mechanism is most applicable to single *Type V* bars of the US west coast, and justifies the early correlation of bar formation with wave steepness. Multi-barred profiles reflect either multiple breakpoints (Dally 1980; Davidson-Arnott 1981) or bar formation by distinct differences in wave energy; for example, an outer bar may be produced under storm waves and an inner bar by less energetic conditions (King and Williams 1949). Water level shifts and coincident shifts in breaker location could also produce a multiple barred system. Oblique, helical vortices were produced under spilling rather than plunging breakers in the laboratory and could account for both single and multiple barred profiles (Zhang 1994). However, the mass transport velocities under reflected standing waves would perhaps best explain the formation of *Type III* multiple parallel bars; simple reflection of the incident waves could not be the cause, since

the length scales of the bars would require much longer periods. The theoretical convergence patterns of drift velocities under standing edge waves provide strong support for their role in forming crescentic *Type VI* bars. Progressive edge waves may be responsible for linear bars of the same group (Huntley 1980). Further, the edge wave periods necessary to produce the length scales found in nature is of the same order as the well-known *surf beat*. However, the generation of these trapped modes of oscillation still remains ill-defined, even though field observations of low-frequency peaks in the near-shore energy spectrum have been made on barred coasts and related to the presence of edge waves (Huntley 1980; Bauer and Greenwood 1990).

Barred topography has long been associated with the occurrence of cellular nearshore circulations (Shepard et al. 1941), and Hino (1974) proposed that an instability of the fluid sediment interface would generate variations in sediment transport resulting in sinuous or crescentic undulations of the surf-zone bed (Fig. 5). Certainly the role of ripcell circulation in bar dynamics has been well documented for *bar-* and *cusp-type sand waves* (Bowen and Inman 1969; Davis and Fox 1972; Sonu 1973; Greenwood and Davidson-Arnott 1975; Wright and Short 1984), for *transverse bars* (Niedoroda 1973), and for *Type VI* bars, both crescentic and straight (Greenwood and Davidson-Arnott 1979). However, it is also possible that the regularity in nearshore circulations is in fact controlled by the presence of edge waves (Holman and Bowen 1982). Whichever mechanism initiates bars, there will be feedback between the topography and the hydrodynamics, perhaps giving rise to some “hybrid” model of formation (Holman and Sallenger 1993).

Bar Morphodynamics

In general, the smaller the *wave-formed bar* the more dynamic it is, as there is less sediment involved in morphological changes (Sunamura and Takeda 1984). However, there is considerable variability in morphodynamic behavior, depending upon bar type, the general environmental constraints, and indeed the antecedent state of the bar (i.e., whether or not it is close to its equilibrium position). Bars also tend to migrate at lower rates as the tidal range increases, since at some stage the bars are being exposed subaerially and remain static at this time. Bar dynamics have generally been related to behavior under specific storm events. However, the magnitude, frequency, and sequencing (*chronology*) of such events may be important in the nearshore, which as a non-linear dynamical system, is extremely sensitive to feedback processes (see Moller and Southgate 1997; Southgate and Moller 2000; see Elgar 2001 for an alternative view). There now exist at least two long time series of morphological change: (1) thirty years of annual profiling along 100 km of the Dutch coast (Ruessink and Kroon 1994); (2) sixteen years of bathymetry recorded at Duck, NC (Plant et al. 1999). Extensive measurements of the cross-shore location, and alongshore bar shape, are now being made successfully on a near continual basis at a number of locations worldwide using video-imagery (e.g., Lippmann and Holman 1990; van Enckvort and Ruessink 2001).

Type I bars are relatively stable in general, although landward migration rates of ~ 10 m per month have been recorded. Under low energy conditions the ridges have been observed to be: (1) destroyed by storms and regenerated in the post-storm period (Mulrennan 1992); and (2) formed by storms (Hale and McCann 1982). *Type II* bars have been shown to migrate at relatively rapid rates, both onshore and alongshore, and *Type III* bars migrate also at a relatively rapid rate. *Type II, IV, and VI* bars have been shown to occur as part of a temporal sequence of beach evolution by Wright et al. (1979), Wright and Short (1984), Sunamura (1988), and Lippmann and Holman (1990). This sequence ranges from fully dissipative (barred profile) to fully reflective (non-barred profile) wave conditions, and therefore, is related to the surf similarity parameter ($\varepsilon = a_b \omega^2 / g \tan^2 \beta$; where a_b = breaker amplitude, ω = incident radian wave frequency, g = the gravitational constant, β = beach slope). In the Australian Model, the two-dimensional shore-parallel longshore bar and trough occurs at the fully dissipative beach stage, the rhythmic bar and beach at an intermediate stage, and the non-barred profile occurs at the fully reflective stage. In regions where a more limited range of waves exist, the beach may simply change between one or two stages, and where the environmental constraints are more restrictive still, then the bars may assume only one characteristic morphology. Further refinement of the stage model used the *Dean Parameter* ($\Omega = H_b / \omega_s T$; where

H_b = breaker height in meters; ω_s = sediment fall velocity in meters per second; T = wave period in seconds). Barred profiles occurred when $\Omega > 0.85$ and non-barred profiles occurred when $\Omega < 0.85$ (Wright et al. 1985). Sunamura (1988) used the dimensionless parameter $K^* = H_b^2 / g T^2 d$, where g is the gravitational constant and d is the grain size, to classify sequences dependent upon erosional or accretional beach stages. Erosion is characterized by $K^* \geq 20$ and is associated with offshore bar migration, slope decreases, and a dissipative state; while $5 \leq K^* \leq 20$ indicates onshore migration and beach accretion. Yet, a further parameter was introduced by Kraus and Larson (1988) to separate barred and non-barred profiles, $P = g H_o^2 / \omega_s^3 T$, where H_o = offshore wave height. A value of 9000 separates barred (greater values) from non-barred profiles (Dalrymple 1992).

Type VI nearshore bars have been found to migrate onshore, offshore, and alongshore, with offshore rates reaching 2.5 m h^{-1} during storms and erosion/accretion rates of 0.05 m h^{-1} (Sallenger et al. 1985; Aagaard and Greenwood 1995). Onshore migration rates are generally smaller, but may still reach 1 m h^{-1} . When the *Type VI* bars are three dimensional, they may migrate alongshore at rates up to 10 m per month (Greenwood and Davidson-Arnott 1975). Ruessink et al. (2000) examined the relative rates of across-shore and alongshore migration using complex empirical orthogonal functions applied to profile data. The alongshore migration rate ranged up to 150 m per day and was strongly related to the alongshore component of the offshore wave energy flux. Short-term variability in bar crest position was shown to be due to changes in the quasi-regular topography, and not to alongshore uniform on-offshore migration. While offshore migration under storms has been clearly related to hydrodynamic forcing, especially the setup-driven undertow (Gallagher et al. 1998) or mean currents modulated by infragravity waves (Aagaard and Greenwood 1995), the onshore migration of *Type VI* bars is poorly known. Generally the motion is attributed to skewed fluid velocities and accelerations (Elgar et al. 2001).

On the Dutch coast a multiple bar *Type VI* system exhibited characteristics of a feedback-dominated system, producing cyclic changes over either 4 or 15–18 years (Wijnberg and Terwindt 1995). Plant and Holman (1997) showed that bars on the east coast of the United States exhibited unpredictable behavior in relation to wave height changes and yet still moved through a sequential pattern of form changes. This paradoxical behavior they related to feedback effects. The forcing for these transitions is as controversial as bar genesis, since direct hydrodynamic forcing has been proposed as well as a self-organization mechanism.

Little work has been done specifically upon bar decay, other than the welding process associated with *Type II* bars (e.g., Davis et al. 1972; Aagaard et al. 1998). However, the one major exception is the study of the multiple bar system

along the Dutch coast. Here, the bar system shifts progressively offshore over time and the outermost bar decays. This has been attributed to the action of highly asymmetric, non-breaking waves (Larson and Kraus 1992; Reussink and Kroon 1994; Wijnberg 1997). Plant et al. (2001) suggest that a morphologic feedback mechanism can lead to bar decay. As bars move onshore under nonbreaking conditions they are also reduced in height; thus they move further away from wave breaking, allowing further bar decay. This has been observed at Duck, NC (Lippmann et al. 1993).

Predictive Models of Bar Genesis and Dynamics

Because of the relatively poor knowledge of long-term bar behavior and the inadequacy of local sediment transport models for the complex nearshore environment, predictive models for the genesis and dynamics of wave-formed bars are still far from complete. In general models proposed are either (1) process-based models (e.g., Bowen 1980), or (2) behavior-based models (de Vriend et al. 1993). The latter range from: (1) highly parameterized models to predict summer–winter (bar–berm) profiles (e.g., Aubrey 1979) or sequential bar evolution (Wright and Short 1984; Sunamura 1988) to (2) statistically based models for predicting bar dynamics (Aubrey et al. 1980) to (3) morphological models to simulate large-scale beach changes (e.g., Cowell et al. 1995).

Cross-References

- Beach Features
- Beach Processes
- Beach Profile
- Net Transport
- Rhythmic Patterns
- Surf Zone Processes
- Wave–Current Interaction

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Bay Beaches

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Definition

Beaches in mostly enclosed bays subject to reworking by locally generated fetch-limited waves.

Beaches are found in bays, sounds, lagoons, and estuaries (here all termed bays) and can comprise a large proportion of the shoreline. Examples include beaches in Chesapeake Bay, Puget Sound, the Tagus River, and arms of the Baltic Sea. The term bay in relation to the open coast is somewhat subjective, and bays such as Monterey Bay, California, may be exposed to ocean waves with energy levels that are among the highest in the world. This discussion is confined to low-energy beaches that occur in mostly enclosed bays where fetch distances for local wave generation are generally less than 50 km. Bay beaches are a subset of low-energy beaches (Jackson et al. 2002) and share many characteristics with beaches in small lakes and reservoirs (Nordstrom and Jackson 2012).

The length of bays greatly exceeds the length of ocean shore in many countries. Beaches are common in these bays, but they are often so small and isolated that they escape attention, except in populated locations. The principal factors affecting the morphodynamics of these beaches are locally generated waves and wave-induced currents, but wind-induced and tidal currents play a role in morphologic change. Fluvial processes may be dominant at estuarine shores in narrow basins or tributaries. At the low end of the wave-energy continuum, other terms, such as stream bank, intertidal marsh margin, or bay bottom, may be more appropriate than beach.

Shore Processes

Waves generated by local winds in bays have low heights (usually mean heights <0.2 m and storm wave heights <1.0 m) and short periods (2.0–4.5 s) (Nordstrom 1992). Ocean waves entering bays play a limited role in beach change where shores do not face ocean entrances. Tidal range affects the vertical distribution of wave energy over the profile, determining the width of the foreshore and the duration that waves break at any elevation. Bay beaches are usually characterized by a steep upper foreshore fronted by a broad, relatively flat bottom, often called a low-tide terrace or subtidal terrace. Spilling waves break in a broad surf zone across the gently sloping terrace at low water levels, but the energy in the waves is low. During high water levels, waves reach the upper foreshore with little loss of energy and usually break as plunging waves. Waves frequently break on the upper foreshore and convert directly to swash without an intervening surf zone. Thus, swash processes play a relatively great role in sediment transport.

Longshore currents are predominantly generated by the breaking of local wind waves, but refracted ocean waves, tidal flows, and wind drift are locally important and may result in flows bayward of the breaking waves that are opposite flows generated by local wind waves. Tidal currents are

important near channels, projecting headlands, and constrictions in the bay, and they may be the dominant agent of sediment transport on the terrace bayward of the foreshore. Ice forms faster and has a greater influence on mid- and high-latitude bay beaches than on nearby ocean beaches because bay waters are colder in winter, shallower, and less saline; ice lasts longer because low wave energies are slow to remove it. Ship and boat wakes are higher on bay beaches than on ocean beaches because vessels can pass close to the shore, but the average energy in the wakes is usually only a small percentage of the average energy of wind waves in all but the smallest bays.

Water level changes can be locally induced by winds blowing across the bay or they can be induced by flow of water through inlets from surges generated on the open coast. Winds can increase water levels on the downwind side of the bay while lowering water levels on the upwind side, but a large opening to the sea on the downwind side of the bay can result in lower water levels downwind.

Beach and Shore Characteristics

The best development of beaches occurs where relatively high wave energies expose abundant unconsolidated sand or gravel in the eroding formations, such as moraines, glacial outwash, fluvial deposits, and landward portions of barrier islands and spits (Nordstrom 1992; Freire et al. 2007). Beaches in small bays, with limited availability of sand and gravel, may be highly localized or confined to ocean entrances. Many beaches have been created by human efforts in urbanized estuaries at locations where no beach would occur naturally because wave energies are too low. Artificial beaches are often wider than natural beaches would be in a similar wave climate. Some new beaches are accidental by-products of landfill operations; some are created intentionally as recreation areas (Nordstrom 1992).

Bay beaches may be unvegetated or partially vegetated and composed of sand, gravel, or shell. Surface sediments are often coarser on bay beaches than on ocean beaches with a similar source. Lag gravel is common on the beach surface, formed from particles exhumed by swash or by preferential elimination of fines by low-energy waves. Individual pebbles move readily over the sand surface, and swash excursions create bands of gravel on the upper foreshore.

The depth of mobilization of sediments on the upper foreshore is shallow (e.g., <0.2 m under storm conditions), and the active beach may be only a thin veneer of unconsolidated material overlying an immobile layer of coarse sediments, clay, peat, or a shore platform. Mobilization of sediments on the low-tide terrace by waves may occur only

to depths of 10–30 mm. Biologic processes can have an influence on sediment mobilization and transport of a magnitude comparable to that of waves and tidal currents alone and play a greater role than wave processes in altering the characteristics of the surface and subsurface of the low-tide terrace (Nordstrom 1992).

Vegetation plays a greater role in influencing morphologic change on bay beaches than on ocean beaches (Nordstrom 1992; Pilkey et al. 2009) because of greater abundance of vegetation in bays and the reduced ability of the low-energy waves to move it. Vegetation helps bind bottom sediment and attenuate wave energies; vegetation flotsam in the breaker and surf zones alters the wave and current characteristics and the likelihood of entrainment of beach sediment; vegetation litter in the wrack line forms barriers to waves, currents, and swash uprush.

Bay shorelines are often composed of numerous isolated beaches with different orientations. The beaches have high variability in morphology and rate of erosion over small areas resulting from local differences in fetch, wind direction, stratigraphy, inherited topography, resistant outcrops on the foreshore, variations in submergence rates, and amounts of sediment in eroding formations (Phillips 1986; Rosen 1980). Beach compartments are isolated into longshore drift cells defined by deep coves or headlands formed by resistant rock, marsh, or human structures.

The net rate of longshore transport on estuarine beaches varies with orientation, fetch distance, and size of each drift cell and ranges from tens of cubic meters to tens of thousands of cubic meters (Wallace 1988). Although rates of transport are low, the magnitude of erosion can be high because the quantities of sediment in transport represent a sizable fraction of the total unconsolidated sediment in the active beach (Jackson et al. 2017). Many bay shores are eroding at greater rates than nearby ocean shores.

The narrow beach widths, wrack deposits, lag gravels, and beach slopes may restrict sand transport by wind, but dunes and deposits of wind-borne sediment covering preexisting topography are more common than perceived. Relatively large dunes can occur as inherited forms where they migrated from ocean beaches or they formed in the bay during more optimum climatic conditions in the distant past or where they have been artificially constructed by human action (Nordstrom and Jackson 2012).

Beach Change

The upper foreshores of most bay beaches are modally reflective. Conspicuous cyclic morphologic change is confined to the immediate vicinity of the foreshore. Sediment removed from the upper foreshore during high-wave-energy events is deposited on the lower foreshore with a change to a concave

upward profile shape. Sediments moved farther offshore onto the terrace form only a thin veneer over the surface instead of forming the break point bar that is prominent on many ocean beaches. Landward and bayward displacement of the entire foreshore profile may also occur, while the profile slope is maintained. This parallel-slope retreat and advance is common when sediment exchange is due primarily to longshore transport and is most pronounced near the ends of drift compartments (Nordstrom 1992).

Resource Values of Bay Beaches

The fronting terrace of a low-energy bay beach has a relatively stable substrate that allows macroscopic plants and fauna to thrive. The upper foreshore is more energetic and may have less species diversity and abundance but nevertheless high ecological value (Dethier et al. 2016). Infauna and macroalgae provide prey to juveniles of commercially valuable fish, and the intertidal area provides habitat for recreationally important clams and numerous species of epifauna and infauna. Fish and invertebrates are prey for foraging birds, especially in the regularly exposed intertidal zone. Wrack from plant litter is inhabited by numerous amphipods and insects. The swash zone and dry upper foreshore are also foraging areas for birds, including upland species.

Bay beaches are not as intensively used as ocean beaches for recreation, but they have important complementary values. They provide convenient surfaces for launching and landing boats and boards for wind surfing. They are favored by parents with small children because they provide a safer environment than on the ocean. Many bay beaches are underutilized for beach recreation because of the unclean appearance of the beaches or lack of awareness of their existence or unique attributes, but ease of access causes bay beaches close to urban areas to have relatively high rates of use (Nordstrom 1992).

Shore Protection and Management

Erosion control strategies for bay beaches may differ from strategies for ocean beaches because of differences in the scale of erosional forces and in the value of resources. Protection programs are facilitated because beach segments are small, isolated drift cells, often under jurisdiction of only one management agency and because small-scale, low-cost protection may be utilized. Low wave energies and gentle offshore gradients make construction of fixed offshore engineering works more practical than on high-energy beaches. Shore-parallel walls are often considered successful because they can withstand direct attack of local waves, they take up minimal space on the beach and adjacent upland, and

they limit the loss of biological resources on the fronting terrace or bay bottom. Detrimental effects of walls include reductions in beach width, riparian vegetation, amount of woody debris, and types of beach wrack and associated invertebrates (Dethier et al. 2016).

Protection projects funded by national or state/provincial governments are often not considered cost-effective, resulting in a fragmented approach to protection by individual property owners. Simple engineering principles are often ignored in constructing small-scale protection structures, including lack of filter cloth or weep holes in bulkheads, failure to build structures deep enough to prevent toe scour or high enough to prevent overtopping, weak fastenings, failure to use adequate sized armor stones, or failure to perform maintenance. As a result, there is much evidence of structural failure. Beach fill is increasingly used for protection or recreation, but fill can cover benthic habitat and eliminate shallow-water areas for aquatic plants. Bayside beach nourishment projects can be inexpensive because only small quantities of fill are required. Fill materials brought in from outside the region may retain their exotic appearance because of limited mixing by low-energy waves.

Beaches in short-fetch environments comprise just one type of shoreline, and vegetated reaches may have greater management interest (Nordstrom and Jackson 2012). Living shoreline projects often eliminate beach environments in favor of marsh habitat. Alternatively, managed realignment projects that involve removing or abandoning shore protection structures can create conditions for new beaches to form. Lack of sediment is an issue facing ocean beaches and bay beaches alike, and sediment resources are likely to be even more restricted in the future (Orford and Pethick 2006). There has been considerable government intervention in decisions on developing bay shores, especially in productive estuaries, but this intervention is rarely conducted to maintain beach resources. Alternative human uses such as transportation, industrial developments, residences, boating, and marsh restoration are compatible with a coastal location according to most policies, and actions to enhance these uses may eliminate beaches. The number and value of bay beaches can be enhanced by implementing beach nourishment projects, altering vegetation, constructing appropriate protection structures, acquiring key sites for public use, and enhancing access. The ease of constructing and maintaining bay beaches and the paucity of quality recreation space in many urban areas make creation of new beaches as surrogates for ocean beaches an attractive option.

Summary

Bay beaches are characterized by short-fetch distances that result in low wave heights and short wave periods, although

erosion rates can be high because of limited sediment availability. Biologic processes play a greater role in beach change than on exposed coasts, and the ecological value of bay beaches is considerable. These beaches provide human-use values alternative to those provided by beaches on exposed coasts, but human alterations can come at the expense of natural values.

Cross-References

- [Beach Erosion](#)
- [Beach Nourishment](#)
- [Beach Processes](#)
- [Cohesive Sediment Transport](#)
- [Cross-Shore Sediment Transport](#)
- [Dissipative Beaches](#)
- [Estuaries](#)
- [Human Impact on Coasts](#)
- [Longshore Sediment Transport](#)
- [Nearshore Sediment Transport Measurement](#)
- [Reflective Beaches](#)
- [Shore Protection Structures](#)

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Beach and Nearshore Instrumentation

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B

Instrumentation in studies of the coast generally, and of the beach and nearshore zone in particular is designed to measure attributes of form and changes in the form (bed) over time, including bedforms; fluid processes related to waves, water level and currents in the water and wind on the beach; and sediment concentration and mass transport rate in the water and on the beach. These measurements may be made at a variety of temporal scales ranging from fractions of a second to months and years and spatial scales ranging from a few square millimeters to hundreds of square kilometers. Some attributes are measured individually, but much of the focus today, and over the past three decades, has been on measurements of morphodynamics, in which the objective is to measure fluid and sediment transport processes and the resulting change in morphology at a temporal scale of minutes to days and occasionally months. Much activity is focused on sandy and to a lesser extent muddy coasts and much of the instrumentation described here is devoted to these, but some work also takes place on the erosion of cohesive clay and bedrock coasts. The highly dynamic nature of the nearshore and swash zone in particular poses many problems for the design of instruments for measuring fluid and sediment transport processes. In addition to the need for very rugged instruments and supports for mounting them, there are difficulties posed by the lack of access to much of the nearshore during storm conditions, and by the presence of bubbles and organic matter in the water column. Ultimately, the instrumentation is designed and deployed to measure particular properties and processes of the beach and nearshore zones, and therefore, this review is organized by the measurement objective rather than particular instrument types.

Measurement of Form and Changes in Form (Erosion and Deposition)

Erosion of Cohesive and Bedrock Coasts

Where the coastline is developed in bedrock, till, and cohesive muds the focus of attention is usually on the measurement of rates of erosion in relation to the strength attributes of the material and the erosional or forcing processes (Sunamura 1992). On a small scale, erosion by weathering and abrasion of rock coasts generally takes place so slowly that measurements are made at point locations on a timescale of months to years. The micro erosion meter originally used to measure

solution of limestone was adapted for use on intertidal shore platforms (Trudgill et al. 1981). It consists of a pointer attached to a micrometer gauge on a mount that can be placed on pins drilled into the rock platform. The mount swivels to allow measurement to be taken at several points around the station so that an average value can be obtained. Measurements are commonly taken at intervals of months or years because of the relatively slow rate of downcutting (Kirk 1977; Viles and Trudgill 1984). A cruder version of the instrument has been adapted for measurements of erodability of tills and clays underwater (Askin and Davidson-Arnott 1981; Davidson-Arnott and Langham 2000). Because the erosion rates are typically up to several centimeters per year, measurements can be made on a weekly to monthly basis.

Measurements of the erodability of fine-grained cohesive muds commonly found in a variety of marine and estuarine environments, have commonly been made with a variety of benthic flumes—essentially inverted channels of various configurations which can be deployed either on the exposed tidal flat or underwater (Amos et al. 1992; Maa et al. 1993; Houwing 1999). Water is circulated through the channel at increasing speeds until the shear on the bed induces erosion and the erosion rate is measured either directly, or indirectly by measurement of suspended sediment concentrations. Recent experiments have also been made with the Cohesive Strength Meter, an automated device which employs a carefully regulated vertical jet of water and monitors the rate of erosion with respect to the impact force (Tolhurst et al. 1999).

Erosion and Deposition of Sediments

While large-scale changes in form can be measured by a variety of techniques described below, these techniques are usually carried out at finite intervals of days, weeks, or months. Measurement of changes in the bed at particular locations on the timescale of dynamic measurements of fluid flow and sediment transport, typically on the order of minutes to hours, has proved to be surprisingly difficult to do in shallow water. One major problem in the nearshore during storms on sandy coasts is the difficulty of distinguishing the bed from the material immediately above it, which is being transported as bed load or suspended load close to the bed. Techniques for measuring changes in elevation at points in the nearshore range from simple erosion rods to optical and acoustical instruments.

Simple measurements of change in bed elevation and the total depth of activation can be made with rods emplaced along a profile or on a grid which are surveyed before and after a storm (Greenwood et al. 1979). The rods can be emplaced by wading and diving. The maximum scour depth can be resolved by placing a washer on the sand surface and then measuring the depth of burial following the storm. Results from a grid of these can be used to measure volume change in the nearshore (Greenwood and Mittler 1984).

Similarly, thin rods can be used on the subaerial beach to measure erosion by wind and thus provide a comparison volume to measurements of sand transport or deposition. Other approaches involving this simple technology in coastal applications include the use of a bedframe device to measure rates of sediment deposition in foredunes (Davidson-Arnott and Law 1990, 1996) and the use of Surface Elevation Tablet (SET) stations in measuring net change in saltmarshes (Cahoon et al. 2000). Recently, automated devices which act in a similar fashion have been developed. One approach uses a vertical array of photocells spaced at a small increment, usually on the order of 1 cm, with the bed being distinguished by either a change in the voltage output or a circuit which can detect where the break is between exposed and buried cells (e.g., Lawler 1992). An alternative method uses the difference in conductivity between sediments and seawater to distinguish the bed level (Ridd 1992). The value of these instruments is that they are relatively low cost and therefore provide the potential for deployment of sufficient sensors to give reasonable spatial coverage across the surf zone.

It should be possible to detect the bed using a small echo sounder mounted on a support above the bed, though this has proved notoriously difficult when there are large amounts of sediment moving over the bed and in suspension. One adaptation of this approach is to mount the transducer on a frame with a sealed stepper motor that permits it to traverse a section of the bed, thus permitting determination of two-dimensional bedform properties and migration rates (Greenwood et al. 1993). Transducers and miniature versions of sidescan sonar have been used more successfully in deeper water where sediment concentrations are much lower. Recent developments in acoustic doppler technology give a much better definition of the bed. Several versions of acoustic doppler velocity profilers (ADCPs) are available which enable the speed of currents to be detected at incremental distances from the sensor. When pointed downward, these are able to distinguish the bed more precisely than simple sonar devices, because the doppler shift is absent from sediments that are not moving (see section below on Sediment concentration, mass transport rate, and deposition for information relating to the ADCP). Small, relatively cheap acoustic sounders are available for use in air and can be used on the beach to measure changes in elevation of the bed or the water surface in wells installed to measure the water table. These devices can also be mounted on tracks to give a profile of changes in elevation during a transport event and the dimensions of any bedforms that develop.

Measurement of Form and Form Change

Measurement of the dune, beach, and nearshore form on a scale of meters to kilometers has traditionally been done using standard survey and hydrographic techniques. Surveys out to the limit of wading have been carried out with levels and

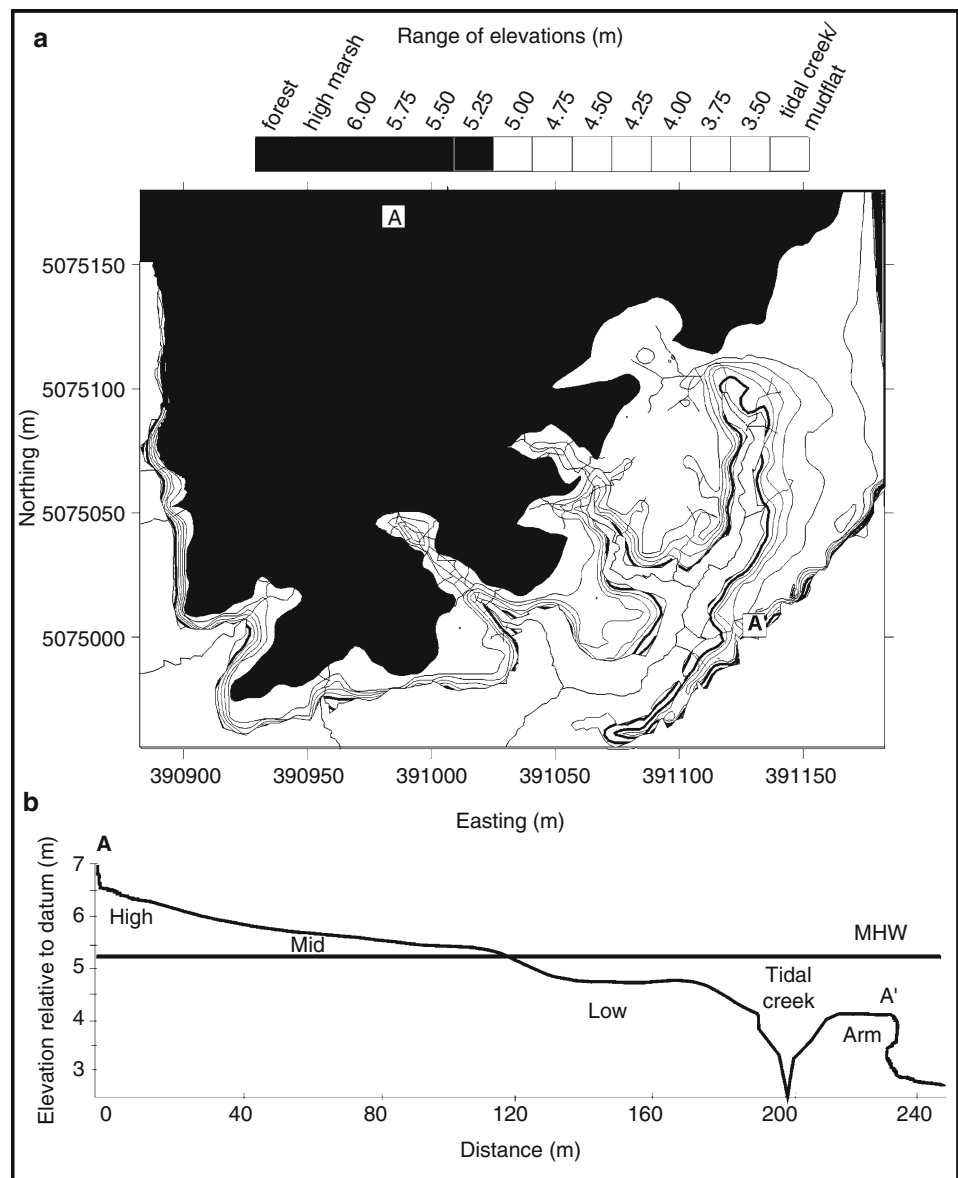
theodolites, and the use of a total station incorporating an electronic distance measurement (EDM) unit and electronic data storage is now standard. These permit rapid surveys over a range of elevations and the output is readily incorporated into a wide range of contouring and geographic information system (GIS) software packages which can produce digital elevation models and permit easy extraction of volume change through repetitive surveys (see Fig. 1). In shallow water, depth has traditionally been determined using standard echo sounders mounted on a boat (Gorman et al. 1998). Digital recording has now replaced the standard paper trace and positional data can be recorded simultaneously using a global positioning system (GPS). Towed arrays or acoustic multibeam transducers can be used to give simultaneous mapping of a wide swath, including information on large bedforms (Morang et al. 1997a). Better definition of the

seafloor and three-dimensional bedform features can be attained with sidescan sonar, which utilizes a towed transducer that emits a signal at right angles to the tow direction and records returns from a swath either side of the transducer (Morang et al. 1997b).

The use of GPS which integrates signals from three or more satellites to determine location and elevation for a variety of surveying tasks, is now becoming standard in measuring beach form and change as it is in so many other fields. Simple systems can give positional accuracy of a few meters and elevation to about 10 m. Much greater accuracy can be obtained through the use of differential systems, which simultaneously capture the signal from the satellites and from a land-based station whose position and elevation is known precisely (see Fig. 1). Moderate priced differential systems make use of Coast Guard beacons, which are set up along the

Beach and Nearshore Instrumentation, Fig. 1

Digital elevation model of a saltmarsh and tidal creeks, Bay of Fundy, Canada, produced from measurements made with a total station and with a DGPS system



coast for navigational purposes. These can give positional accuracy of $\pm 2\text{--}3$ cm and vertical resolution of about double that, though the accuracy decreases with distance from the beacon. More expensive differential systems use a base station set up over a known position and a rover station for the actual survey. The systems can be used to measure the height and position of particular points but they can also be put in a backpack or on a vehicle allowing continuous recording of a traverse. This permits the mapping of linear features such as the waterline, thalweg of tidal creeks, bar crest, and top and bottom of cliffs, thus permitting much better delineation of these features and permitting more accurate delineation of change through repetitive surveys.

A major problem for morphodynamic experiments in the nearshore and surf zone is to obtain measurements of form change during intense storm events. While measurements of sediment transport and nearshore water motion can be obtained throughout an event, most measurements of form change have been obtained through standard surveys carried out during low wave conditions before and after the event. Some data during storms can be obtained from jetties and from specially constructed platforms that span the surf zone. However, some specialized equipment makes data collection during quite high wave conditions possible. These include various sled devices, which can have either a mast with a prism for measurement by a total station or a GPS station to enable position and elevation to be determined. The sleds may be towed by boat beyond the surf zone and winched onshore or a pulley system may be attached to an anchor seaward of the surf zone, enabling the sled to be pulled offshore without recourse to a boat. One highly specialized instrument is the CRAB used extensively at the CERC facility at Duck, North Carolina to carry out a variety of tasks in the water, including surveys in waves up to 3 m (e.g., Plant et al. 1999).

Production of topographic maps from stereo pairs of aerial photographs has been a standard procedure for five decades but photo rectification and automated contouring have required expensive equipment and are rarely used for small-scale beach studies. However, new developments in video technology and digital photogrammetry are making remote measurements of form change much more practical. Video technology has been applied for more than a decade to measure waves and swash run-up (see below) but it has also been applied to measurement of the position of nearshore bars through time exposure of wave breaking (Konicki and Holman 2000; Ruessink et al. 2000; Alport et al. 2001). The intensity of wave breaking is captured by creating time exposure images over a period on the order of 10 min and the resultant smooth white bands outline the zones of wave breaking on shallow bar crests and at the beach. Video cameras can also be used to monitor changes in dynamic features such as tidal inlets and associated ebb and flood tidal deltas (Morris et al. 2001).

The use of digital images from still and video cameras to produce digital elevation models (DEMs) through a variety of computer software packages is of especial interest in mapping coastline changes and changes in the morphology of the beach and foredune area (Chandler 1999). The technique makes use of overlapping pairs of photographs produced either in the traditional way through the movement of a camera installed in a plane, helicopter, or a land-based vehicle, or through the use of images taken from two fixed positions. In the case of aerial photography or moving vehicles, the position of each digital image can be linked to real time positional data provided by DGPS. Where fixed cameras are used on the beach, control points whose position and elevation have been surveyed precisely are used to aid in rectification (Hancock and Willgoose 2001). The advantage of these automated photogrammetric systems is that they can provide a very large number of data points for construction of the DEMs and much of the processing can be automated, thus allowing the evolution of topography over days, weeks, or months to be captured.

On a larger scale, the development of light detection and ranging (LIDAR) technology combined with DGPS permits topographic mapping of both the land surface and the nearshore bed to depths of 10–15 m (Irish and White 1998; Sallenger et al. 2001). The technology makes use of a laser transmitter/receiver, which transmits laser pulses toward the surface and records the traveling time of the reflected pulse. The pulse is reflected from the land surface and, over water, the return from the bottom can also be detected down to depths that depend on the degree of absorption, scattering, and refraction in the water; these in turn depend on sun angle and intensity and on the degree of turbidity in the water. The system can be deployed in a helicopter or fixed wing aircraft. Apart from the unique ability to map both the land and shallow nearshore, the technique offers a relatively low-cost method for determining topographic changes due to major storms and hurricanes (Sallenger et al. 2001), and for surveying changes in areas such as salt marshes and tidal mud flats which are difficult to access with standard surveying approaches.

Winds, Waves, Water Levels, and Currents

Much of the focus in field studies of the beach and nearshore zone is on measuring the morphodynamics of sandy coasts, and to a lesser extent that of muddy coasts. These experiments require measurement of fluid and sediment dynamics over a range of timescales from fractions of a second to hours, days, and months and over spatial scales ranging from a few centimeters to hundreds of meters. Until recently, different instrumentation has been required to measure the fluid dynamics from that measuring sediment dynamics. Over the past three decades, mechanical devices for measuring fluid dynamics

Beach and Nearshore Instrumentation, Fig. 2

Array of cup anemometers and wind vanes mounted on towers to measure wind flow over the beach foreshore at Innisfree Beach, Ireland. Two versions of integrating sediment traps are seen on the left—the smaller traps are cylindrical traps after Leatherman (1978) and the other traps are wedge traps (Nickling and McKenna Neuman 1997)



have been increasingly replaced by solid-state electronics involving the application of a range of direct and remote technologies. Because of the broad range of forcing variables and the spatial scales involved in determination of sediment transport, a wide range and large number of instruments is typically employed. Instrumentation typically involves measurement of wind speed and direction (both for aeolian transport on the beach and for the dynamics of the water surface), water surface elevation, wave form and direction, and water motion.

Wind Speed and Direction

Wind speed and direction have traditionally been measured by some form of mechanical cup or propeller-type anemometer and resistance wind vane mounted on a mast above the water or land surface. These give good resolution of the horizontal wind velocity at a particular elevation. Typically, in studies of aeolian transport on the beach several vertical arrays of anemometers will be deployed in order to obtain measurements of internal boundary layer development and to estimate the bed shear velocity u^* (Greeley et al. 1996; see Fig. 2). The vertical flow can be obtained with systems of three propeller type anemometers or with two mounted at 45° angles as in the K-Gill anemometer (Atakturk and Katsaros 1989). Sonic anemometers now offer the ability to measure fluid motion in all three dimensions, though their size is still large enough to make measurements close the bed (and thus the saltation layer) difficult. Recent modification of the Irwin sensor, a vertical pitot tube that can be mounted flush with the bed (Irwin 1980), offers the ability to obtain direct measurements of wind stress near the bed with little disturbance to the flow.

Mean Water Level

Measurements of water surface elevation are collected routinely to measure changes due to tides, storm surge, and wave set-up and set-down across the breaker and surf zones. Traditional mechanical floats have now been replaced by optical and acoustic sensors installed in stilling wells. Most studies of surf zone dynamics have made use of mean values of the surface elevation measured from wave staffs or pressure transducers, though the accuracy of these measurements is on the order of $\pm$ several centimeters. More precise measurements that can be used for investigating detailed mechanisms of nearshore circulation can be obtained through the use of manometer tubes deployed into the surf zone from the shore (Nielsen and Dunn 1998).

Waves

Field measurements of waves in the inner nearshore and surf zones can be obtained by some form of surface piercing wave staff, which has the advantage of providing a direct measure of the wave form. These systems make use of the conduction of electricity by water, particularly seawater, and record either the change in electrical resistance or capacitance of the system as water rises and falls over a length of uninsulated cable which is part of an electrical circuit (Ribe and Russin 1974; Timpy and Ludwick 1985). The change in resistance or capacitance can be conditioned to produce a variation in an output signal which may be a DC current or a frequency. The sensor itself may be fixed to a support jettied into the sand in the nearshore (see Fig. 3) or attached to some physical structure such as a jetty or platform. The advantage of the surface staff is that it provides a direct measure of the water surface

Beach and Nearshore Instrumentation, Fig. 3

Resistance wave staffs (left) deployed over an intertidal ridge and runnel, Nova Scotia, Canada. A grid of depth of disturbance rods is deployed across the bar (center) and frames supporting OBS and electromagnetic current meters can be seen in the far right



form and they have been used extensively in many studies, particularly in fetch limited areas where installation can be accomplished during calm conditions (Davidson-Arnott and Randall 1984; Greenwood and Sherman 1984).

The disadvantage of wave staffs is that they are subject to high wave forces when deployed in shallow water and they have largely been replaced with some form of pressure transducer housed in a watertight case. These can be deployed some distance below the surface, or on the bed and they are often colocated with other sensors such as electromagnetic current meters and nepelometers (see below). They sense the change in pressure associated with the passage of individual waves. The pressure variation with depth can be predicted from wave theory and thus it is possible to develop a transform function that will relate the recorded variations in pressure to the surface wave form (Lee and Wang 1984). Since there is usually a spectrum of frequencies present in the pressure transducer record, the transform should be performed for all of the frequencies present. This is not a trivial task, though it can be done routinely in a data analysis program. There is some loss of information on the true form of the surface wave as well as the loss of the higher frequencies with increasing depth of deployment, but this is offset in studies in and close to the breaker and surf zones by the ease of deployment and the reduced exposure to breaking wave forces.

The water surface can also be measured remotely using a video camera to measure the change in surface elevation against a graduated pole or screen. This gives a good measure

of the wave form without interference and it enables determination of whether the wave is broken or not. The record can be digitized manually or a computer software algorithm can be used to extract the position of the surface automatically. Video cameras have also been used extensively to extract data on run-up frequencies on the beach (Holman and Sallenger 1985). Recent developments in LIDAR technology may also permit application to measuring waves (Irish et al. 2001).

Individual wave staffs or pressure transducers provide a picture of the variations in water surface elevation through time—that is, they give information on wave height and period but not on the direction of travel. This requires either the deployment of several instruments in an array, which permits determination of the wave direction through a comparison of travel time between various sensors (Bodge and Dean 1984; Howell 1992), or the measurement of both the horizontal and vertical components of water motion using a pressure transducer and bidirectional electromagnetic current meter or an acoustic doppler current meter (see following section).

Because of the rapid oscillatory motion associated with wave action in shallow water, mechanical current meters are generally not useful in studies of fluid processes in the inner nearshore and surf zone, though miniature-ducted impeller current meters have proved useful in some locations (Wright et al. 1982; Masselink and Hegge 1995). Measurements of hydrodynamics in the nearshore and surf zone were revolutionized in the 1970s by the development of electromagnetic current meters (EMCMs) and they have been used extensively in almost all field experiments as well as in the



Beach and Nearshore Instrumentation, Fig. 4 Electromagnetic current meter (left) and three OBS probes mounted on a “goalpost” in the intertidal zone, Skallingen, Denmark. Electronics for the current

meter are installed in the waterproof housing secured to a pole jettied into the sand to the left of the goalpost. A profile line of large depth of disturbance rods is visible at the right

laboratory (Huntley and Bowen 1975; Cushing 1976). Examples of their use can be found in numerous experiments including the Nearshore Sediment Transport Study (NSTS, Seymour 1989), the Canadian Coastal Sediment Study (C^2S^2 ; Willis 1987) and in the various experiments carried out at the CERC at Duck, North Carolina (Birkemeier et al. 1997). The current meters produce a fluctuating magnetic field around the sensor head and measure the voltage generated by fluid flow in the field using Faraday’s law. Typically, the instruments have four sensors mounted orthogonally so as to detect flow along two orthogonal axes. The current meter is usually mounted so as to detect horizontal flow, but it is possible to mount it with one axis vertically. The sensors have a fast response time, permitting sampling at frequencies >5 Hz and are able to detect very small mean flows in a highly fluctuating environment (see Fig. 4). The majority of EMCs used in the field have been made by Marsh-McBirney Inc. of Maryland, USA. Large models have a 10.5 cm diameter head and the small ones, which have been used extensively in the surf zone have a 4 cm head. These current meters have been evaluated extensively (Aubrey and Trowbridge 1985; Aubrey and Trowbridge 1988; Guza 1988) and their widespread use allows for ease of comparison between different studies. While EMCs are now being replaced by various forms of acoustical instrument, they are still useful in the breaker and surf zone because of their smaller sensitivity to the presence of air bubbles.

In the past decade, various forms of acoustical doppler instruments have been developed which have been used in laboratory and field experiments. Essentially, they emit an acoustic signal which is reflected by fine material in the water column from a focal point and received by orthogonally mounted transducers. The relative motion in each axis is then determined by the doppler shift of the signal. The acoustic doppler velocimeter (ADV) is the simplest of the instruments and measures velocity at a single point on the order of a few centimeters from the emitting transducer.

Sediment Concentration, Mass Transport Rate, and Deposition

Obtaining measurements of sediment transport is the third key element in morphodynamic experiments in the coastal zone. There are a large number of instruments available and a variety of approaches have been taken, but much work remains to be done to obtain reliable measurements over a reasonable spatial and temporal scale (White 1998). Estimates of net longshore transport over periods of weeks, months, or years at a location can be obtained from measurements of the amount trapped at a total barrier, either over a short period at a purpose-built groin (Wang and Kraus 1999) or at a large jetty. A number of studies have also used fluorescent or radioactive tracers for measurement of longshore sediment transport or

of transport pathways within the surf zone. However, the focus here is on instrumentation for instantaneous measurement of the transport rate either directly, or indirectly through the combination of measurements of sediment concentration and net water motion. Direct measurement techniques include various traps and acoustic doppler instruments that measure concentration and velocity simultaneously through some portion of the water column. Indirect techniques for measuring concentration include optical and conductivity devices.

Traps and Pumps

A number of devices have been used in attempts to trap sediment suspended in the water column of the net transport in the swash or surf zone, but the oscillatory motion associated with wave action makes this task much more difficult than, for example, under unidirectional flow in a river. Pump samplers have been used with varying degrees of success and various bottles for capturing the suspended sediment load. These all require considerable effort and the logistical difficulties coupled with doubts as to the accuracy of the sampling process have limited their further use. Recently, arrays of streamer traps consisting of long bags of fine mesh fixed to a rigid rectangular inlet have been used to measure transport where there is a net current present. These traps are able to capture large amounts of sediment, but it is still not clear that they can provide reliable estimates of the net transport or that they provide accurate results over a range of conditions.

Optical Devices

Optical devices emit light and give a measure of the sediment concentration in the water column at a point either through the degree of attenuation of the light beam or through the amount of light reflected from particles in suspension. They therefore do not provide a direct measure of sediment transport and thus must be collocated with a device such as an EMCM which measures the fluid flow. The light transmitted is generally in a narrow wave band in the infrared range in order to minimize the effects of natural light in the water.

Transmissometers measure the degree of attenuation of the light over a fixed distance separating the emitter from the receiver. They tend to be relatively bulky instruments best suited for work some distance seaward of the breaker zone in depths greater than 10 m where suspended sediment concentrations are relatively low. A Sea Tech transmissometer with a 5 cm path length was developed for use in shallow water (Huntley 1983) but the much smaller probes associated with instruments measuring reflected light proved more suitable for the inner nearshore and surf zones.

Much of our understanding of the dynamics of suspended sediment transport in the nearshore over the past three decades has come from the use of the optical backscatterance sensor (OBS) originally developed at the University of Washington (Downing et al. 1981) and now produced

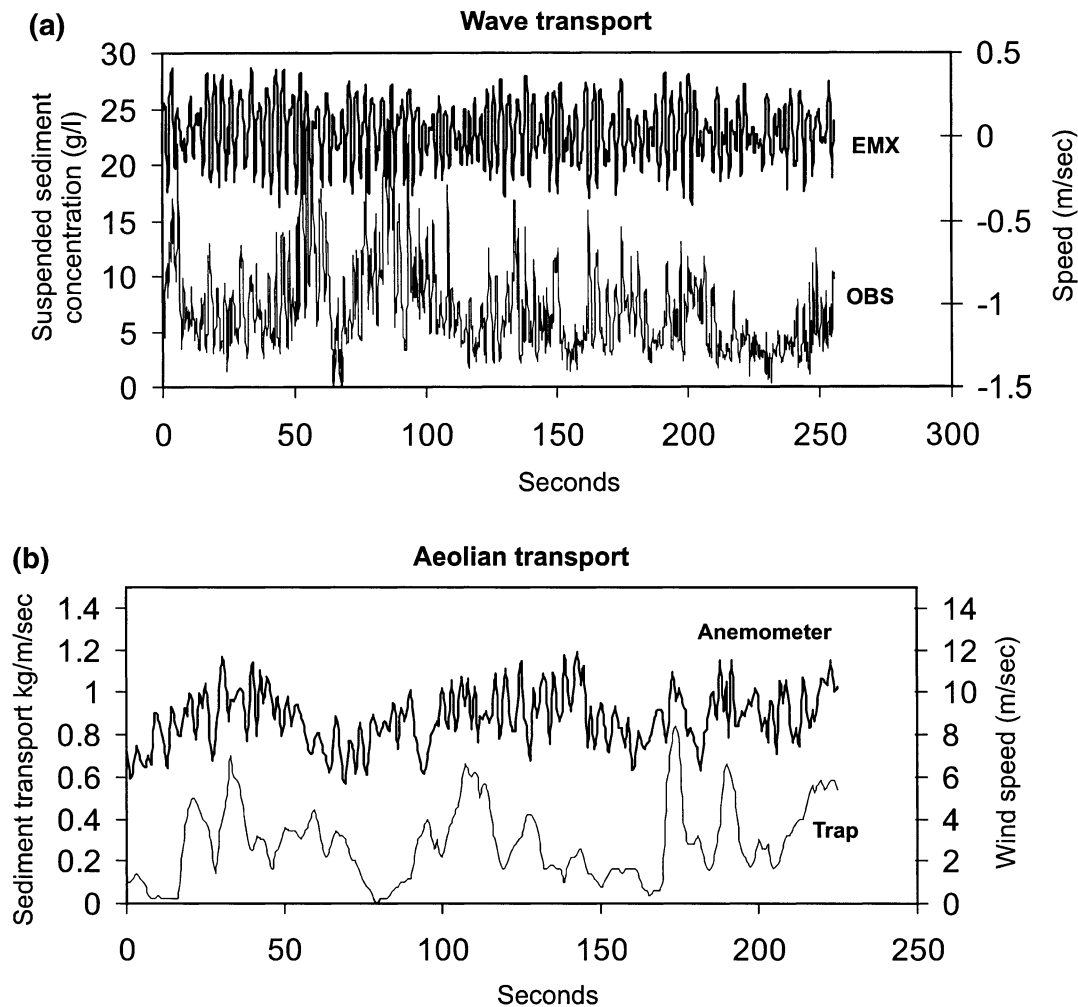
commercially by D & A instruments. The OBS is a miniature nephelometer which measures the backscatterance of light by sediments suspended in the fluid. It utilizes a narrow infrared beam which has the advantage of minimizing interference by sunlight and confining the sampling volume to a short distance from the probe. The sensor is compact (2.1 cm diameter) with transmitter and receiver mounted next to each other at the end of the probe, thus minimizing flow interference and enabling sensors to be mounted in close proximity to other probes or to electromagnetic current meters with which they are often collocated (see Figs. 4 and 5a). The OBS probe is very rugged, enabling it to be deployed in areas of strong currents and breaking wave impacts, and it is clearly superior to other nephelometers for work in the nearshore marine environment (Greenwood et al. 1990). They are designed to measure suspended sediment concentrations in areas where concentrations may be high and/or may vary rapidly over short time periods (i.e., on the order of 0.25 Hz). They have been used in a wide range of marine environments, including the shoreface and continental shelf (Wright et al. 1991; Kineke and Sternberg 1992), the breaker and surf zones (Black and Rosenberg 1994), and estuaries (Kineke et al. 1991).

Provided they are not deployed too close to the bed (interference with the bed itself) or too close to the surface (effects of ambient light) both transmissometers and OBS probes work well. They are linear over a wide range of grain sizes from clay to sand and there has been extensive testing and calibration of both types of instruments in the laboratory and field (Downing and Beach 1989; Osborne et al. 1993; Greenwood and Jagger 1995; Bunt et al. 1999; Sutherland et al. 2000). However, they perform best with a narrow range of grain size and calibration, where there is a wide range of grain size they are subject to considerable error (Bunt et al. 1999).

Routine field calibration of the instruments is difficult and laboratory calibration requires quite complex facilities and involves the difficulties of obtaining representative samples of suspended sediment to be returned to the lab for testing. Recent testing of a laser in situ scattering and transmissometry (LISST) instrument produced by Sequoia Scientific (Traykovski et al. 1999; Gartner et al. 2001; Mikkelsen and Pejrup 2001) offers the potential to measure the complete particle size distribution and concentration simultaneously. Initially, this may offer a means of calibrating cheaper, less bulky sensors but further developments may lead to smaller versions which could be deployed close to the bed.

Acoustic Doppler Velocity Profilers

Acoustical instruments offer the possibility of measuring both particle concentration and velocity simultaneously and over some appreciable portion of the water column, and thus providing a direct measure of the transport rate. Acoustic Doppler velocity profilers (ADVP) use the same basic technology as the ADV described above. However, they measure



Beach and Nearshore Instrumentation, Fig. 5 Comparison of flow and sediment transport in the nearshore and on the beach. (a) On-offshore flow speed measured with an electromagnetic current meter and suspended sediment concentration measured with an OBS probe at an elevation of 0.15 m above the bed on the crest of an intertidal

sand bar (see Fig. 3). The instruments were sampled at 4 Hz. (b) Horizontal wind speed measured with a cup anemometer and sediment transport rate measured with a wedge trap fitted with an electronic balance. The instruments were sampled at 1 Hz and smoothed with a 5 s running mean

the return signal in very small increments of time, thus allowing the determination of velocity and concentration in discrete “bins.” They can be used from a boat with position fixed by DGPS and can thus give a complete picture of flow over bedforms in estuaries and tidal channels (Best et al. 2001).

Aeolian Sand Transport

The design of field instrumentation for measuring sand transport by wind has tended to lag behind that available for measuring transport in the water. Sediment transport from the beach over a period of days or weeks can be measured indirectly by measuring accumulation in vegetated sand dunes by profiling or by the use of a bedframe device (Davidson-Arnott and Law 1990, 1996).

Most direct measurements of sediment transport have been made with simple vertical traps which are oriented

into the wind and which allow the sand captured to collect in the base. The sediment collected over a period of time is weighed to give an average transport rate for the collection period. A major problem is to design the trap so that it is isokinetic, otherwise sand in transport is diverted away from the trap opening by the pressure buildup. Simple vertical traps, which have been used widely may have an efficiency <30%. Wedge-shaped traps have improved aerodynamics and are likely much closer to isokinetic (Nickling and McKenna Neuman 1997; see Fig. 2). However, these traps are more sensitive to changes in wind direction and will undersample when the wind angle exceeds 5°; thus they can only be used for periods of 15–30 min without attention. A variety of other trap designs are available (Goossens et al. 2000) but all have a number of problems with accuracy.

Horizontal traps offer the opportunity to sample all of the transport load and provide a means for calibrating vertical traps. They require a large pit several meters across and again will integrate total transport over periods of tens of minutes to hours (Greeley et al. 1996). Use of a wet horizontal trap can reduce some of the logistics (Wang and Kraus 1999) because the trap need only be a few centimeters deep.

A number of trap designs are now being used to obtain measurements of the instantaneous mass transport rate, thus permitting comparison of the transport rate with measurements of the wind flow. The trap design of Nickling and McKenna Neuman has been modified to incorporate a continuous weighing electronic balance (McKenna Neuman et al. 2000; see Fig. 5b). Bauer and Namikas (1998) used the same trap but designed a combination tipping bucket and strain gage to weigh the sand collected over long time periods. The design of Jackson (1996) uses a similar weighing mechanism to that used by Bauer and Namikas, but the trap itself is a circular collection funnel that is mounted flush with the surface. This avoids the problem of isokinetic sampling associated with vertical traps and has the advantage of omnidirectional collection. However, it measures flux to the surface rather than the total transport rate.

Impact Measurement

The drawbacks of trap designs and the need for high speed, continuous sampling of the transport rate have led to the development of several instruments that measure the impact of saltating grains and then attempt to calibrate this to the transport rate. The saltiphone (Arens 1996) uses a microphone to record impacts and the intensity is then recorded as a voltage signal. While this gives a measure of the relative transport rate, it has proved difficult to calibrate and is sensitive to variations in grain size. The SENSIT (Stockton and Gillette 1990) responds to the impact of saltating grains on a piezoelectric crystal and counts the number of impacts per second. It has also proved difficult to calibrate to give a measure of the mass transport rate and seems to offer only a relative measure of the grains in saltation. Because of the small sampling area a vertical array of the sensors must be deployed in order to measure the total mass transport rate.

Cross-References

- Erosion Processes
- Geographic Information Systems (GIS)
- Global Positioning Systems (GPS)
- Monitoring Coastal Geomorphology
- Muddy Coasts
- Photogrammetry
- Sandy Coasts

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Beach Awards and Certifications

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Definition

A set of administrative and operational elements that, through a process of systematic evaluation of preestablished requirements, guarantee the continuous improvement of the holistic conditions of the beach and publicly recognize its effective management.

Introduction

Beach awards and certifications (BA&C) include all those programs and initiatives that seek public recognition of the management of one or several tourist beaches: awards, eco-labels,

and management systems. This grouping is based on the fact that its evaluation process always seeks the delivery and authorization to use a registered quality symbol, which is identifiable as recognition for achieving management objectives.

This need for clarification is also relevant given that most scientific literature refers more to beach quality rating schemes than to certification schemes themselves. However, the separation between the assessment process (rating schemes) and the recognition of a desired level of quality (certification schemes) is not clearly differentiated in the scientific or technical literature. Therefore, it is assumed that the former are immersed in the latter, since it will always require an evaluation process prior to the delivery of a certification.

The BA&C have been a response to the rapid growth of tourism since the 1950s and 1960s when rapid transport networks were developed in Western Europe and the United States, and the majority of the sectors of society obtain paid vacations. From this increase of mass tourism, it becomes necessary to differentiate destinations and to take better care of the resources that make them attractive. On the other hand, empowerment of civil society for environmental care and quality of products and services has favored the appearance of several certification schemes.

The first matter to clear up is the motivation of a tourist to choose one destination or another, depending on their beaches. Williams and Micallef (2009), like many others, highlight the desire of a visitor to move hundreds or thousands of kilometers to find a satisfactory environmental status, different cultures, and services with competitive costs. Consequently, it can be established that the quality of a destination is based on its natural, sociocultural, and economic elements.

The concept of quality often refers to the degree to which a set of essential characteristics meets preestablished requirements. However, quality can be assessed from several perspectives, such as mercantilist, engineering, statistical, or economical. As an example, the *International Organization for Standardization* (ISO) defines quality according to a product or service and its ability to meet explicit or implicit needs. A more elaborate and increasingly dominant concept of quality is “total quality,” which focuses more on management than on products, as well as seeking to maximize the satisfaction of needs and expectations. Additionally, it should be aware not to confuse quality with luxury or specific services because simplicity can also be a form of quality if it meets the needs of the customer or user.

In brief, concept of quality on beaches can be defined as “the set of essential characteristics that allow to satisfy the expectations of the user of the beach, without risking the dynamic stability of the coastal system.” An example is defined by Williams and Micallef (2009), for whom the quality of the beach is defined by the ability to provide high levels of safety, water quality, facilities, landscape, and

cleanliness. However, each certification scheme usually has its own preestablished requirements to guarantee an optimal level of quality for the visitor of the beach, as well as for the natural environment and local culture.

Evolution of BA&C on the World

The origin of the BA&C can be traced back as much as desired. Corbin (1994) suggests that the earliest schemes could date back to the eighteenth century or earlier, when the “European idle classes” made an itinerary called “the Grand Tour,” which consisted of visiting beaches and coastal towns to enjoy their landscapes (see “► [Tourism and Coastal Development](#)”). However, going back before the 1960s when mass tourism is considered to be born would be more a distraction because beachgoers at that time were very scarce.

Specifically to beaches, the birth of awards and certifications is usually traced back to 1985 when the Blue Flag Program was created in France, having its start in Europe in 1987. This initiative comes from a nongovernmental organization, the *Foundation Europeen Pour L’education Environmental* (nowadays *Foundation for Environmental Education*), which designs a scheme of eco-labeling of beaches based on the evaluation of the quality of bathing water and environmental education. This initiative was reinforced by the European Directive of Bathing Waters of 1976, which included strict levels of environmental quality. Consequently, the European Commission sees an important ally in the Blue Flag Program and begins a symbiotic relationship, the first 12 years with financial support and from 1999 to date in the figure of alliance. Thus, in 1994, there were more than 1,500 Blue Flag beaches in Europe, and at the turn of the century, there were over 2,000 (Cagilaba and Rennie 2005). At the beginning of the twenty-first century, the Blue Flag Program extends beyond European borders reaching Africa (Morocco, Tunisia, and South Africa), the Americas (Puerto Rico, Dominican Republic, Canada, and Brazil), and Oceania (New Zealand); by 2017, more than 4,000 flags were delivered in 46 countries.

From this initiative, a profusion of this type of awards begins, which is accompanied by the sustained growth of mass tourism in Europe, and later in the rest of the world. In the United Kingdom, many similar programs were created, such as the *Seaside Award* (1992), the *Solent Water Quality Awards* (1994), and the *Green Coast Award* (1996); these British schemes are widely referenced, mainly due to the work of the research team composed by Allan T. Williams, Cliff Nelson, and David Tudor. It should be noted that in cold countries, such as the United Kingdom, the culture of beaches is closely linked to landscape and contemplation, whereby the citizen’s interest in beaches becomes the protection of their daily environment, being represented in multiple

organizations with names such as *Keep Wales Tidy* or *Keep Scotland Beautiful*, which, at the same time, administer their own certifications.

On the other side of the Atlantic in the United States, the *National Health Beach Campaign* was created in 1996 as an initiative that emerged from Florida International University and led by Stephen Leatherman, a former student of Allan T. Williams. Likewise, in Latin America, there have been references of at least nine certification schemes since 2003 (Botero et al. 2014; Zielinski and Botero 2015). Apart from these references, there are few reports of other schemes of certification or evaluations of beaches on the planet.

With the turn of the century, along came the catalogs or books that highlighted beaches only for their beauty, or in the best case, with moderately rigorous criteria. Initiatives such as *Guida Blu* in Italy or the *Good Beach Guide* in the United Kingdom present a huge list of beaches, rated according to bathing water quality, safety service, and tourist infrastructure, among other criteria. Within this type of schemes, the list of *Beach Safety* in Australia can also be included, although it is clearly biased to safety aspects. These initiatives are important for sensitizing the general public, but at the same time they weaken the management of beaches and the creation of “collective imagination” of what a “good beach” means.

As a summary, Table 1 lists several schemes for evaluation, classification, and/or certification of beaches, currently applied or in the recent past. These BA&C are notable for having a wide range of aspects of beach quality and an evaluation mechanism with levels that are at least scientifically acceptable. In addition to the schemes included in this table, there are many initiatives that seek to evaluate some criterion of the quality of the beach, but are not considered broad or deep enough to be considered as an integrated beach management tool.

A very quoted initiative is the one carried out for Costa Rica by Chaverri in 1989, which is reviewed by several reference documents on beach management and certification (Cagilaba and Rennie 2005; Vaz et al. 2009; Williams and Micallef 2009). However, there is virtually no information on this initiative since the only bibliographic basis of the references that quoted it is the proceedings of Congress held in the United States in 1989, which are not accessible. The information of a scheme called *Seaside Award* should also be highlighted, which was applied for several years from 1992 in the United Kingdom, but nowadays it is part of the organization that manages the Blue Flag schemes and *Quality Coast Award* in England and Wales (www.keepbritaintidy.org); it should be assumed that this award is no longer valid. Another widely quoted scheme in literature is the *Green Coast Award* which, according to the authors quoting it, was administered by the organization *Keep Wales Tidy*, but no detailed information on its functioning is available.

Beach Awards and Certifications, Table 1 Beach awards, certifications, and rating schemes in the world

Scheme's name	Kind of scheme	Geographical coverage
Blue Flag (www.blueflag.org)	Certification (Ecolabel)	Europe, Africa, Oceania, and the Americas
Bandera Azul Ecológica (www.aya.go.cr)	Certification (Ecolabel)	Costa Rica
Bathing Area Registration and Evaluation (www.bare-beach.com)	Evaluation and rating	Mediterranean countries
Beach Safety in Australia (www.surflifesaving.com.au)	Rating	Australia
Blue Wave & Clean Beach (www.cleanbeaches.org)	Certification (Ecolabel)	United States
Certificación Turística de Playas (www.turismo.gob.ec)	Certification (Ecolabel)	Ecuador
Clean Beaches (www.kabq.org.au)	Award	Australia
Environmental Campaigns (www.encams.org)	Certification (Ecolabel)	United Kingdom
Good Beach Guide (www.goodbeachguide.co.uk)	Rating	United Kingdom
Green Sea Initiative (www.dwrcymru.com)	Certification (Ecolabel)	United Kingdom
Guida Blu (www.legambiente.it)	Rating	Italy
National Healthy Beaches Campaign (www.healthybeaches.org)	Rating	United States
Norma Q (www.aenor.es)	Certification (Management System)	Spain
Playa Ambiental	Certification (Ecolabel)	Cuba
Playa Natural (www.ceadu.org.uy)	Certification (Management System)	Uruguay
Playas y Balnearios de Calidad (www.ambiente.gov.ar)	Certification (Ecolabel)	Argentina
Praia Dourada	Rating	Portugal
Premio Ecoplayas (ecoplayas.rcp.net.pe)	Award	Peru
Premio Ecoplayas (www.ategrus.org)	Award	Spain
Quality Coast Award (www.keepbritaintidy.org)	Award	United Kingdom
Sostenibilidad para Destinos Turísticos de Playa (www.mincomercio.gov.co)	Certification (Management System)	Colombia
Sustentabilidad de Calidad de Playas (www.semarnat.gob.mx)	Certification (Management System)	Mexico

Finally, mention should be made of valuation schemes, which are often confused with BA&C. These schemes are numerous, since they are proposed and, in some cases, implemented by public institutions or universities. One of the schemes is the *NALG Beach Grading* (www.environment-agency.co.uk), which is a litter assessment on the beach and bathing water quality, which is done by the Environmental Agency of England and Wales. On the other hand, it is mentioned in some documents of the program of the Agencia Catalana de l'Aigua (<http://aca-web.gencat.cat/aca>), which includes indicators to monitor three aspects on Catalan beaches: (1) microbiological quality of bathing water, (2) visual quality of waters and sand, and (3) state of access to the beach. These public schemes are important because of their wide implementation, but they have weaknesses regarding the analysis of the results and the updating of the parameters.

From the academic perspective, dozens of beach quality indexes have been proposed by research groups from Spain (Beach Quality Index (Ariza et al. 2010)), South Africa (Beach Evaluation Index (Lucrezi et al. 2016)), Portugal (Beach Quality Index (Semeoshenkova et al. 2017)), Mexico (Integrated Evaluation Index (Cervantes and Espejel 2008)), and Colombia (Environmental Beach Quality Index

(Botero et al. 2015)), among many others. The approaches of these indexes range from the holistic view of the beach as a socio-ecological system to specific aspects such as environmental quality or safety. Although they are rigorous and well-conceived proposals, their systematic application on beaches is almost nil, since in almost no case have they been adopted by institutions in charge of beach quality assessment.

Main Characteristics of BA&C

In addition to being concentrated on the management of the beach, there are four elements that share all beach awards and certifications: (1) voluntary application, (2) granting of labels, (3) promotion of compliance with legislation, and (4) independent evaluation by means of an audit. The first of these, voluntariness, starts from the desire to differentiate a destination from its competitors, since the organization that adopts it is not obliged by law to do so and users to whom it is addressed can choose whether to take it into account or not (UNWTO 2002). As a consequence, voluntary initiatives such as BA&C are seen as a strategy that plays a role in regulating tourism operations while addressing environmental, economic, and social issues.

With regard to the delivery of labels, certifications in their function of differentiating a destination must generate an image representative of the level of quality they endorse. This image is represented on stamps, flags, and other types of symbols, which are publicly exposed on the beach that has met the minimum requirements and wishes to communicate it to their potential customers. The third characteristic of the BA&C, as a promoter of the legislation applicable to tourist beaches, arises from the complex regulations and competences that converge in these coastal systems, as well as from the respect for the legislation that characterizes modern democracies.

Fourthly, audit as a method of verification is based on the principle of credibility required by BA&C to fulfill its primary function: certificate. The audits can be internal or external, the first being used to fine-tune the management of the beach and the second to corroborate the levels of compliance by third parties.

Importance of Quality Certification on Tourist Beaches

Although the importance of certifying quality at tourist beaches may be clear, it is necessary to more widely establish the advantages to undergo the strict procedure that implies for BA&C. Initially, certifications were visualized as a bridge that overcomes the gap between conservation and tourism (Nelson and Botterill 2002). Mass tourism, which is reflected on the beaches as one of its favorite destinations, generates diverse and significant pressures on coastal ecosystems, as well as on the local culture. Despite the fact that sustainable tourism is becoming a paradigm, there are still few beaches where tourism is not accompanied by natural degradation and cultural conflicts. The certifications, therefore, serve as a bridge between tourism vision and the needs of conservation of natural and cultural values, since they demand a commitment of the tourist activity that is developed at the beaches.

On the other hand, the BA&C allow to simplify and communicate the complexity of the beach to meet the needs of a particular audience of interest (Cagilaba and Rennie 2005); this audience refers to the broad group of stakeholders, as well as visitors and tourists on the beach, who must also understand the complexity of the coast. Therefore, BA&C are management tools (see ► “Beach Management Tools”), which require decision makers to be aware that the needs of a particular group may affect the needs of another. Certification schemes, when they are based on a bottom-up approach, become an excellent opportunity to show the consequences of each action on the coast and to prevent socio-environmental conflicts.

Another advantage of the BA&C is that the adoption of a rigorous quality assessment becomes a strategy for improving the beach and the satisfaction of its users (Williams and Micallef

2009). The strictness of following a schedule of compliance with multiple requirements on the beach promotes that all stakeholders are committed, and it becomes a catalyst for agreements between managers, beneficiaries, and funders. Particularly for places where levels of chaos and ecosystemic and cultural complexities are high, the need to follow preestablished procedures, reach agreements, and commit to them is an indisputable advantage and an excellent way of moving toward the regulated teleologies described by Vallega (1999).

BA&C not only promote the organization of stakeholders on the beach but can also be used by the local community to exert pressure on public bodies to take specific actions of general interest (Cagilaba and Rennie 2005). In consequence, when the certification process has been started, the investment of public and private budgets in the improvement of the conditions of the beach is justified with greater solidity. This phenomenon is boosted over time. Once certification is obtained, support in its maintenance will be an obligation of the local authorities under penalty of being blamed for losing a recognition already achieved.

Another relevant aspect of BA&C is the transmission of confidence to users about the tourist quality and socio-ecological protection of the beach (Vaz et al. 2009). In recent years, the scientific community has been insistent on the need to take the perception of users into account, but this feedback is seldom related to the sense of confidence that a certification of beaches can give. Very relevant is the example of beach clubs that have generated such levels of confidence in their clients, that at the end of the summer tourist season have already booked more than 60% of the following year's season (Botero 2013). The benefits represented by BA&C are summarized in Table 2.

Spatial and Temporal Coverage of BA&C

An important aspect to understand certifications of beaches refers to their spatial and temporal coverage. The former is divided into two concepts: (1) geographical area of validity (e.g., countries or places in which certification is valid) and (2) area to certify (e.g., area of the coast that is certified by the seal or flag). In relation to the geographical area of validity, it is an aspect of extreme importance, since it is sought as wide a recognition as possible of the quality conditions of the beach. Botero et al. (2014) found that BA&C in Latin America range from labels at the regional level, such as the case of Playa Ambiental (Cuba), to the Blue Flag applied internationally in more than 45 countries, being the most common national scale which makes it difficult to recognize each scheme beyond the borders of the country of application.

Regarding the concept of area to certify, Botero (2013) identified a surprising vagueness since practically no BA&C clearly defines the criteria to be taken into account to delimit

Beach Awards and Certifications, Table 2 Beach awards and certification benefits (Botero 2013)

Group of stakeholders	BA&C benefits
Tourist brokers	Improving business performance Identification of failures and actions to overcome them Reduction of operating costs Implementation of better food-handling practices Access to technology transfer and technical assistance Differentiation from competitors
Environment and local communities	Protection of ecosystems and local culture Increase and improvement of local conditions of employment Guarantee of free access to beaches and their facilities and infrastructure
Beachgoers	Guarantee of adequate management of the beach Better information to choose the beach to visit Environmental and cultural awareness of the destination to be visited Accessibility for the disabled people and with reduced mobility
Government	Visibility of the destination brand of the municipality or city Stabilization of the local economy, as well as health and environmental protection services Organization and formalization of local activities

the area to certify. The relevance of this aspect is based on the fact that both the beach manager and the certification auditor must be clear of these limits when assessing the requirements since a divergent interpretation can lead to the loss of a certification and the consequent conflict.

On the other hand, it is important to take the temporary coverage of each BA&C into account. All certification schemes have a validity period, after which it must be renewed; however, the renewal periods of BA&C are very diverse. An example is the Blue Flag, which was initially granted at the beginning of each summer and was valid for that summer season, but when it was extended to tropical countries with more than 300 days a year of good weather conditions for sun, sea, and sand tourism, it was adjusted to annual periodicity. The importance of temporary coverage is based on the time distribution of efforts to achieve and improve the quality of the beach while funding such improvements.

The two extreme cases are as follows: (1) a certification with a very short validity, for example, months or weeks, that will be very exhausting in terms of effort and money, so stakeholders will soon be discouraged in their maintenance and will create a constant pressure toward short-term results, and (2) a certification that is renewed every 2 or 3 years,

which generates the perception that the effort must be done only before renovation audits, thus discouraging continuous improvement and long-term management. Although an ideal renewal time cannot be determined, the following criteria may be taken into account: (1) dynamics of the tourist flows of each beach so that the evaluation includes both high and low seasons, (2) ecological cycles in order to evaluate natural conditions at different times, and (3) perception of the stakeholders regarding the progress of their beach so that the motivation is kept constant.

Organizations in Charge of the Certification Process

The evaluation of tourist beaches has an important amount of scientific articles and technical documents. However, issues related to the administration of certification schemes have been less studied. In addition, the information produced by the organizations in charge of the BA&C is few or not disclosed, making it difficult to study.

In general, administration of BA&C is divided into two groups: (1) institutions in charge of administering the certification scheme as a brand and (2) the entity responsible for implementing the scheme on a particular beach. The first group is responsible for multiple functions, which in most cases fall into several different institutions making it difficult to manage. A first function is regulation of the scheme, that is, to define the minimum requirements that must be met on a beach so that certification can be granted. This function is assumed by the institution that creates the schemes, being public institutions in most cases, such as Ministries of Tourism or Environment, nongovernmental organizations, or standardization agencies.

Another function for managing the BA&C is its promotion. Because the certification is an acknowledgment, the massive disclosure of its “brand” is a transcendental factor. That is why permanent efforts in presenting the scheme to the managers of the beaches are necessary in order to motivate its implementation. This function can be performed by the same organization that regulates and creates the certification, but not in all cases. The task of promotion is not very well defined in most of the schemes, becoming additional work distributed “among all” and ending in an action that no one assumes as their own.

In addition to regulation and promotion, there is a central function that must be performed to administer a BA&C: the delivery of the seal or symbol of certification. In many BA&C, the organizations that are in charge of their promotion and regulation are quite clear, but this is not the case with regard to users and even beach managers who must separate the audit visit from the delivery of the scheme.

In addition to certification, there is a higher level of assessment in many BA&C: accreditation. This function refers to the recognition given to organizations that are authorized to audit a particular certification scheme; in other words, it is the accreditation of those who will evaluate compliance with the certification requirements on the beach. This task is usually assigned to a national accreditation body, which is in charge of endorsing the entities that are going to audit the beaches to be certified.

The last function of the institutions in charge of BA&C refers to audit of requirements. This is the most tangible task for those who are requesting a certification of beaches because it is these organizations that go to the beaches and evaluate the fulfillment of the requirements. These audits are carried out mainly by two types of institutions: (1) private companies that have this work as their line of business and (2) standardization institutes, which are normally the exclusive representatives of ISO in each country. The audit is, therefore, a service provided by organizations to those who request it after payment of costs that are usually high compared to other types of services.

This complex set of functions is often distributed differently according to each beach certification scheme. The simplest case is Blue Flag, which is administered entirely by a National Operator, which must be an NGO affiliated with the *Foundation for Environmental Education*, and which receives the exclusivity to administer the eco-label in a given country. On the other side are the BA&C in which each function is performed by a different institution, such as Argentina, Colombia, Ecuador, or Mexico, in which certification depends on a public institution, regulation on standard institutes, auditing on private companies, accreditation on the national agency in charge, and promotion of all at the same time or none at all.

The reason for this dispersion of functions in multiple institutions is supported by the delicate element of “credibility” that must ensure certification. The issue lies in the confidence that should be generated by the public expressions of compliance by companies, which is why third parties, with “recognized prestige and independence,” are required to guarantee that a quality service or product is indeed being delivered. That is especially why an audit is delegated to private companies totally unrelated to the management of the beach, so that they are the ones who “objectively” establish whether the beach meets the requirements to be certified. Although this is the norm in most BA&C, the Blue Flag case, which is the same organization that audits and certifies, is a notable example of well-managed credibility.

On the other hand, it is the second group of organizations which refer to those responsible for the implementation of the scheme on a particular beach. The administration of the BA&C on a beach goes through three stages: (1) preparation of the beach conditions to levels of certification,

(2) application and approval of the certification audit, and (3) maintenance and renewal of the seal or flag.

Stage 1: Preparing Quality Issues on the Beach

The first stage is prior to applying for certification and may take from a few weeks up to several years according to the initial state of the beach. Because of the limited bibliography on BA&C administration, references about timing, costs, and procedures for getting a beach ready to be certified are limited to guides of each BA&C and particular implementation reports.

The first element to prepare conditions of quality on the beach is to characterize the current situation of the beach. The literature on territorial management processes, including ICZM, usually refers to an event triggering the management process, which can range from an external financing project to the conscious mobilization of beach stakeholders (Vallega 1999). In any case, in real situations there is no “start signal” of the process, but the beach manager must identify that conditions are given to undertake this long and expensive path.

Zielinski and Botero (2012) establish five tasks to undertake by beach managers, previous to starting the certification process: (1) cartography of the beach, (2) physical inventory of urban equipment, (3) registration tourist brokers on the beach, (4) register of beach user’s density, and (5) basic environmental diagnosis. These five aspects are important as catalysts for the organization and collective support of improvement actions. Activities such as generation of cartography or environmental diagnosis encourage the knowledge of their beach by the stakeholders, while the inventory of equipment and the registration of tourism brokers direct the process toward organization of economic activities.

It is also important that before publicly announcing the interest in obtaining certification for the beach, a diagnosis of natural, sociocultural, economic, and institutional elements of the destination has been made. By doing this, it is possible to evaluate if the goal is possible, otherwise it will generate expectations that will end up affecting leadership on the beach and future initiatives of integrated management. Particular attention should be paid to local governments and national technicians who, when announcing certification processes for beaches, should consider that these large investments can lead to failures, which undermine the confidence of local communities and hinder future initiatives (Williams and Micallef 2009).

Once the prerequisites have been met, the resources and tools available for the improvement of the beach should be evaluated up to the level required by the BA&C. This is a task that must be planned accurately. To date, no studies on the time or cost of implementing a BA&C are known, but in general it can be established that on a beach that has no formal management process started, the path to certification may take

about 3 years and cost up to one million dollars in case of requiring the construction of infrastructures and provision of facilities. These times depend mainly on the detailed planning of the interventions, since sociocultural processes have very different times from the natural processes and these in turn from the times of construction of civil works. It is not enough to clarify that each process is particular and only a diagnosis of the situation of starting each beach can indicate more precise times and costs.

Stage 2: Requesting Certification of the Beach

This stage begins at the moment when the manager considers that the beach has enough quality levels and formally requests the certification. The most common process is the visit of an audit team which reviews each requirement and evaluates whether it meets with the certification guidelines or not. For this, the audit is based on documents given by the beach manager, visits to the field, and interviews with stakeholders. Notice about the visits is usually given a few days beforehand in order to avoid manipulation of the characteristics of the beach. After the visit, the audit team makes a report which is delivered to the organization that owns the BA&C, who finally decides whether or not to deliver the seal or flag.

A particular case is Blue Flag which performs the entire evaluation procedure internally. Initially, the beach manager must send the National Operator the documentation evidencing compliance with the 33 requirements of the certification. The National Operator must gather all the applications and present them to a National Jury, which is the first evaluator of the application; this National Jury is formed by institutions and experts linked to the services of the beach, such as lifeguards, tourism brokers, environmental organizations, etc. Those candidate beaches that surpass the national evaluation are evaluated later by the International Jury, who finally decides on the beaches to certify. As it is a procedure that includes beaches in almost 50 countries in most latitudes of the planet, they are carried out in two periods: in April for the northern hemisphere and in October for the southern hemisphere and the tropics. Finally, in each country there is a public event of delivery flags to those beaches that approved the evaluation process.

Stage 3: Maintaining and Recertifying the Beach

The third moment starts immediately when the seal or flag is received. Although each BA&C has its own temporary coverage, in most cases an annual follow-up visit is made to revise the quality requirements regardless if the renewal is for 1, 2, or 3 years. Notice of the visits is usually given in advance, since they have a cost that must be assumed by the beach management body, especially in those schemes that are administered by standardization institutes, which have these visits as one of their sources of income.

Maintenance and the renewal of certification goes beyond ensuring quality levels for verification audits. One of the strengths of the management systems and eco-labels is the “dependency” they generate when they are first delivered, in order for the stakeholders and visitors to start recognizing and becoming familiar with the seal or flag. Consequently, if in any of the verification visits the certification is lost, criticism is usually generated to those responsible for the management of the beach. A well-known case is when a Blue Flag is lost, which normally occupies entire pages of local newspapers, being a big problem for the mayor and the technicians in charge of the beach management. Consequently, the best guarantee of continuity is that the certification process is supported by comanagement among all stakeholders, preferably organized around a beach management body (see ► “Beach Management Tools”).

Inclusion of Stakeholders and Decision-Making System

One aspect that reinforces the obtaining and the permanence of the BA&C is inclusion of the stakeholders in the decision-making since the beginning of improvements of the conditions of the beach. As mentioned at the end of the previous section, obtaining certification has the exceptional quality of becoming a common objective, moderating diverse positions on beach management, and acting as a catalyst for collective agreements. The great advantage that the BA&C have over any other beach management tools is that it frames all actions that are carried out on the beach; no other tool is more comprehensive than this.

However, this potentiality can become the greatest enemy of the management of the beach if the processes of participation and decision-making are not carried out properly. As an example, experience of participatory processes in Latin America has been characterized by specific efforts, usually external and based on a technical vision. As a consequence, expectations are generated in local communities, which are later disappointed, either because they expected to continue receiving the economic and material benefits that these external processes generate or because the duration of these specific initiatives is not enough to create habits of self-management.

One of the first precautions that must be taken in the administration of a beach certification is the creation of decision-making centers (DMC), in the sense defined by Vallega (1999). A frequent mistake, as demonstrated by the formulation processes of most BA&C, is the creation of “elite groups,” usually technicians from government institutions who only maintain cooperation between them, turning certification management into a black box, whose operation is unknown to anyone on the outside. This type of DMC creates

a syncretism that promotes generation of rumors and speculations that can even spoil a process of years.

Another relevant aspect in relation to participation in decision-making is the correct identification of each group. In general, there are three kinds of stakeholders: (1) primary, (2) key, and (3) secondary. The bibliography on this subject is very extensive; however, it has been considered relevant to comment on the BA&C of Ecuador, in which the stakeholders were classified based on their importance and their influence for the beach certification process; this classification, in addition to differentiating each type of stakeholders, recommends their participation in the administration of the certification (PMRC 2007).

Weakness of BA&C

Although implementation of beach certification processes is highly advantageous for local communities and ecosystems, it is not without complications. The first is that BA&C should be appropriate for both anthropic and natural beaches so that temptation of urban development in pristine areas leading for certification purposes is controlled. This premise is not currently taken into account by many BA&C, receiving multiple criticisms by conservationists and scientists of various currents, as happens to the Blue Flag (McKenna et al. 2011; Mir-Gual et al. 2015), even though this is the most recognized scheme in the world.

Another aspect, despite the growing technical and scientific bibliography on beach management, is the lack of sufficient rigorous studies on the degree of positive or negative affectation of the local communities where the BA&C are applied. To date, studies have focused on describing the BA&C or assessing their understanding by users, but there is still a gap in information on improving quality of life conditions, meeting human needs and other elements typically defined in integrated coastal management.

A third aspect is the criticism that has been received by the certifications since the 1990s because of its marked top-down approach, as well as the lack of interest in the perception of beach users (Nelson and Botterill 2002). Virtually all certification schemes establish the active participation of the local community as a premise of their implementation; however, when their actual implementation begins, groups of experts are hired, who many times have never even visited the beach, to lead the certification processes. Unfortunately, research is also not frequent to show that BA&C are achieving bottom-up processes.

The current BA&C have also been criticized for not considering aspects related to the beach surrounding environment. Indeed, most of the schemes have serious deficiencies to define the beach limits to certify, normally circumscribed to the surface between the bathing area and the upper limit of

public use, forgetting the connection with ecosystems such as dunes, seagrass, or coral reefs.

The last weakness of BA&C refers to the little information offered to the tourist before visiting the beach. In order to select a destination for their beaches, or even to decide which beach to visit, the tourists need information about the environmental status of the different beaches so that they can choose the one that has the best conditions for their enjoyment (Ariza et al. 2008). However, most BA&C require information from the beach only on the spot, so the decision to visit one beach or another prior to heading to the destination is still a utopia. It is precisely because of this weakness that BA&C have been criticized as merely a promotional tool, although in most cases, the implementation of certification schemes implies real improvements in the quality of beaches. This lack of information has also generated suspicions among scientists on the subject, such as when Williams (2004) posed if BA&C would be, instead of beach management tools, only a placebo for local governments and the community to believe they are doing management; it is a question that is not yet confirmed or rejected with scientific evidence.

Concluding Remarks

The BA&C are tools that cover all the management actions of tourist beaches in environmental, services, safety, information, and ordering aspects. Although the main certification is Blue Flag, there are dozens of similar initiatives in regions such as Europe, North America, and Latin America. Nevertheless, after 30 years of the creation of the first BA&C, scientific information on its operation and effectiveness, at least in English language, is still scarce.

In general terms, most initiatives related to beach quality are divided into two groups: (1) academic models of measurement and (2) awards for compliance of particular aspects in beach management. The former are much more rigorous than the latter, but their level of application is almost nil; moreover, the second are those that are being applied and guiding the management of tourist beaches but with levels of questionable rigor. The question to be solved, therefore, should not be how to evaluate the quality of the beaches, as it has been shown that it is analyzed a lot, but how to pass so much knowledge acquired into the hands of managers interested in certifying their beaches.

Cross-References

- [Beach Management Tools](#)
- [Beach Use and Behaviors](#)
- [Carrying Capacity in Coastal Areas](#)
- [Coastal Management Practices](#)

- [Economics of Beaches](#)
- [Rating Beaches](#)
- [Tourism and Coastal Development](#)
- [Tourism, Criteria for Coastal Sites](#)
- [Tourist Beaches](#)

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Beach Drainage

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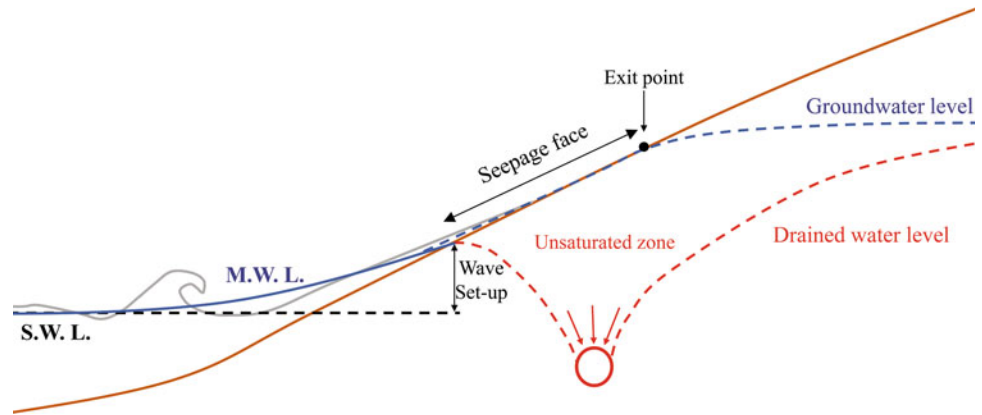
Definition

The Beach drainage is counted among soft coastal protection systems, able to defend sandy beaches from erosion with a low impact on environment. It is constituted by one or more drains placed parallel to the coastline, under the emerged beach face, aimed at enhancing sea water infiltration inside the beach. The pipes are connected through blind pipes with a well, from where the water is pumped out and then transported back to the sea or to another destination. The system induces a water table lowering and, consequently, an increase in the thickness of the unsaturated zone. The relative low water table enhances up-rush fluxes infiltration, allowing sediments deposition on beach face and reducing sea-ward sediment transport. To date, drainage efficacy in restoring eroded beaches is not well defined. For this reason, it can be considered as an auxiliary system for defense works management, such as a beach nourishment in order to increase its lifetime.

Overview

The system was casually discovered in Hirtshals, on the north-eastern coast of Denmark, since a relevant sand deposit was observed around vertical wells used to fill an aquarium with filtered sea water. An on-shore movement of shoreline of about 20÷30 m was observed, and during winter storms, a reduction of shoreline withdraw occurred, resulting in a general accretive behavior of the beach. Based on these observations, the researchers of the Danish Geotechnical Institute began to study the drainage and its possible application as a coastal defense intervention with some adjustment. The vertical wells were substituted by sub-horizontal drain pipes.

Beach Drainage, Fig. 1 General sketch of the BDS



The first patent of a commercial drainage system corresponds to what is normally called Beach Management System (BMS) or Beach Drainage System (BDS), as registered by the Danish Geotechnical Institute in 1985.

Drains consist of PVC pipes with holes uniformly distributed on their contour in order to drain sea water and surrounded by a geotextile layer in order to avoid the blockage by sand, during water infiltration. The system aims at interacting with the swash hydrodynamics by promoting the deposition of sediments transported by waves during up-rush and contrasting their offshore movement during back-wash. In general, the system artificially increases beach absorbing capacity, producing the water table lowering, and consequently, an increasing of the thickness of the unsaturated zone (Fig. 1). The lower elevation of the water table influences the interaction between groundwater and in/ex-filtration within the swash zone, affecting accretional or erosional trends of the beach face at the shoreline.

The Physical Process

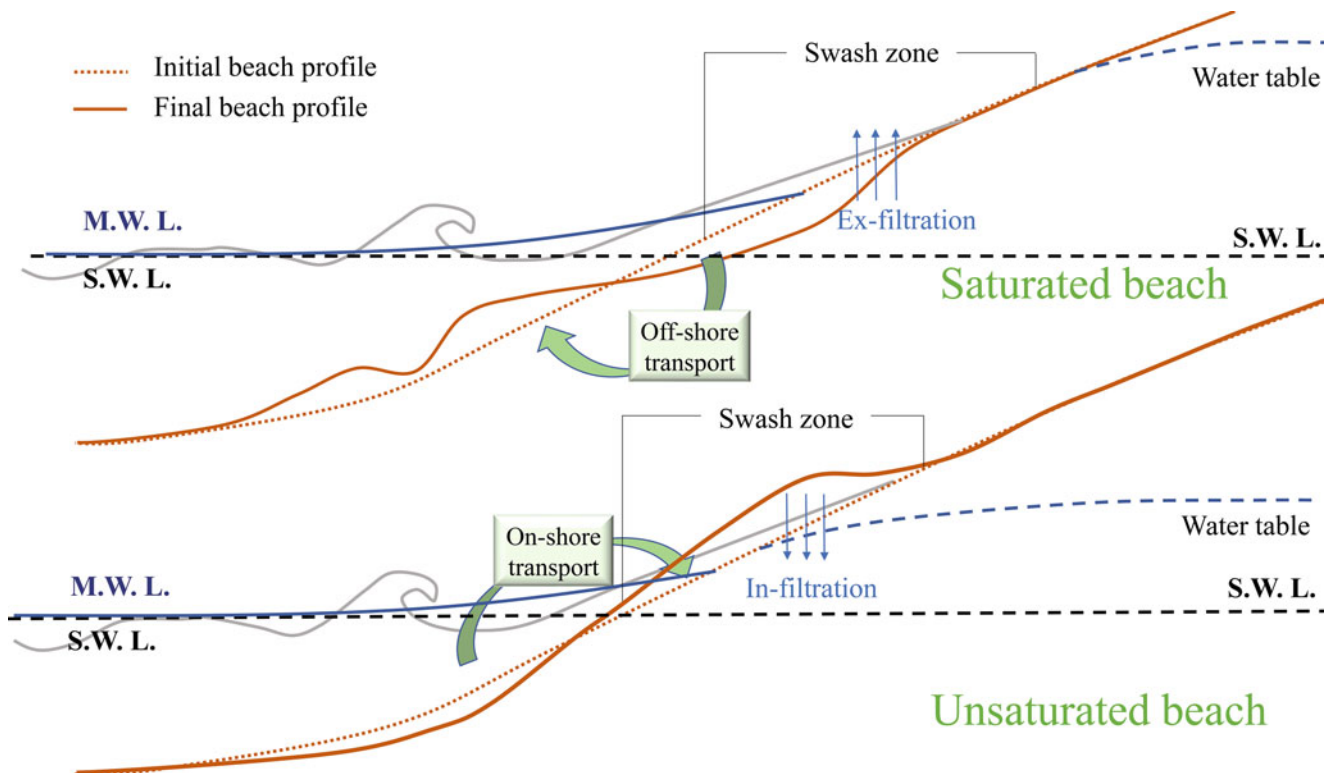
In the swash zone, the sediment motion is affected by several factors (i.e., near-bed velocity-driven shear stress, bore collapse, sediment advection from hydraulic jumps, in-exfiltration). Among them, the water filtration inside the beach through the shore face plays a relevant role. A critical factor driving swash-groundwater interaction is the elevation of beach groundwater relative to the mean sea level. The local elevation of the water table, linked to wave characteristics, influences groundwater processes and their interaction with swash zone hydrodynamics and sediment transport. In particular, a relative low water table enhances swash infiltration and sand deposit on the beach face, while a high water table promotes large back-wash fluxes and a consequent offshore sediment transport. The influence of these processes on beach profile shaping was effectively demonstrated firstly by Duncan (1964) and then observed

by other authors (e.g., Oh and Dean 1995). Results showed a sand volume accretion during the rising tide with a relatively low water table. On the contrary, during the falling tide due to an elevated water table, an increasing of eroded volumes and an overall destabilization of sand were observed.

Such different behaviors can be compared with the typical seasonal barred/normal profile (Fig. 2). When the beach is interested by long-period waves (prevailing during summer), the run-up flow can infiltrate into the partially saturated beach layer, inducing a reduction of the back-wash and the settlement of suspended sediments. Moreover, an enhance of sediment stability occurs, due to a down-directed vertical infiltration, as reported in Oh and Dean (1995), who demonstrated by a simple seepage model, that outflow across the beach face reduces the effective weight of surficial sediments. On the contrary, when the beach is interested by winter storms (characterized by higher wave energy and shorter period), the back-wash flux prevails on the up-rush one, since the water cannot infiltrate through the sand and flushes down, due to the high saturation degree of the beach. The net result of the run-up/down processes is an offshore sediment transport causing the shore erosion.

Field and Laboratory Experiences

Several BDS field installations are available at many sites around the world (e.g., Chappell et al. 1979; Bowman et al. 2007; Bain et al. 2016). Field campaigns provide different responses on system efficiency. Usually they show a sand volume increasing on the emerged beach, with a consequent seawards shoreline movement and a sand stabilization. Moreover, a berm formation with a steep foreshore and a reduction of the swash zone width were observed. However, field results are sometimes difficult to be interpreted and generalized, since many factors influence full-scale systems and avoid a proper experiments control, such as the absence of an adequate



Beach Drainage, Fig. 2 Beach evolution trends in winter (top panel) and summer (bottom panel)

long-term monitoring or difficulties encountered in isolating BDS effects from the natural beach evolution trend.

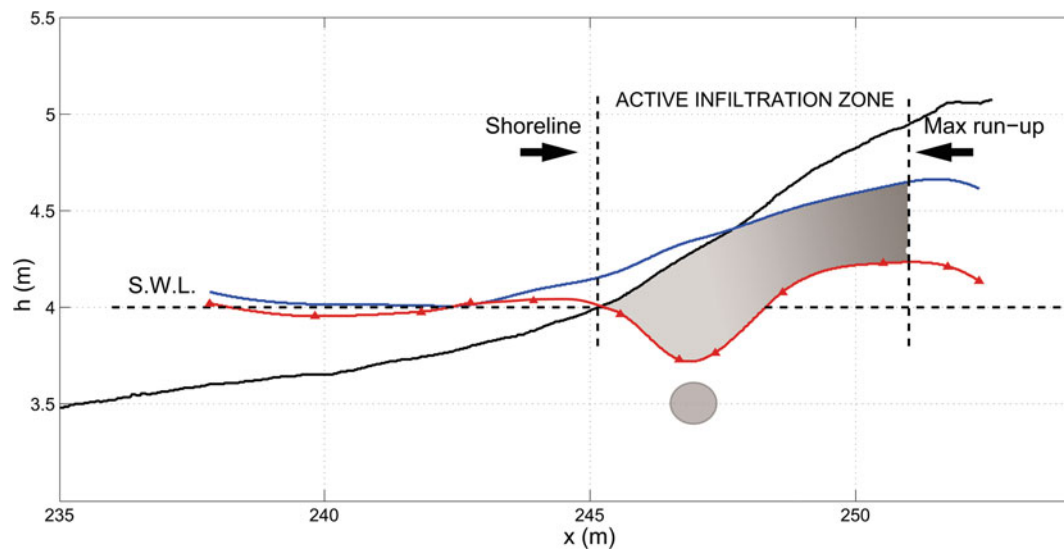
Kawata and Tsuchiya (1987) tested in laboratory the efficiency of a *subsand system* as system of beach erosion control, under controlled conditions. According to obtained results, they showed that under swell waves the subsand filter allowed to accelerate the sediment accretion in the foreshore zone through the development of a berm with a consequent shoreline advance. When wave conditions changed to stormy waves, a decreasing of sediment loss from the foreshore was observed. Ogden and Weisman (1991) found that under both erosive and accretive conditions, the drainage had no effects on shoreline seaward movement, but on berm development and on increasing of the overall beach slope steepening. More recent laboratory tests performed on both small and full-scale physical models (e.g., Damiani et al. 2009, 2011a, b; Contestabile et al. 2012) demonstrate the effects of drain in modifying hydrodynamics in the swash zone.

The comparison between drained and undrained conditions allows to highlight that the system is capable of reducing the wave reflection, wave-induced set-up (Fig. 3), thus affecting the nearshore circulation, namely with a reduction of the undertow currents. Under accretive wave attacks, tests performed in drained conditions showed that beach tends to a configuration comparable with the equilibrium profile typical of gravel beaches. Under high wave energy conditions,

the effects of drain were less evident with a reduction of shoreline withdraw velocity.

General Design Guidelines

Even if design criteria are not yet well defined due to the lack of experience, much information can be derived from recent laboratory experiences. In order to promote drainage effects on beach stabilization, the knowledge of both emerged and submerged beach foreshore represents one of the most relevant factor driving the installation of a BDS. The sediment grain size distribution greatly influences its permeability and, indeed, the beach absorbing capacity. A low permeable porous media would totally neglect drainage effects; besides, coarse sandy as well as gravel beach could be considered naturally drained. According to Wentworth grain size classification, the optimal mean grain size (D_{50}) would range between fine (0.125 mm) to medium (0.5 mm). Foreshore slope has to be quite low, in order to enhance infiltration and great volumes of sediments in shallow waters (i.e., submerged bars) have to be available in the nearshore submerged beach, as a sediments reservoir. The drain planimetric position can be defined as a function of wave run-up. Laboratory experiments, indeed, demonstrate that such a position maximizes drain efficacy, defining an *active infiltration zone*, which



Beach Drainage, Fig. 3 Example of wave-induced set-up reduction in drained conditions, showing the *active infiltration zone* (Damiani et al. 2011a)

properly interacts with swash dynamics. A drain placed landward from the shoreline would reduce the drained flow and, consequently, its efficacy. Otherwise, a system closest to shoreline would cause an increasing of drained water (drained directly from the sea) with a consequent increase of both installation and management costs, without any benefit in terms of beach stabilization. Besides, a drain close to the shoreline would be easier to be exposed and definitively damaged. Concerning the drain depth with respect to the Mean Water Level, the performed experiments do not allow to define an optimal depth. Preliminary numerical simulations (Saponieri and Damiani 2015) based on full-scale physical model (in static conditions) show that a *limit drain depth* can be defined, highly depending on beach permeability. It can be identified as the maximum drain depth in correspondence to which the water table reaches its maximum and corresponds to the depth at which drain pipe begins to work in pressure.

Summary

The Beach drainage can be considered as a useful assistance for sandy beach management, under certain specific conditions as described, even if its effects have not yet been fully addressed. It is able to modify swash zone hydrodynamics and sediment transport, since it artificially induces the water table lowering inside the emerged beach, within the swash zone. Such a new hydrodynamic condition promotes the on-shore sediment transport and sand settlement. Nevertheless any specific design criteria have been yet implemented, both field and laboratory experiences remarked the importance of preliminary studies on beach characteristics (i.e., foreshore slope, sand grain

distribution, fall velocity, permeability) as well as wave climate, in order to improve drain efficacy in coast-stabilization.

Cross-References

- [Beach Erosion](#)
- [Beach Profile](#)
- [Cross-Shore Sediment Transport](#)
- [Swash Zone Dynamics](#)

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Beach Environmental Quality

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Definition

State of the beach as a socionatural system in a certain time, concerning its ecosystem functionality and satisfaction of human needs such as leisure, subsistence, and identity, within sanitary, ecological, and recreational dimensions.

Beach Environmental Quality

Environmental quality is a concept with a fuzzy definition in the literature, since beaches are moving between biocentrism approaches to sanitary parameters. Similarly, the general definition of beach quality helps to consolidate the concept of beach environmental quality (BEQ) because it suggests integration of indicators which represent the natural component with human welfare. According to Botero et al. (2013), most of the references related with BEQ concept focus on the description of natural characteristics of the beach environment, along with the condition of beach pollution caused by human activities or natural dynamics.

In general terms, approaches of BEQ are linked to the generic assessment of “beach quality,” which concentrates almost exclusively on the evaluation of the welfare perceived by beach users in terms of esthetic, hygiene, and cleanness conditions. Another approach commonly registered in scientific

literature is the degree of physical security to which users are exposed in beaches, in terms of health risks associated to the contact with water and sand. Since most leisure activities linked to sea and sun tourism imply a direct contact with the elements of the environment, beach sand, and water have been separately inquired and in detail while conceiving them as environmental quality indicators. Further approaches associated to the concept of beach environmental quality include the design of management measures, because they give a significant importance to the environmental condition of the beach (Williams and Micallef 2009). In addition, the terms “environmental health” and “ecosystem health” are frequently found on scientific publications, which consider beaches as the research purpose. Finally, considering the aptitude of this coastal unit, there is an approach joined with beach uses, which seeks to satisfy human needs such as recreation, protection against natural events, and conservation of natural heritage (Ariza et al. 2008).

In general terms, beach environmental quality has three components: sanitary, ecological, and recreational. The first is related with the risk to human health which is caused by human themselves through the water pollution or dangerous litter on the sand. Parameters to measure these components used to be microbiological pathogens (e.g., coliforms and enterococcus) or with presence of vector-animals (e.g., dogs and pigeons). The second component is linked with natural functions on the beach as a natural system, which is studied from ecosystem health and ecosystem services approaches. Sentinel species and ecosystem stressors are parameters useful to measure this component. Thirdly, tourist beaches should satisfy some human needs such as subsistence, leisure, and identity, which is clearly explained in the human scale development approach (Max-Neef 1993). Safety, beach cleaning and scenery are some parameters to measure this component. To sum up, BEQ should be analyzed from a wide approach which integrates these three components of the socionatural system.

The measuring methods of BEQ depend on the component which will be evaluated. To begin, the sanitary component is commonly measured from samples in the bathing water and sand, as well as counting of particular elements on the beach. Within this component, microbiological parameters and litter are commonly evaluated. Second, the ecosystem component is estimated by methods connected to the biological sciences, such as counting species and assessing the degradation of the ecological climax of beach ecosystem and its adjacent ecosystems. Finally, the recreational component is quantified based on users’ surveys and expert evaluations, such as the case of the coastal scenery that has been applied in many countries in the world (i.e., Ergin et al. 2004; Anfuso et al. 2017) or physical security (Botero et al. 2014). However, there is not yet a condensed version of standardized techniques to measure the BEQ, which serves as a general guide

for beach managers, as happens with the measure of fresh and potable water since decades (APHA 2009).

Within the variety of parameters to be evaluated, the measuring of microbiological variables is marked for its importance as a pollution indicator. The samplings of water should follow the conventional protocols, which start from maintaining the cold chain, applying stabilizing chemicals, and performing analytical laboratory analyzes. The sampling of sand is less documented, so to proceed with the samples in the area with the greatest presence of users, making small holes to take the sediment is a common practice. Around the world, the most recognized recreational water quality standards are the European Directive 2006/7/EC (European Parliament and the Council 2006) and the guide published by the World Health Organization (WHO 2003) which suggest the use of the filtration technique by membrane in order to identify fecal coliforms (*Escherichia coli*) and *Enterococcus*. Furthermore, a variety of literature is focused on the quantitative assessment of user's risk to exposure of water contaminated with *fecal coliforms*, concluding on many situations this happens due to wastewater discharges from hotels or deficient wastewater treatment. The importance of analyzing the health risk in BEQ is because of the possibility of developing infections caused by *Enterococcus* as a consequence of the use of contaminated marine waters. Researchers from around the world have reported a strong correlation between low quality of bathing water and gastrointestinal disease in swimmers, as well, skin infections from occasional visitor to beaches with a low BEQ.

The BEQ's sampling from spatial information has been proved as the most innovative due to the usefulness for coastal areas. Geospatial techniques have been explored as an instrument to measure beach environmental quality, which integrate acquisition of data (i.e., satellite systems, airborne platforms, or video camera monitoring systems), storage, integration, and representation of such data through GIS platforms and processing programs (Pereira 2014). Drones or unmanned aerial vehicle (UAV) systems are also a promising technique because they allow an economic mean of data acquisition, considering the amortization of the investment with the possibility to observe several beaches. Besides, these systems are tied to specific software that facilitates the analysis of the images captured by the UAV, generating geotagged photos and mosaic of images that allow the observation of the phenomena taking place on the beach (Pranzini and Rossi 2013). Such outputs allow to identify beach equipment and civil infrastructure that characterize the degree of organization and stiffening of a beach, which are part of the recreational component of BEQ (Botero et al. 2014).

Several research teams and institutions have proposed general frameworks to study BEQ, although few of them have covered a holistic view. In Colombia, the ICAPTU project establishes a model which integrates the three components above mentioned in order to define an index including three

indicators with the purpose of synthesizing the environmental quality of a single beach in a single number (Botero et al. 2015; Pereira 2015). Another example was proposed by Cervantes and Espejel (2008), who designed an integrated index to assess recreational sandy beaches using descriptive beach indicators, user perception and economic value indicators in Mexico, Brazil, and United States. Ariza et al. (2010) also proposed the Beach Quality Index based on the analysis of natural, protection, and recreation functions of beaches, which includes 13 subindices to evaluate the quality of urban and urbanized beaches in the Mediterranean area. On the other side, Semeoshenkova et al. (2017) presented another proposal of Beach Quality Integrated Index in order to assess the "environmental quality" and "human well-being and health" in three types of beaches in the Adriatic coast (urban, semiurban, and semirural). Other example was proposed by McLachlan et al. (2013) who conducted a review of physical, ecological, and socioeconomic factors important to determine two simple indices: one index about conservation value and the other index refers to recreational potential. When these two indices are combined, a beach can be simply classified as suitable for intensive recreation, conservation, or mixed use. Lastly, in Chile, two conservation and recreational indices, plus urbanization levels, were used to identify deficiencies in the infrastructure levels and tourist security offered to users (González and Holtmann-Ahumada 2017). To conclude, although several indexes have been proposed during last decade, information about their application for monitoring is still inexistent.

Summary

Beach environmental quality refers to the state of some specific conditions which are measured directly or indirectly on the beach, within the DPSIR framework. Although, scientific literature about environmental factors of beaches is numerous, the majority of them are related with few parameters and do not consider the holistic approach.

The BEQ as a whole includes sanitary, ecosystem, and recreational components. Each one of them has their own conceptual framework and parameters. As a result, several methods and techniques could be used to measure each environmental variable, from field sampling to lab analysis and remote sensing. As a conclusion, it is important to mention that some proposals of indicators have been developed during the last years, covering different levels of analysis and integration, but none is yet universally accepted.

Cross-References

- Beach and Nearshore Instrumentation
- Beach Awards and Certifications

- Beach Management Tools
- Beach Use and Behaviors
- Carrying Capacity in Coastal Areas
- Marine Litter
- Tourist Beaches

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Beach Erosion

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Definition

Beaches are loose accumulations of sand, gravel, or a mixture of the two that bound an estimated 30% of the world's coasts (Bird 1996). Beach erosion is a process whereby a beach loses its sediment, resulting in a depletion of its sediment budget. This process implies that the beach can no longer balance energy produced by waves, currents, and water piling up against it, leading to further net sediment loss and lowering and retreat of the beach. Sometimes, beach erosion also occurs concomitantly with beach retreat through overwash by waves. Basically therefore, beach erosion may be viewed as resulting from an imbalance between, on the one hand, the energy inputs and, on the other, the resistance of the beach bed and sediment liable to be mobilized by the fluid forces. The erosion process itself is thus a way of eventually reestablishing balance through dissipation of energy.

Beaches: Natural Buffers of Wave Energy

Because they consist of more or less loosely packed non-cohesive sediments, beaches act as buffers that absorb, reflect, and dissipate energy delivered to the shore by waves. By doing so, they shelter areas behind the beach, especially during storms, from wave attack and flooding. Such back-beach zones may be cliffs, dunes, or low-lying marshes and lagoons. Beaches also provide various other ecosystem services such as nutrient cycling, water purification, nursery habitats for resource species, and feeding-breeding habitats for various species of shorebirds and endangered sea turtles, while also serving as important nodes for economic development and cultural use (Nel et al. 2014). On many coasts of the world, beaches have been taken up by economic development (Pilkey and Cooper 2014). A lot of this development has occurred over the last four to five decades, thriving on the

Beach Erosion, Fig. 1 Chronic erosion near Cannes, French Riviera, that has led to an extremely narrow beach, thus endangering housing and infrastructure on the upper beach



worldwide growth in domestic and international tourism and largely favored by the diversification of beach recreational activities. The boom in coastal development has been matched by an increasing awareness that the beaches that form the foundations of prosperity of many communities are eroding in many places, and increasingly in peril (Pilkey and Cooper 2014; Bird and Lewis 2015). Lack of foresight in construction and development planning has, in many cases, led to massive and irreversible coastal urbanization that renders many communities vulnerable to the insidious effects of beach erosion (Fig. 1). Similar effects have been caused by rampant beach aggregate mining. Erosion impairs the capacity of a beach to act as a buffer against storms. This means that beach erosion may have serious negative repercussions for low-lying island states, for shorefront communities, and for beach-based leisure activities on which depend many jobs and from which many coastal communities draw income. Erosion also has negative ecological impacts (Schlacher et al. 2014), as shown by the destructive effects on many ecological beach-dune communities in the wake of superstorm Sandy (Burger et al. 2017).

Beach studies started in earnest mainly in connection with military activities, notably during the preparation of the World War II landings on the French coast. Since then, they have increased dramatically, especially over the last four decades, with increasing diversification of concepts embracing the abiotic (physical/geomorphic), biotic (biology/ecology), and application (management, resource use, health, climate change and sea-level rise) domains (Nel et al. 2014). Evaluating the implications of beach erosion necessitates a clear definition of what beach erosion is and how it is measured, notwithstanding the fact that the physical processes involved in the dynamics of erosion are still not well understood.

Perception of Beach Erosion and Measurement of Erosion Rates

Good beach management requires accurate quantification of beach erosion rates. This is important, especially as regards the establishment, within a long-term coastal management framework, of construction and development setbacks. Although beach erosion has received great attention from coastal scientists, government agencies, administrative authorities, and beachfront owners, its perception and exact definition are controversial issues (e.g., Smith et al. 2017), mainly as a result of the diverse interests of the different parties involved in beaches and/or their management (Esteves and Finkl 1998). This holds true for beaches in many developed countries. Designation of eroding east coast beaches in Florida by the Florida Department of Environmental Protection has, for instance, been found to be problematic, since 65% of this shoreline is currently identified as eroding when, in fact, only 20% eroded between about 1971 and 2007, a period of widespread beach nourishment (see below) to combat erosion (Houston and Dean 2015). There are no common standards for objectively classifying beach erosion, and the process is not commonly perceived as a problem on undeveloped shores. Perception of the problem generally tends to be associated with developed shores in urban areas mainly where sandy beaches are deemed as important to the economy, but it should be noted that nondeveloped beaches may have high ecological value (Armaroli et al. 2012).

The difficulty inherent to a standardized definition of the shoreline (Boak and Turner 2005) also implies that there are no common standards for quantifying rates of beach change and for determining, for instance, the high-tide line, one yardstick for identifying the beach shoreline (Ruggiero and

List 2009; Smith et al. 2017). Beach erosion is generally quantified through some statistical treatment of retreat rates and volumetric losses (Anfuso et al. 2016). The input data are either from field surveys that have gained in accuracy with the advent of electronic stations and differential global positioning systems (GPS) or from numerically rectified aerial photographs, maps, land-use documents, as well as from data generated by other newer methods such as video monitoring (e.g., Holman and Stanley 2007; Angnuureng et al. 2016), airborne laser scanning (Koroglu et al. 2017), or light detection and ranging (LiDAR) (e.g., Levoy et al. 2013; Pye and Blott 2016). Specific treatment of readily available LANDSAT images can also yield rates of beach change at a high resolution (Liu et al. 2017; see also entry on “► [Shoreline and Coastal Terrain Mapping](#)”). High-resolution photogrammetry based on the use of drones or light aircraft (Brunier et al. 2016; Turner et al. 2016a) is a technique of beach monitoring that is developing rapidly. Beach erosion rates and volumetric losses may also be estimated from the depletion of beach nourishment volumes (e.g., Roberts and Wang 2012; Bird and Lewis 2015, see entry on “► [Beach Nourishment](#)”). Rates of beach erosion may range from net moderate losses of less than a meter a year to several meters following just one storm event. Such rates may also vary alongshore, attaining a maximum in “hot spots” (List et al. 2006) or high-tide shoreline areas subject to the most severe perturbations. There are a few examples at the global scale of reliable quantification of long-term (decadal to multi-decadal) rates of beach change from regular profile surveys (e.g., Ruggiero et al. 2016; Turner et al. 2016b).

Beach Erosion Processes Within the Profile

The beach is a three-dimensional (3-D) sediment body that extends alongshore from the upper limits of wave run-up to the outer limits of significant wave action, the so-called wave closure depth, in the nearshore zone. However, whereas the upper limits may be relatively easy to identify using geomorphic features, the offshore limits are not, for obvious reasons. Beach erosion may be a short-term (order of hours to seasons) process that reflects adjustment to wave energy changes (storms, and rarely, tsunami waves), or a longer-term (order of years) one that reflects an increasingly deficient beach sediment budget (Fig. 1). On sandy beaches, short-term erosion is commonly part of a cycle of morphodynamic adjustment of the beach profile to seasonal or nonseasonal changes in wave energy. Seasonal changes commonly correspond to the classic winter profile flattened by storms, and the summer profile that accretes under fair-weather conditions. Beach-profile adjustment generally leads to better absorption of the nearshore and incident wave energy, resulting, over a more or less long period of time (hours to months), in an

equilibrium situation and shoreline stability. The period of adjustment depends on the wave energy, the beach morphology, and the sediment volume. Rapid beach recovery is quite common, but rates of recovery are also generally variable and recovery may sometimes take several years following major storms (Brooks et al. 2017). Such recovery is important for beach resilience, an important parameter in the long-term sustainability of a beach. Sandy beach morphology and sediment volume are intricately related, defining either short, steep, reflective beaches, commonly associated with coarse sediment, or wide, flat, dissipative beaches, commonly with fine sand (e.g., Wright and Short 1984; Aagaard et al. 2013; see also entries on “► [Reflective Beaches](#)” and “► [Dissipative Beaches](#)”). In the former, much of the sand is locked up in the intertidal beach, forming especially a voluminous sub-aerial beach sometimes comprising an upper beach terrace called a berm. In the latter, much of the sand is stored in shallow intertidal to subtidal bars. High-energy waves impinging on steep reflective sandy beaches result in the fastest response times, resulting in erosion of the upper beach and berm, and seaward removal of the sand to form barred intermediate and dissipative beaches. In such situations, erosion of the upper beach is compensated by accumulation on the lower beach, without there being necessarily a net loss of sediment. However, coastal managers are sensitive to changes in subaerial beach volume, so that such short-term upper beach and berm erosion, especially where severe, may raise anxiety.

Beaches characterized dominantly by gravel (Fig. 2) differ in their behavior. Such coarse-grained beaches have generally steep, narrow reflective profiles that are commonly inert and unresponsive, to a certain degree, to increases in wave energy. This is a result of the spatial organization of the constituent clasts and the capacity of these coarse-grained beaches to absorb wave energy through high percolation rates (Buscombe and Masselink 2006; Orford and Anthony 2013). Macrotidal beaches, found in areas with large tidal ranges (>4 m at spring tides), also commonly show slow or moderate response to high-energy events, compared to their more common microtidal counterparts, because of their wide, multi-barred profiles and the rapidity of migration of the wave domains as a result of the important tidal excursion (Wijnberg and Kroon 2002; Masselink et al. 2006).

On any sandy or gravelly beach profile, short-term morphodynamic changes may be embedded in longer-term changes involving net sediments gains or losses, the latter being synonymous with overall beach erosion throughout the profile. Whatever its origins, a net loss of beach sediment results in durable changes in beach morphology as the beach seeks to adjust to this situation of sediment deficit. Durable erosion may generally result in a permanently scarped upper beach profile (Fig. 3). In some cases, erosion can lead to the total disappearance of a beach. The sediment lost accumulates



Beach Erosion, Fig. 2 Gravel beaches, such as that of in Nice, on the French Riviera, are generally highly reflective. Seawalls behind gravel beaches can prevent onshore roll-over of gravel clasts during storms, a

mechanism that can lead to the maintenance of these coarse-clastic beaches over time. Chronic erosion has necessitated nourishment (see below) of this beach since 1976

Beach Erosion, Fig. 3 Durable beach erosion is commonly manifested by an upper beach scarp. The erosion here, which threatens coastal settlements and has led to inland relocations of several villages and an international highway, has occurred downdrift of the port of Cotonou in Benin, West Africa, on a coast subject to strong longshore drift



elsewhere, either further alongshore, on other beaches, in estuarine and lagoonal sinks, or in offshore sinks.

In spite of a considerable amount of beach research over the past five decades, the sediment transport processes operating on beaches, notably those involved in cross-shore

sediment movements, are still poorly known (Aagaard 2014; see entries on “► [Beach Processes](#)” and “► [Surf Zone Processes](#)”). “Normal” waves and wave shoaling processes in the nearshore zone lead to net shoreward flows that may result in sediment drifting alongshore or from offshore working its

way toward the beach. In essentially reflective beach systems, these shoreward flows are balanced by gravity-driven seaward return flows down the beach face. The net sediment budget of the beach would then depend on its ability to diminish the seaward return flow volume through processes such as infiltration and grain size and bedform adaptations to flow strength. These processes depend on beach slope and grain size, both interrelated, and on the water table on the beach. Low, dry season water tables on microtidal beaches composed of medium to coarse sand in West Africa favor exceptionally steep beach slopes (up to 18°) that result from berm buildup through sand deposition by swash infiltration. Higher rainy season water tables have the opposite effect of encouraging little infiltration, thus favoring sediment transport down the beach. Gravel beaches (Fig. 2) are essentially reflective and subject to very high percolation because of their coarse grain sizes. The dynamics of these beaches are fully discussed elsewhere (Buscombe and Masselink 2006; Orford and Anthony 2013; see entry on “► [Gravel Barriers](#)”).

The complexity of beach erosion processes can only be apprehended from process monitoring that relies on the use of rapidly developing modern techniques involving field and, less commonly, laboratory (wave flumes) experiments, and, increasingly, using numerical wave and beach morphological models. On sandy beaches subject to episodic high-energy erosive events, a variety of related meteorological and hydrodynamic factors combine to enhance beach stripping. These are setup of the water level close to the beach, due to waves, onshore wind forcing, and sometimes the direct exposure of the beach to the low-pressure systems that generate storm waves, and durable saturation of the beach face through both rainfalls that commonly accompany stormy weather and diminution of percolation during multiple swash run-ups. The sediment-stripping processes depend on the complex interplay of various modes of fluid motion near the beach. These include incident gravity waves, or infragravity waves generated by transfer of energy from the former, alternations of high and low waves or wave groupiness, wave-induced currents, tidal currents, currents due to wind forcing, wave-current interactions, and patterns of energy concentration related to the way these various fluid forces interact with the morphology. On a few of the world's coasts subject to strong, sustained winds, eolian processes may also contribute to the long-term removal of sand from the beach. Generally, water pileup on the seaward-sloping beach must be balanced by seaward return flows, or in low-lying sand and gravel barrier systems, by over wash. In the course of these high-energy events, liquefaction and removal of beach sand may be accompanied by deposition of these sediments in offshore areas where the energy of seaward flows peters out or is balanced by shoreward-directed energy. The beach profile retreat is also a short-term mechanism of creating

accommodation space for the temporary water pileup. The sediment transported seaward may initially travel via major rip current pathways generated by incident and infragravity waves. Sand may flow alongshore from these pathways, because of the commonly oblique incidence of waves and wind setup, feeding strong longshore currents. Subsequently, as larger waves and strong winds lead to more pileup of water on the beach, the seaward return flows may no longer be simply canalized in rip channels and mass balancing seaward flows may occur, resulting in generalized beach stripping and offshore sediment loss. The intensity of beach erosion depends on various factors such as the wave energy level and the duration of high-energy conditions, the antecedent beach morphology, wind speed, and orientation relative to the coast, beach grain size, tidal range, and tidal state. Gravel beaches are commonly subject to landward clast transport by overwash during storms (Orford and Anthony 2013; Almeida et al. 2017), but this important process of long-term gravel barrier maintenance can be blocked by seawalls, leading to beach erosion. Seawalls are also detrimental to sandy beaches by favoring greater seaward reflection of wave energy.

Erosion of Pocket Beaches

Many beaches along the world's coastlines are short (tens to a few hundreds of meters long) pocket beaches that are commonly associated with rocky shorelines (Trenhaile 2016), the dominant shoreline type on Earth (Emery and Kuhn 1982). Pocket beaches are commonly swash-aligned beaches, but examples of pocket beaches subject to longshore sediment displacements have also been documented (Turki et al. 2013). Swash and drift-aligned beaches (see entry on “► [Drift and Swash Alignments](#)”), respectively, designate beaches associated with weak and strong rates of longshore drift (Davies 1980). Pocket beaches occur on coasts of all wave-energy ranges from high-energy Atlantic coasts (Dehoucke et al. 2009) to low-energy Mediterranean coasts (Bowman et al. 2009) and variably exposed coral-reef settings (Jeanson et al. 2013). Pocket beaches are particularly abundant along the rocky shores of the Mediterranean, niched in embayments that commonly have limited sediment accommodation space as well as sediment supply (Anthony 2014), but some beaches are sourced by episodic inputs from ephemeral streams (Pranzini et al. 2013). Pocket beaches are likely to be particularly prone to sea-level rise, especially along urbanized coasts, because of limited sediment supplies and the lack of space for retreat (Monioudi et al. 2016). On many developed coasts, notably in the Mediterranean, artificial pocket beaches have also been created (Bertoni and Sarti 2011; Anthony 2014), commonly bound by engineering structures (especially T-groins) that are built to contain the beach sand or gravel and minimize erosion.

Longshore Manifestations of Beach Erosion

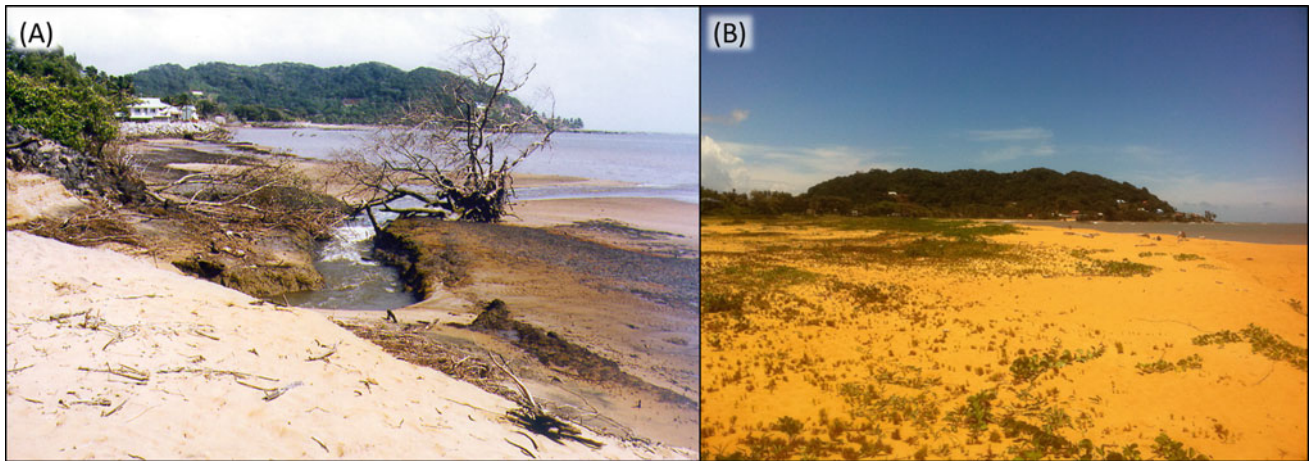
A beach may comprise one or several sediment cells (Carter 1988; van Rijn 2011) with bounding limits to longshore drift. In many cases of beach erosion, the process is a subtle, insidious one that does not require the high-energy events described above (although these may spectacularly enhance erosion rates) other than seasonal increases in wave energy to which the beach is generally well adapted. This is particularly the case on coasts subject to strong longshore drift that determines the overall stability of the beach. Erosion functions essentially where major engineering structures block the sediment load drifting alongshore. Continuity of sediment transport downdrift by the longshore current is assured by beach erosion. The strongest drift rates, sometimes exceeding one million $\text{m}^3 \text{a}^{-1}$ of sand or gravel, are found where large frequent swell waves impinge with marked obliquity on long, open beaches, as on the Gulf of Guinea coast in West Africa, in Australia, New Zealand, South Africa, parts of the coast of Brazil, and the Kerala coast of India.

The longshore manifestations of beach erosion have received attention in the literature, both in terms of the plan shape of freestanding beaches (as opposed to short, headland-bound bay beaches) and of the effects of major engineering structures. The plan shape of freestanding beaches may change rapidly in response to sediment depletion. These changes basically reflect sediment cell divisions (Carter 1988) that may also involve switches from drift to swash alignment, in an attempt by the beach to adjust to sediment deficit by diminishing longshore transport. The large-scale changes in beach plan shape are also accompanied by beach textural and profile reorganizations. Downdrift of jetties, a major cause of beach erosion (see next section), the high-tide shoreline morphology in plan commonly defines a log-spiral curve (Hsu et al. 2010; see also entry on “► **Headland-Bay Beach**”) that may extend several kilometers. This shape illustrates the more severe retreat that affects the beach just downdrift of such structures. Continuity of sediment transport by the longshore current after the jetty is assured by sometimes rapid and significant beach erosion. Erosion diminishes downdrift of this “hot spot” as the longshore current becomes increasingly charged with sediment, leading to a more linear high-tide shoreline. At some distance downdrift, erosion becomes nil and the high-tide shoreline may even show advance from the accumulation of sediment eroded from the beach updrift. These longshore changes are sometimes manifested by a fast “erosion front” and a slow “erosion front” separated by a salient, or “bump,” that may exacerbate erosion downdrift (Bruun 1995). The existence of such two fronts along any eroding beach probably reflects two sediment cells on either side of a central downdrift accumulation terminus fed by beach erosion within the more updrift cell. The accumulation “bump,” or salient, influences wave

incidence angles in such a way as to minimize drift and capture sediment, thus aggravating erosion within the following longshore cell.

Research into shoreline instabilities and complex shoreline shapes has been significantly enhanced by the high-angle wave instability model (Ashton et al. 2001). The model focuses on large-scale (kilometer and greater) alongshore sediment flux generated by high incident wave angles. Although wave breaking is associated with a complex range of smaller-scale longshore processes related to a diverse range of fluid flows, the model neglects these smaller-scale interactions. Deep-water waves approaching directly onshore (i.e., where the angle between deep-water wave crests and the shoreline is 0°) do not generate longshore transport. Such transport increases as the relative angle between deep-water wave crests and the shoreline increases from 0° or decreases from 90° , with a maximum alongshore transport for some angle between 0° and 90° . This transport maximum implies a fundamental instability in shoreline evolution. As waves approach from angles greater than the maximizing angle (“high-angle” waves), the sediment flux along the crest of a perturbation in plan-view shoreline shape must converge, implying that more sediment enters control volumes along the crest than exits, and in the long term, this forms a growing shoreline bump. The high-angle model has been invoked to explain several types of natural shoreline instabilities such as capes, flying spits, and sand waves (Ashton et al. 2001; Ashton and Murray 2006).

Embayed beaches commonly show oscillating patterns of erosion and accretion, but it is also not uncommon for seasonal or longer-term changes in the predominant direction of wave approach to induce changes in longshore drift and generate beach rotation, which is the periodic lateral movement of sand towards alternating ends of the embayed beach (e.g., Thomas et al. 2016; Bracs et al. 2016). The process results in erosion at one end of the beach, while the other accretes. A recent study has shown, however, that offshore sand bars involving cross-shore sediment fluxes can induce significant alongshore variability, especially in embayed rotating beaches where headland sheltering of oblique waves is pronounced (Harley et al. 2016). In rare cases, beach rotation is due to short- to medium-term (order of a few years) changes in nearshore bathymetry that affect wave refraction and dissipation patterns. The massive mud banks delivered by the Amazon River to the muddy coast of South America migrate westward toward the Orinoco delta, inducing changes in incident wave energy levels by strongly modulating wave refraction and diffraction patterns (Anthony and Dolique 2004; Brunier et al. 2016). These generate lateral movement of sand in embayed beaches between bedrock headlands in Cayenne, French Guiana, resulting in alternations in erosion (Fig. 4) and accretion over time, without net sediment loss (other than through illicit sand extraction).



Beach Erosion, Fig. 4 An example of a beach in Cayenne, French Guiana, affected by periodic rotation, with temporal alternations of erosion (a) and accretion (b), concomitant with accretion and erosion

at the opposite end of this embayed beach, due to the effects, on beach refraction and longshore beach sand transport, of large nearshore mud banks migrating alongshore

Similar effects on beaches elsewhere may be generated by changing sand bank configurations offshore, as in the case of the sandy beaches of northern France bounding the English Channel and the southern North Sea.

The Causes of Beach Erosion

The sediment that accumulates on the shore to form a beach may come from various sources. Any poorly consolidated material on which waves and currents impinge may be a source. Such material may be an initial coastal and nearshore deposit of diverse origin such as glacial till or fluvial sediments, or may be delivered to the shore through landslides or by volcanoes. These sources are usually cut into coastal cliffs and underwater slopes that recede as they deliver sediments to the shore for beach construction. Dunes may also deliver sand to the beach but the beach and dunes, especially those immediately bounding the beach, should be considered as an inter-related system with sand interchanges. Some infilled estuaries and many sand- or gravel-rich deltas also supply sediment to beaches, especially at times of high river discharge.

Any natural or human action that affects the supply capacity of a given source and the cross-shore and longshore sediment transport processes on beaches may result in erosion. In many cases, especially on long open beaches, several factors, the specific roles of which are difficult to disentangle, may jointly cause beach erosion. The most readily discernible causes of beach erosion are where identified human actions and activities perturb the beach sediment budget (van Rijn 2011). This cause of beach erosion dramatically developed in the twentieth century with the multiplication of dams across rivers and large-scale urbanization of the coastal zone worldwide (Pilkey and Cooper 2014). On many coasts of the world,

the construction of dams has, over the long run, led to coastal sediment starvation and beach erosion. The effects of artificial structures on the shore (see “► [Shore Protection Structures](#)”), and especially beaches, have received considerable attention in the literature (Cooper and Pilkey 2012; Pranzini and Williams 2013). One important cause of beach erosion worldwide is the construction of jetties, harbors (Fig. 5), and ports. Deepwater ports constructed on the sandy beaches of the Bight of Benin in Togo, Benin, and Nigeria, beaches subject to one of the strongest rates of longshore sand transport in the world, have resulted in dramatic beach erosion downdrift of the port breakwaters, and in equally spectacular beach accretion updrift (Laïbi et al. 2014). The continual beach erosion on this coast (Fig. 3) has led to successive inland relocations of coastal communities and of the major international highway linking the three countries, at great cost to their already beleaguered economies. In many coastal communities, as along the eastern United States, the lagoons behind barrier islands are important economic waterways the inlets of which need to be deepened by dredging and kept open permanently by groins and breakwaters. These impede the longshore drift of sand that is vital in nourishing beaches downdrift (Houston and Dean 2015). Esteves and Finkl (1998) estimated that 90% of beach erosion in southeast Florida prior to 2000 was caused by human action, mostly the construction of deepened inlets with protective jetties. The modifications carried out on most of the 21 inlets along the 588 km of sandy shoreline on the Florida east coast have usually caused significant downdrift shoreline erosion (Houston and Dean 2015).

Another source of perturbation of beach sediment budgets and a cause of beach erosion is coastal urbanization, which involves the development of urban fronts with high-rise condominiums and hotels on the upper beach, as along the Mediterranean rivieras (Anthony 2014). In many cases,

Beach Erosion, Fig. 5 Beach accumulation (foreground) and retreat (background), respectively, updrift and downdrift of the breakwater of the harbor of Hornbaek, northern Zealand, Denmark. To maintain the beach downdrift, bypassing of sand is carried out to the tune of 20,000 m³ a year (personal communication, Aart Kroon)



related dune systems have been flattened or severely degraded, and this has had a dramatic effect on beach stability. Dunes tend to be overlooked as the “savings account” of the shore while the beaches act basically as a “checking account.” The dunes store important volumes of sand that help in balancing the beach budget. Uncontrolled shorefront urbanization has commonly entailed narrowing of many beaches, diminishing in time their wave-energy buffering capacity, and leading to beach erosion. This practice is still frequent on the shores of many emerging economies where development pressures are intense (Fig. 6). Beachfront urbanization also often requires defense structures, notably walls and revetments emplaced on the upper beach. Depending on their design, these structures may act as static barriers that reflect wave energy offshore, thus aggravating beach erosion (Fig. 2). In the past, urbanization and the development of road and rail networks has sometimes involved the direct quarrying of sand or gravel from beaches with fragile sediment budgets.

To stabilize already eroding beaches, groins are sometimes built across the beach with the aim of trapping sediment drifting past. Breakwaters are also sometimes built off the beach to dissipate some of the wave energy that erodes the beach. In playing an efficient role sometimes in alleviating beach erosion, these structures may be instrumental in simply transferring the erosion problem further downdrift, often to the detriment of another community. It is not uncommon to see groin fields sprouting downdrift, increasing in numbers as the problem goes from one community to the next. The gravel barrier beach in Normandy and Picardy, France, is a clear

illustration of this downdrift “march” of erosion and of the attendant groin field. In Picardy, a groin field (Fig. 7) emplaced to stabilize an eroding gravel beach grew from 6 groins in 1976 to 1996 groins in 2000 over a distance of 10 km. The initial erosion of this beach started with the construction of several jetties updrift in the nineteenth and twentieth centuries and was aggravated by the artificial consolidation and stabilization of several sectors of cliffs that hitherto liberated gravel flint clasts to the beach longshore-drift cell. This practice of cliff stabilization has, in some cases like this one, led to beach sediment starvation and erosion. The situation is similar along many beaches, especially on developed coasts.

Natural sediment depletion should also be considered as a cause of beach erosion. On many of the world’s coasts, especially in areas where sea level over the past 5,000–6,000 years has been relatively stable, the sand forming the beaches was derived from sediments on the inner continental shelf. There are, however, very few examples where this mechanism has been demonstrated, because of the requirement of having long-term datasets that cover the simultaneous evolution of both the beach and the nearshore zone. Examples have been shown from Denmark (Aagaard 2011) and the US Pacific coast bounding the Columbia River mouth (Ruggiero et al. 2016). Along large stretches of the Australian, West African, and Brazilian coasts, drowned near-shore deposits are considered to have been reworked by waves and sand derived from these reworked deposits driven onshore to form barriers, sometimes several kilometers in width, comprising multiple beach ridges and dunes.

Beach Erosion, Fig. 6 Strong tourism development pressures on the Mediterranean coast of Morocco, near Tetouan, are leading to dune flattening and beach erosion. The dune sand is used directly in situ in the building industry



Beach Erosion, Fig. 7 An example of a gravel beach with multiple groins that have been successively constructed as erosion has “marched” downdrift (background) following updrift (foreground) stabilization of cliffs that hitherto supplied flint clasts to the beach



This process, called progradation, has stopped in most areas as the nearshore sediment supply has petered out. The beaches bounding these prograded coasts are sensitive to any long-term changes in wave energy, resulting, for instance, from greater storminess or sea-level rise. Although exhaustion of

nearshore sediment stocks is commonly invoked as a cause of beach erosion, this claim is hard to substantiate because of the lack of records of long-term beach and nearshore profile changes, compounded by our poor understanding of cross-shore sand transport processes (Aagaard 2014).

Beaches, as mentioned earlier, may show short-term changes in profile in response to storms and fair weather conditions. Apart from the various causes of sediment depletion evoked above, changes in the state of the sea can also lead to durable beach erosion. These changes include greater storminess, short-term variations in sea level related to major changes in sea surface temperatures such as those associated with El Nino events, and, increasingly, secular sea-level rise imputed to global warming (e.g., Masselink and Gehrels 2014; Jimenez et al. 2017; Serafin et al. 2017). Changes in offshore wave energy are due to storms, such as the northeasters in the eastern United States and cyclones, or may reflect more subtle increases in wave energy due to greater storminess and sea-level rise. Exceptional waves generated by submarine landslides or earthquakes may also lead to significant beach erosion, as in the case of the Tohoku-Oki tsunami in 2011 (Tojo and Udo 2016). These events generate destructive high-energy waves that remove the beach sediments offshore (Yoshikawa et al. 2015). The seaward flows may lead to temporary, and sometimes permanent, losses of sand beyond the offshore limits of the beach profile. On beaches bounding low-lying coasts, permanent losses of sediment may also occur inland as waves wash over the shore. Greater storminess implies more frequent episodes of higher incident wave energy often accompanied by strong wind setup of water level inshore. Many beaches do not have the available sediment stocks to adapt to such increases in wave power and to the currents resulting from wind and wave forcing. Extreme water levels (e.g., Serafin et al. 2017) and sea-level rise, either on a short-term basis, due to short-term events such as El Nino, or forced by intra-seasonal meteorological and oceanographic processes unrelated to storms (e.g., Theuerkauf et al. 2014), or of a secular nature due to global warming, would similarly favor wave energy impingement higher up the beach face (see entry on “► [Changing Sea Levels](#), ► [Sea-Level and Climate Change](#)”). New sediment commonly does not move in from alongshore to balance the increase in wave energy, and the beach erodes as its sediment stocks are transferred seaward. Depending on the wave energy regime and the rate of sea-level rise, such sediment may be permanently trapped offshore as the base of wave action moves upward through sea-level rise. This pattern of beach erosion resulting from sea-level rise has been extensively debated in terms of what has become known as the Bruun rule.

Managing Beach Erosion

Good beach management requires both accurate bookkeeping on rates and patterns of beach change and implementation of the right strategies in the face of erosion. Beaches are a multi-resource asset in many ways, involving huge sums of money in developed economies, both in terms of revenue and for

design and implementation of management policies. As a result, the number of parties involved in beach management may be considerable, ranging from state legislators and engineering bodies, through recreational and tourist agencies, to scientists, beachfront home owners individually or as associations, and environmental and ecological pressure groups. Beaches are, as such, objects of conflicting interests. In many developed economies, and increasingly in emerging economies, beach erosion has become the fundamental coastal zone management problem (Pilkey and Cooper 2014), and a national issue in several countries bordered by long stretches of densely developed low-lying shores, such as the Netherlands and the United States. Beach erosion has also become a cause of major concern for low-lying island states and coral-reef islands subject to sea-level rise (Forbes et al. 2013; Romine et al. 2016).

In the face of beach erosion, the management options are very few indeed. These include the determination of beach development setback lines in order to accommodate future erosion. In the absence of precise determination of beach erosion rates, this strategy may fail, as on the Bight of Benin coast in West Africa. A second strategy is that of letting erosion take its course, generally in undeveloped areas where the process does not constitute a hazard. However, this should not detract from the value of undeveloped beaches that could have high ecological value (Armaroli et al. 2012). A third strategy is that of containing erosion, especially where vital national, economic, or military interests are at stake, as in the Netherlands. On some developed shores, such as the city front of Nice on the French Riviera and parts of the coast of Florida, the value of beach real estate and the revenue from beaches are such that the high-tide shoreline position has to be maintained, generally through the implementation of costly solutions, especially regular beach nourishment (Anthony et al. 2011; Roberts and Wang 2012; Armstrong et al. 2016) but also through efficient bypassing of inlets, such as in parts of south Florida (see entry on “► [Shoreline Response to Littoral Drift Barriers](#)”), and bypassing of harbors (Fig. 5). Beach nourishment (Fig. 8) can also be used to maintain wide beaches that protect urban and industrial estates, as on the North Sea coast of Dunkirk (Spodar et al. 2017). This “soft engineering” technique have been discussed in numerous papers in scientific journals, especially the *Journal of Coastal Research* and *Shore and Beach*, as well as in regular newspaper commentaries in many countries. A clear correlation has been found, however, between regular beach nourishment and development in places especially vulnerable to damage. In a comprehensive analysis of all shorefront single-family homes in the state of Florida, houses in renourished zones are significantly larger and more numerous than in non-nourished zones, thus suggesting a positive feedback between nourishment and development that could aggravate exposure to risk in zones already characterized by high vulnerability



Beach Erosion, Fig. 8 An example of a recently massively nourished beach on the North Sea coast of France near Dunkirk. Photo credit Adrien Cartier, Géodunes (see also Spodar et al. 2017)

(Armstrong et al. 2016). In fact, from about 1971 to 2007, a period of widespread beach nourishment in Florida, only about half of the nourishment sand was placed on eroding shorelines, while about half was placed on shorelines that accreted or were stable from about 1869 to 1971 but where encroachment by development had made the nourishment necessary (Houston and Dean 2015).

A specific comment needs to be made here on engineering practice in managing beach erosion. Engineering practice in some countries, notably the Netherlands, is now being conducted on the basis of a systems approach that *builds with Nature* (Deltares 2014), rather than simply in terms of deterministic engineering models, and this approach is now setting the trend in many areas of environmental management. Among the initiatives are massive nourishment (Brown et al. 2016), as embodied, for instance, in the “Sand Engine” on the Dutch coast (Stive et al. 2013), also referred to as the Zandmotor (ZM). This combined beach/shoreface mega-nourishment, built in 2011, consists of a total sediment volume of 21 million m³. A comprehensive multidisciplinary 5-year monitoring program was launched in 2011 to investigate various aspects of the initial ZM evolution and the coastal response (de Schipper et al. 2016; Luijendijk et al. 2017). However, beach management based on hard engineering is still very much entrenched in many countries and is leading to the demise of many beaches.

Summary

Beaches are a multi-resource asset in many ways, involving huge sums of money in developed economies, both in terms of revenue and for design and implementation of management policies. Beaches provide various important ecosystem services chief among which is that of protecting the coast by absorbing wave energy. Beaches also help in nutrient cycling and water purification, provide nursery habitats for resource

species and feeding-breeding habitats for various species of shorebirds and endangered sea turtles, while also serving as important nodes for economic development and cultural use. Beach erosion is a process whereby a beach loses its sediment, resulting in a depletion of its sediment budget. Beach erosion can therefore lead to a loss of valuable multi-resource assets and coastal ecosystem services, notably that of protection against waves. The complexity of beach erosion and the large number of parties involved in its management should call for a sensible and balanced mix of science with a systems approach, engineering expertise, past and present experiences, and the specificities of the local context in which erosion occurs.

Many developed countries are today faced with minor to critical beach erosion problems, largely because of lack of foresight in coastal development patterns. While they may have the resources to combat beach erosion, the same is not true for emerging economies which cannot divert much needed money toward beach management, often considered as a “low priority” area. These countries are increasingly subject to the pressures of often rapid economic development and of beach-based tourist activities (e.g., El Mrini et al. 2012), whereas beaches are particularly vulnerable in the face of the threat of sea-level rise from global warming (Masselink and Gehrels 2014; Jimenez et al. 2017). It is perhaps reassuring that because of the still moderate level of coastal development in many of these countries, they could still have the opportunity of avoiding the mistakes made in the past by the developed countries by planning such development in a way as to ensure sustenance of the beach resource.

Cross-References

- [Beach Nourishment](#)
- [Beach Processes](#)
- [Changing Sea Levels](#)

- Dissipative Beaches
- Drift and Swash Alignments
- Gravel Barriers
- Modes and Patterns of Shoreline Change
- Reflective Beaches
- Sea-Level and Climate Change
- Shoreline and Coastal Terrain Mapping
- Shoreline Response to Littoral Drift Barriers
- Shore Protection Structures
- Surf Zone Processes

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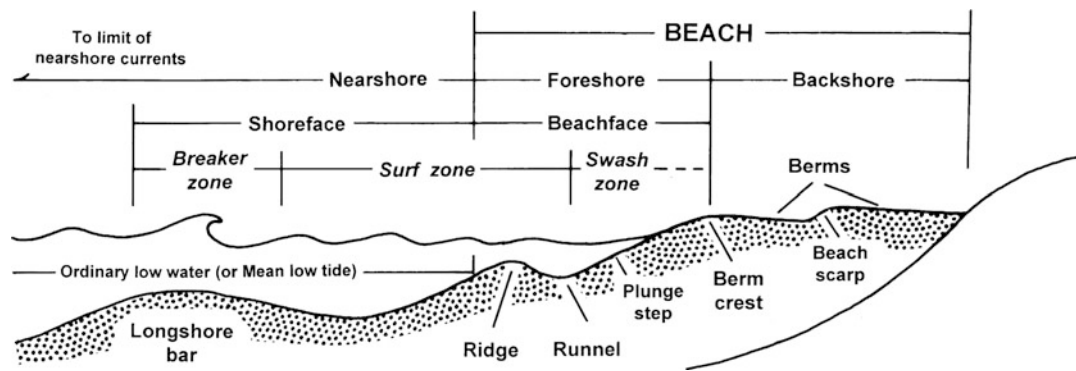
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Beach Features

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Limits and Formation of Beach Features

When discussing the types of features that can be observed along a beach, it is important to first consider the boundaries in the coastal zone that define the limits of a beach. In everyday usage and in the scientific literature, there are some differences in defining these limits, primarily with regard to the seaward limit. Recreational beach users will often consider the beach to extend no farther seaward than the shoreline, and thus limit the beach to an entirely emergent



Beach Features, Fig. 1 Generalized beach and nearshore profile showing names of major beach features and zones

feature, having a width that varies with changing water level. Scientific usage typically extends the beach out to the maximum limit of low water regardless of the water level at any particular time. In some scientific usage, such as in the discussion of coastal sediment dynamics, the seaward limit of the beach may be considered to extend out to the breaker zone (Fig. 1) well beyond the low-water shoreline. The most useful definition, and the one used here, is that the beach refers to the zone containing unconsolidated material that extends from the limit of ordinary low-water (or mean low-tide level) on its seaward side to the limit of influence by storm waves on its landward side (Fig. 1) (Hunt and Groves 1965; Baker et al. 1966; Coastal Engineering Research Center 1984).

Based on morphology, the beach is divisible into two zones. The *backshore* is the more landward and higher part of the beach and is typically a near-horizontal to gently landward-sloping surface. The backshore is not affected by the run-up of waves except during storm events, and so this is the typically dry part of the beach. The landward limit of the beach, which is the limit of influence of storm waves, generally is marked by a change in material, a change in morphology, or a change to a zone of permanent vegetation. Examples of such a landward limit include dunes, cliffs or bluffs, or even engineered structures such as bulkheads or revetments. The *foreshore*, also called the *beachface*, is the more seaward part of the beach. The foreshore has an overall seaward slope, but may include one or more ridges and troughs on its lower slope. Because the foreshore extends to the limit of ordinary low-water, at times of high-water the lower part of the foreshore is submerged.

Critical in the definition of a beach is the presence of unconsolidated materials. These unconsolidated materials are what make a beach, and it is the erosion, transport, and deposition of these materials that results in beach features. Worldwide, the most common beach material is sand-size sediments composed of mineral, shell, or rock fragments. Coarser beach materials include gravel, cobbles, shingle, and even boulders. The beach will be made from whatever

is locally available for the waves to rework. Along shores impacted by commercial or industrial activity, it is not uncommon to find beaches composed in part or completely of bricks, broken concrete, demolition debris, or any other material that may have been dumped along the shore and subjected to movement and redistribution by wave action.

The types of features that may occur along and across a beach vary in time, scale, and relative position. The primary agent in forming beach features is wave action (Davis 1985). Other important agents are the rise and fall of water level, currents, wind, and ice in settings where coastal ice can form. Some beach features can form and persist indefinitely with minimal change in shape or location, but these are the exception. Because the beach is a dynamic setting, most beach features are ephemeral. Once formed, most beach features will only persist until new wave, water level, current, or wind conditions destroy them and replace them with new features.

Beach Slope

Beach slope, which refers to the slope of the foreshore or beachface, deserves special mention as a beach feature because it is one of the characteristics used to distinguish different beaches. Slope is a dynamic feature that changes with changes in wave conditions as well as the gain or loss of different sediment sizes on the beach face. In general, the slope angle, measured relative to a horizontal plane, increases as the grain size increases; thus beaches composed of material such as pebbles or cobbles will tend to have a steeper beach face than ones made of sand. This slope difference relates to the greater permeability of the larger materials (Bagnold 1940; Bascom 1951; King 1972). The wave run-up (or swash) can percolate downward through the interstitial spaces of the larger materials, and this minimizes the erosional influence of the run-back (or backwash). Storm conditions will flatten the beach slope as beach sediment is eroded and transported

seaward. In the calmer wave conditions following the storm, beach slope recovers to a steeper slope as material is accreted to the beach.

Major Beach Features

Major beach features are here defined as those having large topographic expression or large areal extent. One of the beach features that can have the largest vertical expression on a beach is a *beach scarp*. Beach scarps are erosional features that occur when the slope of the beachface is lowered during storm events, and the beachface migrates landward by cutting into the backshore. The result is a near-vertical slope along the limit of this erosion (Fig. 2). The height of a beach scarp may be just a few centimeters or a meter or more depending on the degree of wave action and the type of beach material. Beach scarps are commonly observed in areas where a new supply of sediment (i.e., beach nourishment) has recently been applied in an effort to replenish and build up the beach and wave action has cut into this nourishment and redistributed the sediment in the process of reestablishing an equilibrium beach profile.

One of the reasons that beach scarps are prominent beach features is that these near-vertical erosional features are cut into a near-horizontal area of the upper beach. This upper beach, in many cases, is a broad, near-horizontal to gently landward-sloping area called a *beach berm*, or simply a *berm* (Fig. 1). Berms are depositional features formed from the wave-induced onshore accumulation of sediment. Local coastal conditions may preclude formation of a berm along some beach segments, while other beach segments may have two or more berms at different elevations. When more than one berm occurs, the lower berm(s) (sometimes called the

ordinary berm) is a result of average or more typical waves, and the higher berm(s) (sometimes called the *storm berm*) is a result of the less frequent larger waves. A beach scarp may exist between two berms having different elevations. The seaward margin of the berm is typically defined by a rather abrupt change in slope from the near horizontal surface of the berm to the inclined surface of the beachface. The line defined by this change in slope is called the *berm crest* or *berm edge*. The berm crest is the distinguishing beach feature that divides the beach into the foreshore and backshore zones (Fig. 1).

When low-water occurs, large-scale beach features are exposed that formed underwater and have a morphology influenced by waves, water-level changes, and associated currents. *Ridges and runnels* are the most common of these features. Ridges are elongate low mounds of beach material that are parallel or subparallel to the shore; runnels are the low areas or troughs that occur between the ridges and on the landward side of the shoremost ridge. A single ridge–runnel set may occur with the runnel on the landward side of the ridge or, if the lower foreshore is a broad, low-slope area, multiple sets of ridges and runnels may extend across this area. Such multiple sets of ridges and runnels form a washboard or corrugated topography across the lower beachface that contrasts with the smoother surface across the upper beachface. Another term for these features is “*ball*” referring to the ridge, and “*low*” referring to the runnel. The term “*trough*” is also sometimes applied to the runnel.

Ridges and runnels, when present in multiple sets, are one example of the types of repetitive patterns that can be observed in beach features. Another major beach feature with a repetitive nature is the *beach cusp*. Beach cusps are low mounds of beach material, separated by crescent-shaped troughs, occurring in a series along the shore. Yet another repetitive feature, and one of potentially large vertical scale, is the *beach ridge*. Beach ridges are depositional features, formed by the mounding of beach material by wave action, usually during storm events. Beach ridges are formed in the backshore zone, in some places at the most landward limit of wave influence, and they can extend as a nearly continuous linear feature for many kilometers along the shore. Wind transport may contribute sand to the tops of these ridges and form superimposed dunes (i.e., dune ridges). A single beach ridge may develop, persist for some time, and then be destroyed by renewed storm action. Sequential beach ridges may form through a series of depositional events and, with time, the juxtaposition of these ridges will contribute to the progradation of the coast.

A major beach feature common to barrier island beaches is the *washover fan*. During storm events, elevated water levels and large wave run-up can combine to transport large volumes of beach material across the beach to be deposited in broad, lobate accumulations on the landward margin of



Beach Features, Fig. 2 Beach scarp resulting from recent storm erosion along a sand beach on the Illinois shore of Lake Michigan at Illinois Beach State Park. (Photo by Michael Chrzastowski, Illinois State Geological Survey.)

the beach. These deposits, called *washover fans*, essentially extend the landward limit of the beach and play an important role in the landward and upward migration of the beach during conditions of rising sea level (Kraft and Chrastowski 1985). On barrier island beaches, the formation of washover fans results in the burial of back-barrier marshes and filling along the barrier margin of lagoons.

Coasts that are subject to seasonal ice formation, such as the Great Lakes coasts of North America, can have various beach features that are related to the presence of coastal ice. A hummocky topography may develop along the upper foreshore as a result of sediment pushed into ridges and mounds by wave-thrusted ice. Shallow depressions may also develop across the upper foreshore where wave run-up may remove thin slabs of ice-cemented sand. Once an ice complex forms along the shore, a hummocky topography may develop in the lower foreshore by the action of grounded ice and the scour and fill by waves and currents around the edges of the ice. On the outer edge of the ice, an erosional trough may develop caused by the downward deflection of wave energy as waves impact the ice face. This trough can be a half-meter deep and 2–3 m wide (Barnes et al. 1994). The trough location will shift toward or away from shore as the ice margin shifts in these directions. These troughs and all other ice-related beach features are relatively short-lived. Once the ice conditions cease, any ice-induced modifications to the beach morphology are quickly eliminated by ice-free wave conditions.

Minor Beach Features

Minor beach features are defined here as those with minimal topographic expression or small areal extent. Although limited in height and area, some of these beach features can be visually prominent because of contrast in color, texture, or materials compared to the surrounding beach. Prime examples of such prominent, small-scale beach features are the *tidemark* which is the high-water mark left by tidal water, and *swash marks* formed along the landward limit of wave swash on the beach face. The tide mark is generally a nearly continuous wavy line defined by an accumulation of driftwood, seaweed, and other floatable debris collectively called *flotsam*, left on the beach by the previous highest tide level. Swash marks are a series of superimposed scalloped or fan-shaped patterns defined by fine sand, mica flakes, or bits of seaweed deposited along the most landward reach of the swash. Swash marks are beach features that are in a nearly continuous state of formation and destruction with each new swash event. So too are *backwash patterns* which are diagonal patterns formed on the beach by the dispersion of backwash flowing around small obstacles such as a shell or pebble. A falling tide or the lowering of water level after a storm can contribute to the formation of *rill marks* which are small,

erosional furrows or channels across the beachface caused by the seaward flow of water as the water table in the beach lowers and water percolates out onto the beach-face in a spring-like manner. *Air holes* may also occur on the beachface as water percolates and forces air from the pore spaces up to the surface.

Near the ridge and runnel on the lower foreshore slope, a subtle linear feature may occur called the *step* or *plunge step*. This is a subtle decline in the foreshore profile that is caused by the final plunge of waves before running up the beachface. The plunge step is best developed in settings of low tidal range and steep foreshore slope (Davis 1985).

Because one or more of the berms of a beach are elevated above the influence of the swash, fine sand across these berms is typically dry and can be influenced by wind action. Although the berms are located on the beach, features can develop here that are common to dunes and desert settings. Sand ripples may develop, small dunes may form in wind shadows such as behind logs or other driftwood, and wind deflation areas may occur where the fine sand has been removed to lower the surface and leave a concentration of coarser particles similar to a desert pavement.

Cross-References

- [Beach Nourishment](#)
- [Beach Processes](#)
- [Beach Profile](#)
- [Beach Ridges](#)
- [Drift and Swash Alignments](#)
- [Rhythmic Patterns](#)
- [Ripple Marks](#)
- [Scour and Burial of Objects in Shallow Water](#)

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Beach Management Tools

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Definition

In recognition of the beach as a complex socio-ecological system, a beach management tool is any method or technique whose application is itself a mechanism of decision-making affecting the beach dynamic. This category includes (but is not necessarily limited to) the following items: beach classification, beach user density, beach carrying capacity, beach zoning, beach litter cleaning, beach environmental quality, beach management bodies, and beach awards and certifications.

Instrumentation of Management

The maturity of beach management as a set of concepts and methodologies was established in 2009, when Allan T. Williams and Anton Micallef published their book about the topic in question (See “Beach Management”). As they mention there, “*Current management is essentially practical (...)*” (Williams and Micallef 2009: 44), which infers that some minimum instrumentation is required. Nevertheless, there are very few proposals of tools designed specifically for beach management as a socio-ecological system and almost none from the complexity approach. In consequence, the first step should be to differentiate “*tools for beach management*” from “*beach management tools*” (BMT). The former are all techniques, methods, and equipment used to generate useful and worthy information for managing one or more beach aspects in the sense that their *outcomes* will be used as *incomes* of management actions on a particular beach. Examples of these kinds of tools range from geographical information systems to erosion monitoring, user’s perception, and ecological inventories, among others. Referring to the latter, although they are also methods and techniques, their mere use is as a management action itself; in other words, their function is not to generate

information for supporting management unless it is to make decisions directly over the beach dynamic. This epistemological difference is the milestone on which this entry is based.

With reference to Williams and Micallef (2009), they propose to use three tools for beach management: dimension analysis, function analysis, and environmental risk assessment. The first is an evaluation technique which has not been developed for beach management, but according to the authors, it is very effective for scoping and extremely adaptable in obtaining essential facts for problem definition. The second is useful for considering the environmental goods and services by addressing the characteristics of that environment; this tool analyzes four types of functions: production functions, carrier functions, information functions, and regulation functions. Finally, the third is an assessment process that allows managers to evaluate any proposed development case-by-case and quantify its impacts, with the aim of providing a basis for making decisions about the significant effects of the project. In addition, Williams and Micallef (2009) propose a beach management tool developed in its entirety by them: the Bathing Area Registration and Evaluation (BARE). This innovative tool starts with a detailed registration of the bathing area, considering it as the beach or rocky platform and its surrounding area (~500 m). Afterwards, the five main issues of concern to beach users, the so-called *Big Five*, are evaluated and rated on a four-scale system (A-D), with support of several formats and tables. Lastly, each issue is analyzed using a classification of 1–5 stars, which allows the identification of poorly rated quality criteria and suggests improvements to upgrade the bathing area classification.

With the exception of this proposal, there are few sets of beach management tools available. Perhaps, the most recent compendium is the book titled *Beach Management Tools* (Botero et al. 2018), which seeks to highlight concepts, methods, and case studies world-wide. Notwithstanding, the book has 48 chapters, which prove that the majority of decision-making on beaches is already done by *tools for management*, not by *beach management tools*. As a matter of fact, in the book only four chapters consider innovative tools, while 37 of them are examples of several tools applied to beach management. Last, but not least, there are seven groundbreaking chapters that contextualize each of the seven parts of the book. To sum up, *beach management tools* is a topic which is growing rapidly and in a disorderly manner; therefore, this entry seeks to summarize efforts made by a myriad of researchers around the world and to support a clear taxonomy of the beach management tools.

Beach User Density

The number of people on the beach at a certain time and in a certain space is commonly recognized as Beach User Density

(BUD), and its measuring unit is square meters per user (m^2/user). The vast majority of papers about quantity of beach users are related to carrying capacity (See ► [“Carrying Capacity in Coastal Areas”](#)), but papers about BUD are less frequent, although some values of BUD could be found. Nonetheless, information about the number people on the beach and where they are is extremely important to define beach management strategies and actions. In a paper written by a group of European researchers (Jiménez et al. 2007), an analysis of the temporal and spatial pattern of user distribution in a Spanish beach was done, and the usefulness of counting users was amply demonstrated. As a general rule, BUD is the main *status variable* to manage tourist beaches, because beachgoers are the focus of management actions.

BUD is a tool with several applications for managers. First of all, it allows them to identify quantity and spatial distributions of users, which is the basis to establish some actions such as: number and location of accesses, toilets and signage, potential risk areas, channels of vessels, among others. Secondly, the monitoring of BUD brings worthy information about the duration of tourist seasons and beach zones preferred by users; indeed, this data is highly necessary in tropical beaches where the affluence of tourists is not defined by the summer season. Similarly, the daily and hourly changes of BUD help beach managers make better decisions about several issues: the schedule of lifeguard service and sand cleaning, the prevention of risks at peak times, and the reduction of conflicts among different types of users (e.g., families versus surfers). In conclusion, recording very simple data, such as the number and distribution of beachgoers, could support the majority of management measures on the beach.

In contrast to the importance of counting users, techniques to measure BUD are not well developed. Different methods could be discovered within scientific literature, from very simple hand counting to high-tech equipment. Initially, one of the simplest methods, as described by Pereira (2015), is a set of 20-m strips which are placed every 200 m along the beach, and, later, users are counted by hand by two people; this procedure is repeated every 2 h during the day and finally registered into a spreadsheet (e.g., OpenCalc). Other techniques are a mix between data acquired by specific equipment (e.g., drones, airplanes) and hand counting. A common example are the aerial images in which beachgoers are counted one by one and registered into a database; Pereira (2014) made a description of several geographical techniques used for measuring recreational parameters on beaches. Another semi-manual method is recently used in Portugal, where users are counted by an infrared sensor located at the main access on the beach; in this case, each person who passes the access is summed by the sensor and recorded in a database. Finally, totally automated mechanisms have been developed in recent years, such as those described by Jiménez et al. (2007) in which photographs are taken periodically during the day,

making use of video cameras installed directly on the beach, and later sent via the Internet to a computer which automatically counts users through an algorithm based on the pixels of the image. In summary, BUD can be measured by several methods according to the specific budget and capabilities of each beach manager; however, regardless of the technique used, information about the number and distribution of beachgoers are crucial for pertinent decision-making on beaches.

Beach Zoning

Coastal zoning is a concept frequently used to define sectors of coastal areas with a spatial scale of tens to hundreds of kilometers (1:50,000), which is too broad for beach management. Beaches are commonly strips of sand (longer than they are wide) in which several human activities are done in hundreds or, maximum, thousands of meters (1:1000), meaning a geographical view two or three degrees smaller than coastal zoning. Within this competitive space, several conflicts between users and uses emerge as a consequence of the simultaneous activities taking place. Therefore, organizing uses on the beach into specific zones helps managers to get better use of this desired space. Even though these zones could have any shape and size, a definition of the spatial and temporal pattern of beach users is recommended as the first criterion for zoning. Moreover, this BMT is strongly linked to beach user density, because more beachgoers imply more potential conflicts among them and is commonly related to their territorial behaviors and tolerance to proximity (Silva 2002).

Zoning has several applications as a beach management tool. First, conflicts among uses such as fishery, nautical sports, sunbathing, sea bathing, and beach games are reduced because each one will have their own exclusive area. At the same time, conflicts among users are also prevented with implementation of zones for each kind of activity or user group (e.g., families, surfers, resorts). Zoning is also useful to define the location of facilities such as toilets, lifeguard towers, or shading implements (e.g., umbrellas, tents) according to the spatial dimensions of the beach and BUD patterns. Ecosystem protection is also facilitated by zoning because the sensitive natural areas of the beach (e.g., seagrass meadows, coral reefs, dunes) could be demarcated, preventing the presence of uses and infrastructure that affect their health. Finally, zoning is useful to prevent risks on the beach and improve safety, through the definition of dangerous areas forbidden for beachgoers or with a smaller carrying capacity.

Techniques to implement beach zoning should include three main criteria: space, time, and season. The first is related to the width and length of the beach, defining areas or strips with specific activities allowed within them (e.g., sunbathing,

swimming, boating). The second criterion allows managers to define zones according to temporal pattern of beach use, mainly during the day. As aforementioned in BUD, some types of beachgoers visit the beach at different times; therefore, zoning should be done in order to allow everyone to enjoy the beach without interrupting the enjoyment of others. Lastly, the third criterion is linked to the seasonality of tourism and its impact on activities done year-round on the beach, such as fishery or recreation (See ► “[Tourist Beaches](#)”). Tourist seasonality depends on climate characteristics (e.g., tropical versus tempered countries) and cultural traditions (e.g., Easter Week, school holidays); consequently, zoning could change according to BUD or specific events. In conclusion, beach zoning is an exercise to organize space and time, starting with a general framework and finishing with a very specific design.

A relevant reference about zoning is a document published half a century ago by the Spanish Minister of Public Works (MOP 1970), which includes the majority of concepts about zoning. Detailed figures recommending zoning for Atlantic and Mediterranean beaches and examples of beaches with reserved areas for specific activities (such as fishery) are presented. Additionally, this book presents criteria to ensure comfort on each zone of the beach with diagrams of space occupied by different kinds of umbrellas and to establish the importance of the “*facilities zone*.” Another reference of spatial zoning is done in Colombia, where eight strips parallel to the coastline are defined by Decree 1766/2013 (MCIT 2013), from land to the sea: (a) facilities zone, (b) public space linked zone, (c) transition zone, (d) resting zone, (e) active zone, (f) swimming zone, (g) boat access zone, and (h) nautical sports zone. Application of this zoning establishes a general model for the whole country and a guideline about priority zones on narrow beaches. Although this zoning is already done on several beaches in Colombia, its implementation is respected in some places (Fig. 1) and, in others, it is not (Fig. 2).

As a whole, zoning is a technical procedure made with temporal patterns and the geographical information of facilities, ecosystems, and activities on the beach. Notwithstanding, the participation of stakeholders is a key factor to ensure real implementation, mainly by those who worked directly in the sand and water, such as owners of umbrellas and recreational boats. An example worthy of mentioning was done in Santa Marta (Colombia), on a beach called Playa Blanca, where stakeholders and the maritime authority established parallel and perpendicular zoning of the beach in a totally bottom-up process that allowed them to keep this zoning for over 10 years without major conflicts (Fig. 3).

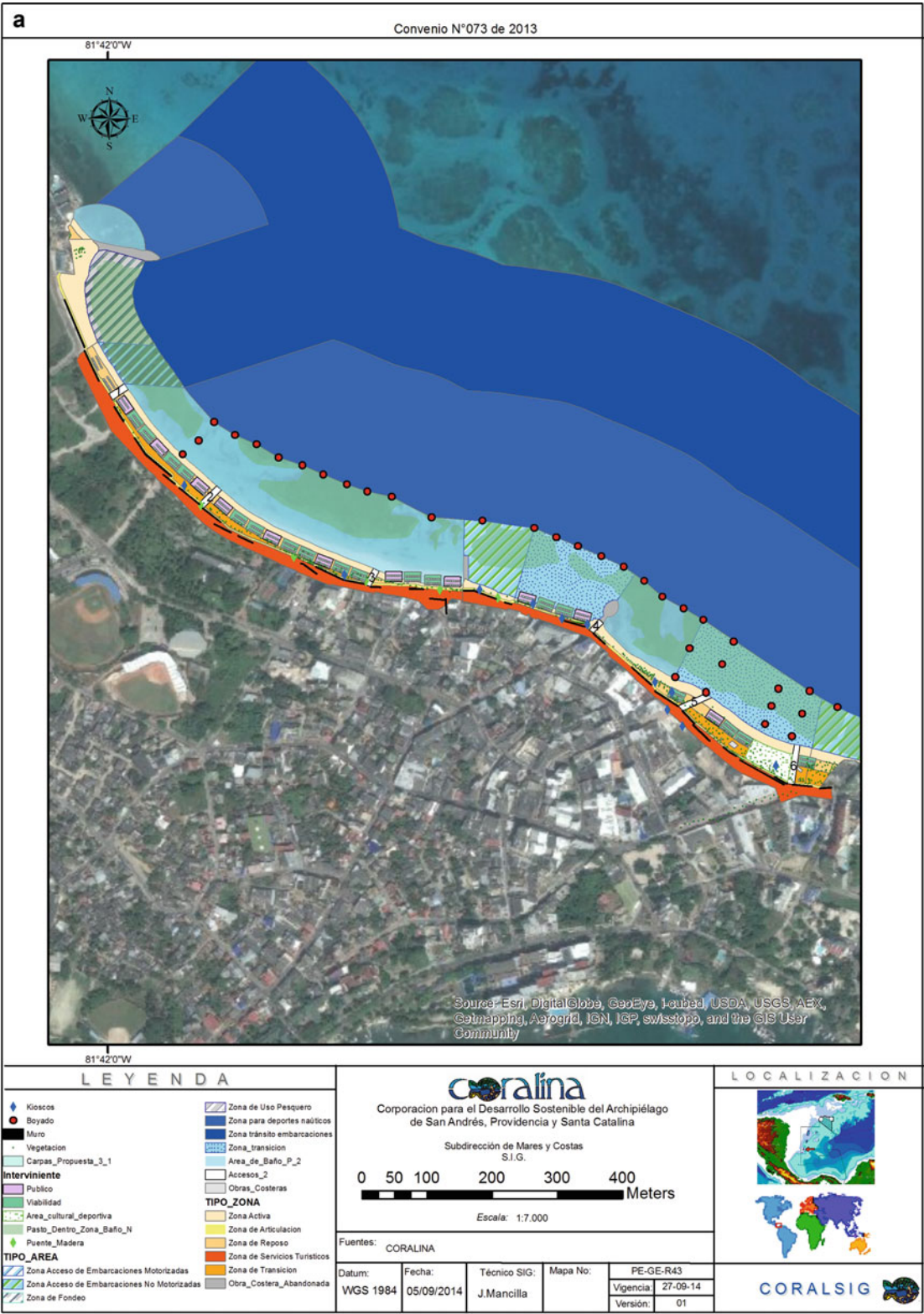
Beach Management Body

Beaches should be managed in a geographical scale smaller than the municipal area: a microlocal scale. Almost all

approaches of Integrated Coastal Zone Management (ICZM) include the local scale as the minimum planning level, which forces coastal managers to make the same decisions for all beaches within their jurisdiction. However, decision-making at beach level is done by stakeholders directly related to the human uses and economic activities on the beach: public service entities (e.g., sand cleaning, ecosystem conservation, lifeguards), tourism brokers (e.g., restaurants, umbrellas, motorcrafts), and beachgoers (e.g., tourists, local visitors). Despite beach management usually being part of ICZM programs, decision-making at beach scale should be a process of microlocal governance, supported by a bottom-up approach and self-regulation mechanisms. Successful beach management requires key stakeholders to do the decision-making and to be organized in a single body which proposes and implements its own decisions (Botero 2013). Furthermore, as Herrera (2010) demonstrated, this beach management body (BMB) needs strong support from local authorities (albeit with a minimum amount of interference) in order to reach an institutional legitimacy for its decisions. A general model of a BMB is shown by Fig. 4, in which blue elements represent the core of the BMB and key stakeholders, while green elements are the local institutions according to their level of intervention in the BMB decisions.

The main functions of a BMB are related to the implementation of daily actions within a tactical approach. The first task should be the short-term planning to reach and maintain good environmental status and tourist quality within an ecosystem services approach. A BMB should have the capacity to plan and adjust its planning in a quick and pertinent loop, creating an adaptive management system in periods of weeks or months. Secondly, a BMB should be the first entity to resolve conflicts among stakeholders interacting on the beach. Prior to involving an active legal or economic mechanism, stakeholders should try to resolve their conflicts through the BMB to which they belong, reducing external interference of institutions that do not have a complete knowledge of beach-system reality. The third function is related to the direct implementation of its decisions, which implies planning, financing, and evaluating each action done on the beach. Nowadays, the majority of actions applied on beaches are decided at the local scale in a classic example of the top-down approach, reducing the interest of microlocal stakeholders in their application and success. Finally, a fourth and more holistic function of BMB is its self-regulation, which includes approval of its own management rules, an administration budget for beach management actions and monitoring the status of the beach (e.g., environmental quality, beach users' perception). All four of these tactical functions should be immersed within a strategic plan of the beach, approved by the BMB, and in harmony with the ICZM plan of the municipality.

Clear examples of a BMB are still rare in the world. One of these few examples occurs on a beach aforementioned (Playa



Beach Management Tools, Fig. 1 (continued)



Beach Management Tools, Fig. 1 Beach zoning and implementation in Spratt Bight, San Andres Island (Colombia)

Blanca, Santa Marta, Colombia), where tourism brokers that work directly on the beach created an organization in 2008 and for 10 years have been overseeing almost every management action on the beach. This very interesting case was studied by Herrera (2010), who was at the same time the maritime authority and reinforced this bottom-up process from his position. Subsequently, other beaches have started similar BMBs formed by self-organized tourism brokers directly related to beach activities.

Conversely, the majority of BMBs are formed by law, and their members are mainly public institutions at the local level. Some examples were found in Latin America, where the PROPLAYAS NETWORK (www.proplayas.org) provided information directly from experts and managers of 16 countries and 42 working teams. The most successful type of municipal BMB is identified in Mexico, where the *Comites de Playas Limpas* (Clean Beaches Committees) have been working in more than 40 municipalities of the country since 2003. This kind of BMB is comprised of the mayor of the municipality, a delegate of the national authority of water (CONAGUA), representatives from public institutions at the federal, regional, and local levels, representatives of civil society, and beach users. Yearly, these committees have a meeting to exchange experiences among themselves and receive beach awards such as the international ecolabel “Blue Flag” and the national beach award “White Flag.” A more recent example of a municipal BMB is presented in Colombia, where Decree 1766/2013 created the *Comites Locales de Organizacion de Playas* (Local Committees for Beach Organization), which are formed by three institutions: the Ministry of Tourism, the Mayor’s Office, and the

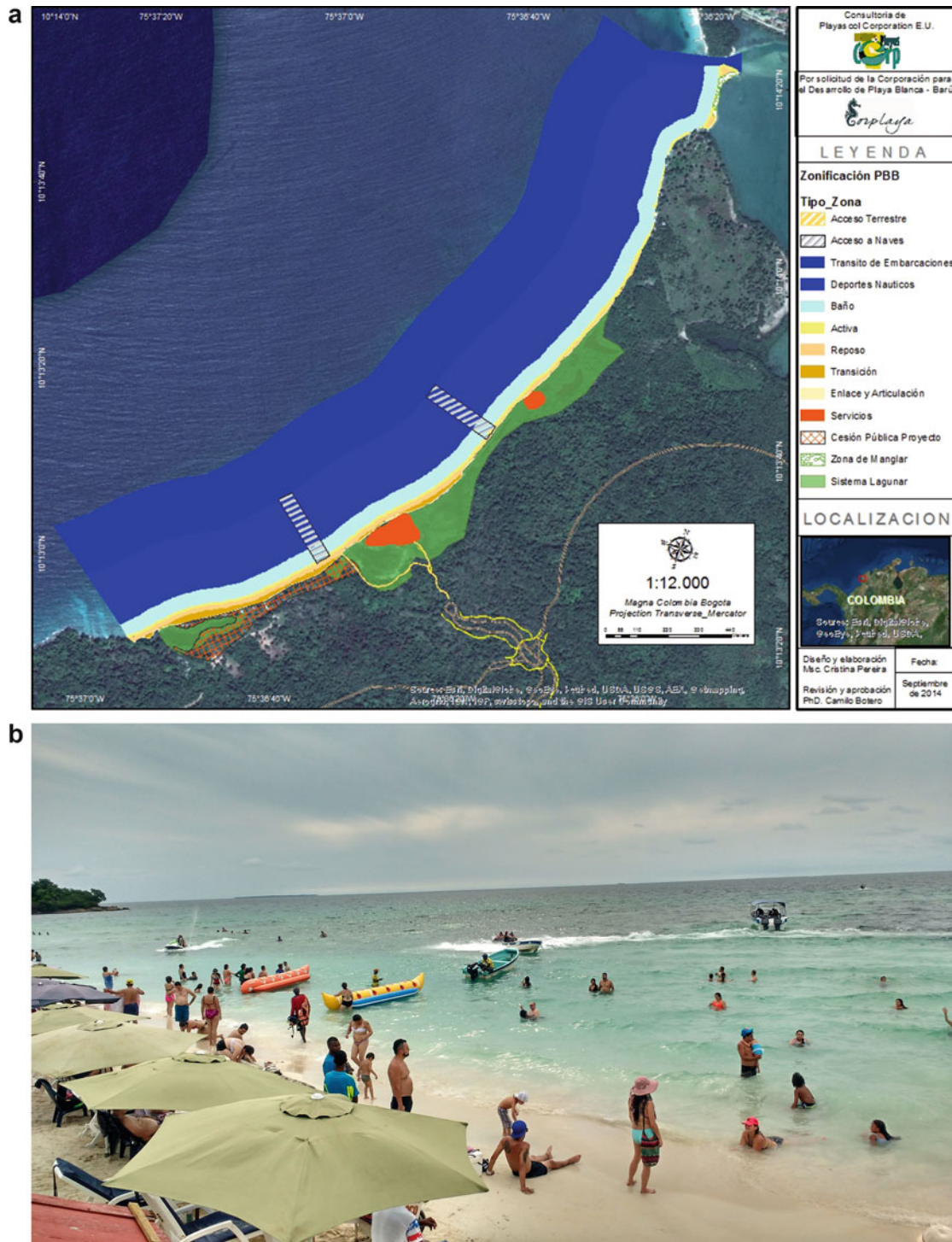
Maritime Authority, although any other institution, organization, or person could be invited to any given session. By 2017, 26 of these committees had been created in this country, covering 56.5% of its coastal municipalities. Their main actions have been to define zoning in principal tourist beaches of each municipality and resolve conflicts with private beach-front constructions, which is forbidden in Colombia. Some other examples could be mentioned, such as the *Junta de Playas* in Puerto Rico and the *Comité Participativo de Playas* in Uruguay, which have their own characteristics. In summary, BMBs are not common in the majority of countries, and those that exist have rarely been studied by scientific literature, although there is no doubt about their crucial role as a beach management tool.

Other Beach Management Tools

Equally important to the above-mentioned BMT, there are at least four others worth mentioning. Because these BMT have more scientific and more advanced examples of application for the world, they have their own entry within this third edition of the *Encyclopedia of Coastal Sciences*. Nevertheless, a brief overview of each one will be provided with the goal of highlighting their capability as a BMT and as an invitation to read these other entries in a common framework of analysis.

Beach Classification

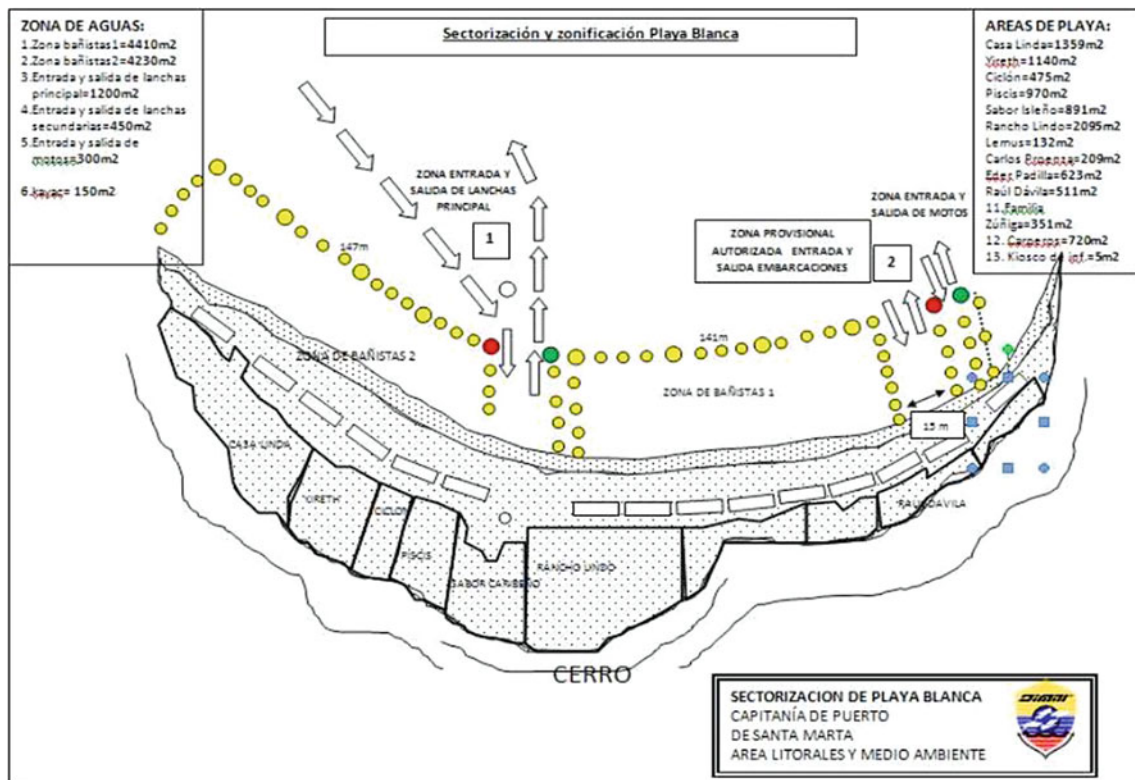
Every beach in the world is unique and at the same time, similar to many others. Stemming from this controversial



Beach Management Tools, Fig. 2 Beach zoning and implementation in Playa Blanca Baru, Cartagena (Colombia)

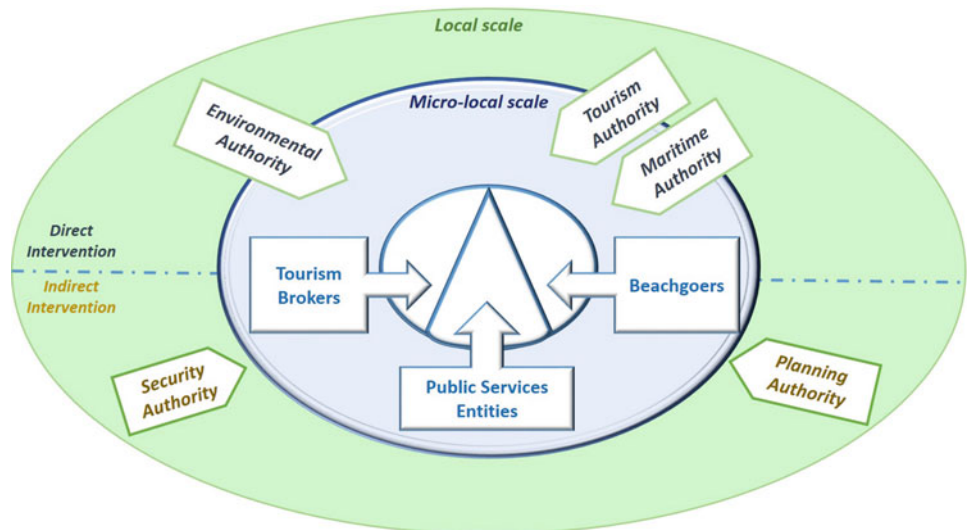
reality, a tool to differentiate one beach from another without creating a vast taxonomy is necessary and almost compulsory. The classification of beaches is a frequent topic in scientific literature, such as Williams and Micallef (2009) highlighted, although a universal list of beach typologies does not exist. The majority of beach classification is from specialists in

natural sciences with emphasis on geomorphology or biology; however, such classifications are of little use to socio-natural systems, such as the vast majority of beaches are. A classification is worthy of management if it is able to integrate ecosystem characteristics with human uses and institutional frameworks, allowing some decisions and restricting others.



Beach Management Tools, Fig. 3 Beach zoning made by stakeholders in Playa Blanca, Santa Marta (Colombia) (Herrera 2010)

Beach Management Tools, Fig. 4 Beach management body – general model



Williams and Micallef (2009) give a clear example when they propose five types of beaches (urban, village, rural, remote, and resort) and define specific management actions for each one. Additionally, almost all BMT will be affected by the beach classification chosen because tools such as BUD, carrying capacity, litter cleaning, or zoning will depend directly of the kind of beach in which it will be applied. In short, beach

classification should be the first BMT applied in any coastal area, when a beach management process is selected (See “Beach Management”).

Beach Carrying Capacity

One of the most popular BMT is the carrying capacity. This concept was first applied for tourism purposes by Cifuentes

(1992), in the Galapagos Islands (Ecuador). Since this study, tens of papers have been published with study cases all around the world, Da Silva (2002) being the most quoted about this tool applied on beaches. In general, beach carrying capacity (BCC) is a tool which seeks a balance between ecosystem resilience and tourist use while keeping the user's comfort at its core interest. BCC is closely related to BUD, because both have the same amount of users as a primary output, although their interpretation is almost the inverse: the latter is the number of users in a certain time and space, the former is the maximum limit of this quantity. Despite the popularity of BCC, several methods have been proposed and none have yet to be accepted by the majority of the scientific community; nevertheless, Cifuentes (1992) has been the most influential reference. The usefulness of BCC is divided into two sides: ecological and tourism. The first is related to the capacity of the beach as an ecosystem to support the development of human activities, such as fishery, tourism, recreation, dumping, among others. Based on this perspective, BCC should be one of the main criteria to decide how, where, and when a certain human action should be done in order to assure the beach ecosystem will resist and recover its natural functionality. The second side is about tourism factors, which has at least three corners: user perception, facility planning, and environmental attitude. User perception is related to the beachgoers' comfort and the quality of the experience during their stay; the better the perception, the higher the BCC. Facilities on the beach are one of the Big Five established by Williams and Micallef (2009) and require careful planning in order to reduce their environmental impact, which is a complex function of quantity of facilities, quality of materials, location, environmental management and maintenance; the better the planning, the higher the BCC. Thirdly, the environmental attitude of beachgoers is the most direct guarantee of better behaviors and less impact on the ecosystem and other users' comfort; the more positive the environmental attitude, the higher the BCC. As it can be seen, BCC is a complex and complete tool, with many references to apply and innovate (see "Beach Carrying Capacity").

Beach Litter Cleaning

The absence of litter on the sand is another point of the Big Five. Hundreds of papers have been published about beach litter, the majority of which have been about the quantity and characteristics of items trashed on the sand. Conversely, research about the sources of these items of litter is rarely and with more difficulty analyzed than counted and weighed. To illustrate this BMT, potential risk to the health of beachgoers is one of the most important effects of beach litter due to exposure to harmful items such as broken glass, syringes, and feces. Another very important item of litter for beach cleaning are those smaller than 5-cm long, such as cigarette butts and

bottle taps, which are difficult to clean by manual methods. As a result, nowadays several machines to clean beaches exist, from large equipment for cleaning more than 26,000 m²/h to small machines with their own engine, with rates between 1250 to 2500 m²/h. However, this mechanical cleaning should be used in certain types of beaches, or areas of the beach, without high ecosystem values, such as dunes or nesting turtles. Despite the perception of these machines having a high environmental impact, scientific evidence of this damage is very scarce and those existing demonstrate the great resilience of the beach ecosystem (Gheskiere et al. 2006). Furthermore, technology in beach cleaning is improving rapidly, as demonstrated in the Caribbean, where the invasion of an algae (*sargassum*) had covered hundreds of beaches since 2013, and nowadays, some machines have been designed for its cleaning. Perhaps, the only litter items not yet possible to clean with machines are the vegetal residues carried away by big rivers, which is a difficult challenge in tropical areas where these inlets transport tons of wood and garbage from river basins and finish directly on the sand (Fig. 5); the cleaning of these items requires large machines and a method that prevents sand from escaping off the beach by bulldozing. To conclude, beach litter cleaning is a compulsory BMT on any tourist beach, although a better method of doing so is by making the decision according to each particular beach (See ► "Marine Litter" and "Cleaning Beaches").

Beach Awards and Certifications

On top of beach management tools are the beach awards and certifications (BA&C). Generally, BA&C are acknowledgments given to a certain beach when a high standard of management is attained, usually through the fulfillment of specific criteria. The first and most known certification is the *Blue Flag Program* (www.blueflag.global), an ecolabel created and administered by the *Foundation for Environmental Education* since 1985. This award is applied in almost 50 countries around the world, with more than 4000 beaches awarded in 2017. Furthermore, many other BA&Cs have been created in Europe and Latin America with either a national or regional scope, and a limited application; Botero (2013) identified more than 40 BA&Cs around the world, and Botero et al. (2014) describes the 9 BA&Cs existing in Latin America. Despite their popularity, the BA&C have several critics and aspects to improve, such as several authors have highlighted during more than 15 years (Nelson et al. 2000; Zielinski and Botero 2015). Nevertheless, BA&C are the most complete BMT because they cover every aspect related to management of a beach, have criteria to assess advancement, and publicly promote this effort to the tourism sector and beachgoers. Moreover, the BA&C could be an award for reaching good management, but they also trigger better management in municipalities with low institutional governance

Beach Management Tools,
Fig. 5 Vegetal debris carried
 away by big rivers in Colombia (**a**
 Salgar Beach, affected by
 Magdalena River; **b** El Totumo
 Beach, affected by Atrato River)



due to politicians who are easily tempted by the public recognition given to awarded beaches. Even though the coastal scientific community has not reached a consensus about the best way to implement the BA&C, their popularity is unquestionable, as the yearly increase in Blue Flag interest has demonstrated; the challenge remaining shall be how to reinforce the bridge of this popular BMT with the scientific community and the key stakeholders on the beach (See ► [“Beach Awards and Certifications”](#)).

Summary

Beach management tools are any method or technique whose application is itself a decision-making mechanism

affecting the socio-natural dynamic. This entry starts with a discussion about management instrumentation and what the differences between “tools for beach management” and “beach management tools” are; in addition, some examples of sets of BMT are mentioned. Later, three BMT are defined in depth, explaining their usefulness, state-of-the-art features, evolution, and some examples: beach user density, beach zoning, and beach management bodies. Likewise, another four BMT are explained briefly, because they are already in one or more entries of this encyclopedia: beach classification, beach carrying capacity, beach litter cleaning, and beach awards and certifications. This entry is not exhaustive about any possible BMT existing, but it covers some of the most common tools used nowadays on the beaches worldwide.

Cross-References

- [Beach Awards and Certifications](#)
- [Beach and Nearshore Instrumentation](#)
- [Beach Use and Behaviors](#)
- [Carrying Capacity in Coastal Areas](#)
- [Coastal Scenery](#)
- [Lifesaving and Beach Safety](#)
- [Rating Beaches](#)
- [Rip Currents](#)
- [Tourist Beaches](#)

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Beach Nourishment

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Introduction

Beaches occur where there is sufficient sediment for wave deposition above water level along lakes, open ocean coasts, embayments, and estuaries. Beach nourishment most commonly takes place along marine beaches, which are among the most dynamic environments on earth. On a global scale, estimates of marine sandy beaches (see entry on ► [Sandy Coasts](#)) range from about 34% (170,000 km) (Hardisty 1990) to 40% of the world's coastline (Bird 1996). Beaches form essentially 100% of the coast of The Netherlands, 60% in Australia, and 33% in the United States (Short 1999). Comprising a significant proportion of the world's coastline, beaches are important considerations for coastal recreation and storm protection, while others are used for residential, commercial, and industrial purposes. Although they serve as natural barriers to storm surge (*q.v.*) and waves (*q.v.*), today about 75% of the world's beaches are subject to erosion (Bird 1985). In the United States, the percentage of eroded beach-front is somewhat greater than the world average and is estimated by some coastal researchers to approach 90% (e.g., Leatherman 1988). During the last century, many erosion-control techniques were developed to mitigate the unwanted impacts of erosional events, especially those associated with accelerated rates of erosion in the vicinity of groins, seawalls, or jetties along developed shores. Traditionally, coastal armoring structures such as seawalls, breakwaters, and groins were relied upon to reduce wave energy approaching the shore or to catch sediment moving across or along the shore, and thus provide protection from coastline retreat. Engineering works, however, provide only partial protection and in some cases actually exacerbate the problem they were designed to cure. During the last century, beach nourishment was recognized as an environmentally friendly

H. Jesse Walker: deceased.

method of shore protection, especially along the coasts of the western world. Today, artificial beach nourishment is the method of choice for shore protection along many developed coasts with eroding beaches (Fig. 1).

Despite the fact that beach nourishment has been used in many hundreds of locations under a wide variety of environmental conditions (e.g., Psuty and Moreira 1990; Silvester and Hsu 1993), and frequently integrated with hard structures as part of strategic shore protection efforts, there is much debate about whether the procedure is the best solution to problems of coastline retreat. Although there are many arguments against beach nourishment, artificial supply of beach-sand remains the most practical method of protecting against coastal flooding from storm surges, for advancing the shoreline seaward, and for widening recreational beaches.

Definitions, Terminology, and Concepts

The term *beach nourishment* came into general use after the first renourishment project in the United States at Coney Island in 1922 (Dornhelm 1995). In engineering parlance, the terms *beach (re)nourishment*, *beach replenishment*, and *beach restoration* are often used more or less interchangeably in reference to the artificial (mechanical) placement of sand along an eroded stretch of coast where only a small beach, or no beach, previously existed. There are, however, subtle connotations in the application of each term. Sediments that accumulate along the shore in the form of beaches are naturally derived from a variety of sources such as fluvial transport in rivers to deposition of sediments in deltas, from preexisting sediments on the offshore seabed, from chemical precipitates (e.g., oolites on carbonate banks in tropical and subtropical environments), or from organisms living along the shore (e.g., shells and exoskeletons from marine organisms). When the natural sediment supply is interrupted, beaches become sediment-starved and the shoreline retreats landward due to volume loss. Efforts to artificially maintain beaches that are deprived of natural sediment supply thus attempt to proxy nature and (re)nourish the beach by mechanical placement of sand. The beach sediment is thus *replenished* by artificial means. *Beach restoration* implies an attempt to restore the beach to some desired previous condition. A nourished or *constructed beach* could be placed along a previously beach-less shore, whereas a restored beach is revitalized by the mechanical placement of sediment.

Beach nourishment projects involve placement of sand on beaches to form a *designed structure* so that an appropriate level of protection from storms is achieved. The placement of sand is commonly by methods such as dredging sand from borrow areas on the seafloor (e.g., Finkl et al. 1997),

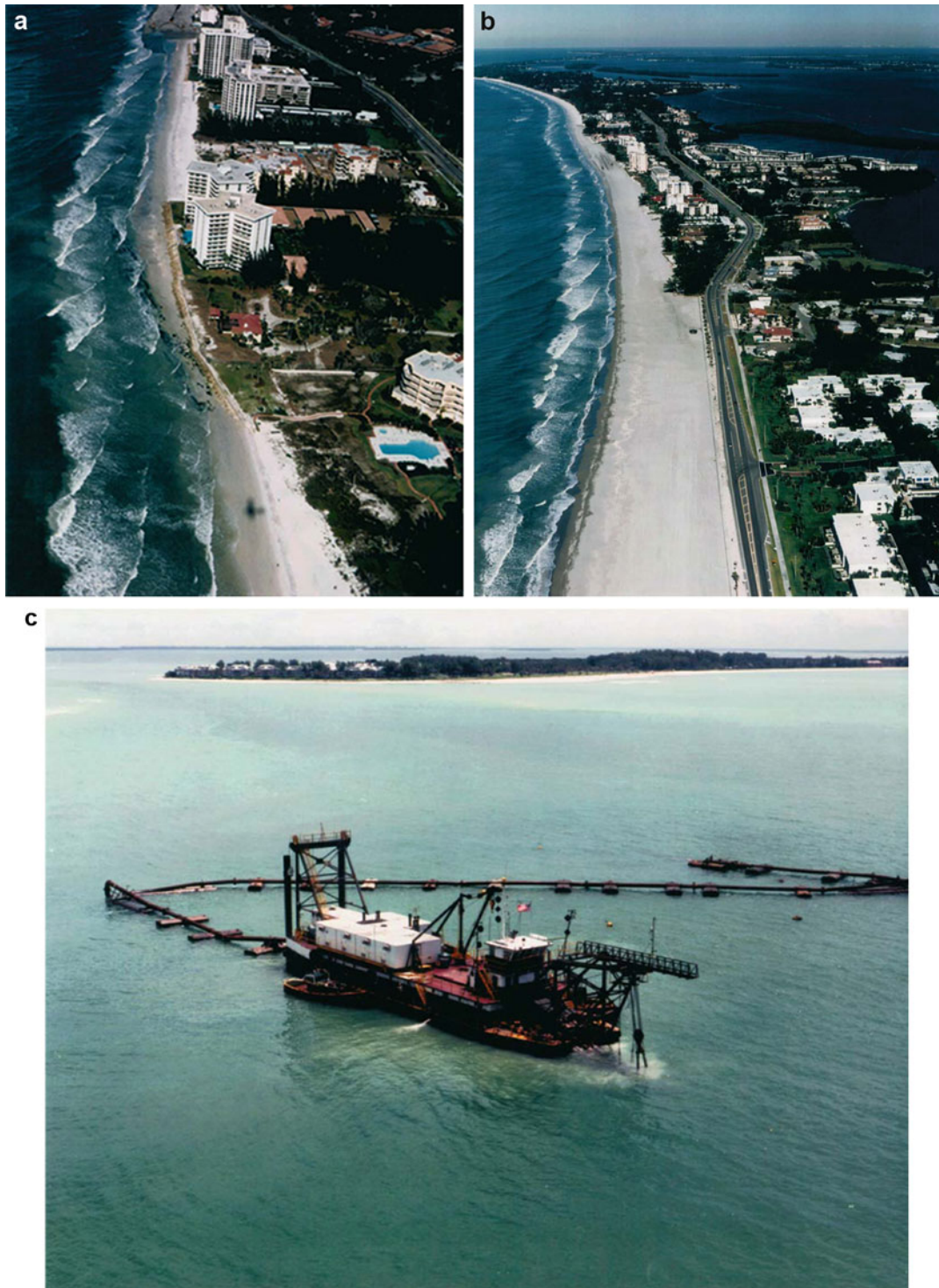
bypassing sand around deepwater inlets or other obstructions (e.g., groins) along the coast that interrupt the littoral drift, or overland delivery of sand from inland quarries to the coast. Although pumping of sand from offshore is the most widespread method of application, due to the large volumes of sediment that are required for most projects, other developments feature placement of sand by trucking or barging from quarries or construction sites, as well as removal of sand from dunes, or relocation of sediment on the berm via *beach scraping* (e.g., Bird 1990; Healy et al. 1990; McLellan 1990; Verhagen 1996). It is now known, however, that removal of sand from dunes is not an appropriate option for sand supply because dune and beach sediment budgets are inextricably interlinked (Psuty 1988). Although most beach nourishment projects deal with sand-sized particles on low to moderate energy coasts, shore protection efforts are also undertaken in very high-energy environments where gravel beaches are featured (e.g., Zenkovich and Schwartz 1987; Peshkov 1993).

Beach nourishment, which entails the construction of new beaches where none existed before or restoration of degraded beaches, usually takes place on developed shores. Along undeveloped shores, beaches provide natural habitat (e.g., nesting grounds for sea turtles and shorebirds) but on developed shores beaches additionally protect coastal infrastructure from storms and are important recreational sites for a globally expanding tourist industry. When beaches are degraded by decreased width and lowering of berms (see entry on ► [Beach Features](#)), many communities choose to replace lost sediment by pumping beach-quality sand from offshore to selected renourishment sites onshore.

Arising from reviews of replenishment activities in the United States, a new terminology was developed to describe the ruggedness or persistence of sand placed along the shore to form artificial beaches. Beach durability defines how well the beach performed under a variety of conditions. The definition of *beach durability* by Leonard et al. (1990a) states that "... the time between placement and loss of at least 50% of the fill volume ..." represents the half-life of a beachfill. The identification of *profile evolution*, performance of the beachfill after placement, is an important consideration of durability and longevity.

Historical Background to the Deployment of Beach Nourishment for Shore Protection

The beach at Coney Island, New York, was the first to benefit from a concerted effort at beach nourishment when, in 1922, more than 1×10^6 m³ of material were dredged from New York Harbor and transported to Coney Island (Hall 1953; Dornhelm 1995). Based on the apparent success of this new



Beach Nourishment, Fig. 1 (c) Beach renourishment on the Gulf Coast of western Florida. (a) Renourishment on Longboat Key, west-central Florida coast near Sarasota showing an eroded beach partly protected by rock revetments and seawalls. In the central foreground there is no beach at high tide. Placement of beachfill is advancing from north to south, as shown in the top of the photo. (b) The restored beach, now 100 m wide, provides a degree of protection from storm surge while

providing a much enlarged recreational beach area. (c) Small coastal suction, cutterhead dredger obtaining sand from an offshore borrow near Captiva Key, southwestern Florida. Beach-quality sand is pumped ashore in a slurry via a floating pipeline. About 75% of the renourished beaches on the Florida west coast have a half-life longer than 5 years (Dixon and Pilkey 1991) (photos: Courtesy of Coastal Planning and Engineering, Inc., Boca Raton, FL)

shore-protection measure, there soon followed a number of other projects along eroding shores elsewhere in New York and along coastlines in New Jersey and southern California. As in the case of Coney Island, beachfills in these early artificial renourishment projects were dredged from sediments in harbors and ship channels. In 1939, Waikiki Beach, Hawaii, was artificially nourished as a recreational beach.

Some other early beach nourishment projects included efforts in Africa and Europe. The construction of jetties for the Durban, South Africa, harbor entrance in 1850 initiated erosion of an adjacent downdrift beach. Groins were built along the shore but they did not stop the aggressive beach erosion. Based on the recommendations of a Belgian engineer, additional long, low groins, in combination with sand bypassing, were added. This was the first attempt at *beach stabilization* using renourishment techniques in South Africa (Swart 1996). Around 1850, seawalls and groins were deployed along the North Sea coast of Norderney in an effort to stabilize this eroding German barrier island. Although these engineering structures prevented dune erosion, they did not stop the loss of sediments from the beach. In an attempt to alleviate problems associated with coastline retreat due to beach erosion, the first large-scale beach nourishment project in Europe was initiated on Norderney in 1951. By 1989 the beach had been renourished an additional six times (Kunz 1993).

Beach nourishment projects have been carried out in many other countries including Australia, Belgium, Brazil, Cuba, Denmark, France, Great Britain, Japan, New Zealand, Portugal, Russia (see discussion in Walker and Finkl 2002). Even though the basic goal of beach nourishment is to elevate the beach and advance the shoreline in order to realize all of the consequent benefits of multiple use but especially increased storm protection, the techniques of sand transfer to the shore and design parameters differ among national approaches. In the United States, the State of Florida has a long and distinguished record of beach nourishment along the southeast coast that involves such notable achievements as: (1) the first sand bypassing weir jetty in the world (Hillsboro Inlet, Broward County), (2) the longest continuously operating fixed sand bypass plant (South Lake Worth Inlet, Palm Beach County) in the world, (3) longest half-life of any renourished beach in the United States (Miami Beach, Miami-Dade County), and (4) the first successful groin-aragonite beachfill project in the United States (Fischer Island, Miami-Dade County) (Finkl 1993; Balsillie 1996).

Needs for Beach Nourishment

Although beach erosion (see entry on ► [Erosion Processes](#)) is common along most coastlines, it is often difficult to recognize in the field unless there are obvious indications of sediment removal. The development of beach scarps in the berm,

dune breaches with overwash, presence of tree stumps or marsh muds on the beachface, and location or damage of buildings precariously close to uprush levels are all signs that beaches are moving landward due to sediment loss. Young et al. (1996) describe these features as *geoindicators* that are helpful for evaluating coastline change along beaches. Such indications of beach erosion are important parameters for estimating the sensitivity and extent of the beach-erosion problem and remediation. The removal of beach materials is by wave action, tidal or littoral currents, or wind. Ranges of countermeasures provide protection from beach erosion, foremost among them during the last quarter of the twentieth century, being *artificial nourishment* (i.e., the mechanical placement of sand on the beach).

Beach protection measures are necessary because beaches are important natural resources that support multipurpose activities. When well maintained, beaches provide storm surge protection, flood control, recreational activities, and habitat for numerous species of plants and animals (Wiegel 1988). Lack of proper coastal maintenance may allow beach erosion to reduce dunes and other natural upland protection, increase loss of natural habitats, degrade a major source of tourist revenue, and shrink the overall economy (Strong 1994; Finkl 1996). Beaches thus need to be protected because they reduce vulnerability to coastal development in high-velocity areas (see entry on ► [Global Vulnerability Analysis](#)).

Although beaches provide a measure of protection to the shore from damage by coastal storms and hurricanes (typhoons and tropical cyclones), their effectiveness as natural barriers against surge flooding depends on their size and shape and on the duration and severity of storms. Beaches are also highly valued as recreational resources that contribute to the economic well-being of many coastal regions in the world. The trend of increasing beachfront development since World War II has resulted in the replacement of dune systems with buildings. This practice has increased exposure of buildings to damage from natural forces (Fig. 2), especially high-energy secular events. The presence of buildings close to an eroding coastline enhances the reduction in beach width because the stabilized shore cannot move landward as it would under natural conditions (see entry on ► [Human Impact on Coasts](#)). Fixation of the coastline by construction in turn adversely impacts both natural storm protection and recreational quality of affected beaches (NRC 1995).

The deterioration or degradation of beaches is regarded as undesirable because beaches provide natural protection from storms and have economic value. The unwanted effects of beach erosion commonly place life and property at risk, usually from flooding, and decrease a community's ability to maintain a viable tourist-based economy. Commercial and residential development on upland areas behind beaches and in close proximity to eroding beachfronts are mainly jeopardized by decreasing (eroding) beach widths (e.g., Wiegel



Beach Nourishment, Fig. 2 Example of an overdeveloped coastal segment along Balneário Camboriú Beach (Santa Catarina State, southern Brazil), where construction of condominiums and tourist facilities restricts the natural dune-beach interaction with phases of seasonal storminess. During perigean spring tides, and storm surges resulting from the passage of atmospheric cyclonic fronts, the beach is under

water and the beachfront road and adjacent shops are flooded. Periodic beach replenishment is required to widen the dry-beach width and add height to the berm. Without beach nourishment, developed coastal segments such as this one lose socioeconomic amenities associated with a wide recreational beach and become increasingly vulnerable to flooding

1988, 1994). Increased potential for economic loss and safety concerns for human life thus drive desires to remediate beach erosion by artificially replacing sand that is lost to erosion.

In addition to the use of beach nourishment for combating coastal erosion, the procedure has been advocated because it: (1) tends to be less expensive and easier to construct than hard structures, (2) is aesthetically desirable, “user friendly” (e.g., Nelson 1993) and “environmentally green” (Finkl and Kerwin 1997), (3) provides a source of sand for wind-created or artificially created dunes which add to the protection of inshore areas (e.g., Psuty 1988; Malherbe and LaHousse 1998), (4) utilizes “waste products” from dredging or construction projects (e.g., Hillyer et al. 1997), (5) contributes to the littoral sediment budget and may benefit down-drift locations (e.g., Lin et al. 1996), (6) capitalizes on natural processes (e.g., Charlier and De Meyer 2000) and thus is more acceptable to society, and (7) restores habitat for biota (e.g., Finkl 1993; Verhagen 1996).

Causes of Beach Erosion

Artificial beach nourishment became necessary only when beachfronts were developed for recreational, urban, industrial,

and military uses. It is often difficult to understand at once the causes of shore erosion because they can be natural or introduced by human activity along the shore. When induced or accelerated by engineering structures the process is sometimes referred to as *structural erosion* (Pilarczyk 1990). Beach erosion is influenced by numerous factors such as uplift (e.g., neotectonism) or subsidence (e.g., groundwater withdrawal, compaction of sediments) of the land surface, change in climate patterns (especially storm frequency, deviation of prevailing wind direction), interruption of sediment supply to the coast, eustatic fluctuations of the sea surface, blockage of littoral drift, and construction on the coast. An increase in relative sea level (i.e., drowning of the coast and landward retreat of coastlines) is, however, often cited as the primary geophysical cause of beach erosion (e.g., Leatherman 1988; Douglas et al. 2000) but many other factors are involved. Construction of dams on major exorheic rivers withholds delivery of sediment to the coast. In the case of the Mississippi River, the sediment load today is about half what it was in pre-dam construction days. Further, sediments that bypass a dam are usually fine-grained and therefore more likely to be carried out to sea and lost to coastal accumulation. Dredging of deep inlets, navigational entrances (see entry on ► [Tidal Inlets](#); ► [Navigation Structures](#)), and the construction of shore protection structures

such as jetties are other interrelated factors that contribute to the degradation of natural beach systems. This list is by no means comprehensive and yet it must be concluded that the causes of beach erosion are manifestly complicated and often interrelated.

Although a range of natural processes contributes to the landward retreat of coastlines, urban development along the shore necessitates placement of sand on eroding beaches in an effort to protect infrastructure. The essence of the problem is not the dynamic adjustment of coastlines to fluctuating ambient conditions but construction too close to the water. Most coastline development is deliberate for reasons of access, proximity, or aesthetics. Whatever the initial impetus for developing coastlines, the result has been an expensive exercise in what is usually nationally funded coastal protection. Coastline retreat in Australia, for example, is not as problematical as it is in the United States because the Commonwealth government established Crown lands along most of the coast keeping urban development some distance inland. In other countries such as The Netherlands where land has been reclaimed from the sea, large dikes and other engineering structures (see Walker 1988) such as dunes and renourished beaches are part of efforts to hold back the sea. Although

these areas have multipurpose uses (e.g., storm protection, conservation, recreation, water catchment), they are not open to intensive urban uses (Pilarczyk 1990).

Coastal development often includes the dredging of ports and harbors and the navigational channels that serve them. Jetties that provide protection from waves in the channel are used to stabilize the geographic migration or orientation of many entrances. Jetties and deep inlets, as well as other shore protection structures such as groins, interrupt the natural littoral drift by impounding sediment or causing it to be jetted offshore beyond the longshore sediment transport system. It is now widely recognized that the interruption of sediment transport along the shore by artificial structures causes downdrift shores to become sediment-starved, which in turn results in shore erosion and retreat of the coastline (Fig. 3). Large littoral drift blockers (e.g., deep navigational entrances, groin fields) can initiate downdrift erosion that propagates for several tens of kilometers (Bruun 1995). In a study of 1238 km of Florida coastline, distributed among 25 coastal counties and covering about 95% of the state's beaches, Finkl and Esteves (1998) concluded that littoral drift blockers on Florida's Atlantic and Gulf coasts accounted for 72% of the statewide beach erosion. Along the southeastern coast where there are



Beach Nourishment, Fig. 3 Hillsboro Inlet, Broward County, southeast Florida. Erosion of the downdrift coast, seen in the coastal offset (photo center), was caused by stabilization of this inlet by jetties. Subsequent construction of a weir jetty on the updrift side of the inlet permitted sand to overwash into an interior sand trap on the landward side of the jetty. A floating dredge sucks the sand from the trap and pumps it to the downdrift side of the inlet via a submerged pipeline.

Dredging is mostly conducted during storms when sediment is overwashed through the weir because the trap quickly fills with sediment. If the sand trap is not cleared as it fills, excess sediment spills into the navigation channel where it becomes a hazard to boaters. This sand bypassing arrangement moves 100% of the estimated net littoral drift to the downdrift beach which is thus maintained in a healthy state

numerous stabilized inlets, and the volume of sediment transported in the littoral drift is relatively small (e.g., $<50,000\text{--}100,000\text{ m}^3\text{a}^{-1}$), littoral drift blockers appear to cause about 90% of the beach erosion. Most of the beach erosion here is thus anthropogenically induced and is, at least theoretically, quite preventable from a technical point of view. In practice, however, remediation of the beach erosion problem is politically recalcitrant.

The causes of beach erosion are complicated interacting processes and it is emphasized that beach nourishment only treats erosional symptoms and does not eliminate the causes. Beachfills are sacrificial in the sense that they are not permanent solutions to the beach erosion problem; they thus provide only temporary protection and it is anticipated that replenishment will be repetitive.

Design of Beach Nourishment Projects

Beach nourishment projects are designed to: (1) increase dune and berm dimensions (i.e., height, length, and width), (2) advance the coastline seaward, (3) reduce storm damage from flooding and wave action, and (4) widen the recreational

beach area. Beach nourishment projects are complicated technical procedures that require careful preparation for successful execution of site-specific engineering design (Finkl and Walker 2002). The scale of mechanical sediment supply is quite variable, ranging from large enterprises that involve federal and local partnerships where sometimes tens of millions of cubic meters of sand are involved to small-beach restorations that may require less than $50,000\text{ m}^3$ of sand. Equipment for obtaining, transporting, and placement of sand on the beach varies with the scale of the project. Large projects require robust equipment that can handle large volumes under high-energy conditions for open ocean dredging (Fig. 4). Placement of sand on small, protected beaches (e.g., pocket beaches, embayed shores) (see also entry on ► [Bay Beaches](#)) can often be achieved with small dredges during fair weather conditions (Fig. 5). In either case, sediment is often redistributed along the shore by front-end loaders, graders, and tractors to achieve the design profile (see also Figs. 1 and 6).

Emergency repair of *erosional hot spots* (localized coastal segments where there is increased erosion and rapid shoreline

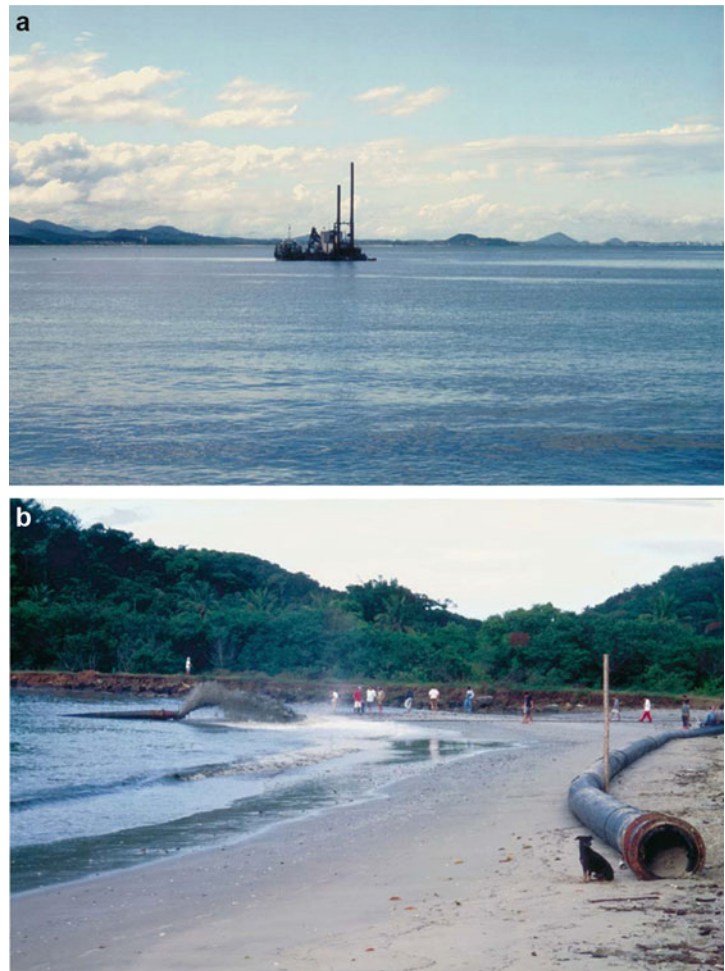


Beach Nourishment, Fig. 4 Large ocean-going dredge operating off the coast of southeast Florida. The *Illinois*, owned and operated by the Great Lakes Dock and Dredge Company (Oak Brook, IL, USA), is one of the largest suction cutterhead dredges in the United States. The 98-m long dredge, which digs to a nominal depth of 32 m, has a 760 mm discharge diameter and can pump sand 7635 m without booster pumps. With total installed power of 8400 kW, and 1400 kW cutter power coupled with 662,000 L fuel capacity, the *Illinois* is ideal for working offshore for long periods in rough weather conditions. The dredge is not

self-propelled and must be moved to projects by tugboats. When on site, the dredge has a large swing radius using a system of anchors and cables to maneuver within about a 300 m range without resetting the anchors. This dredge can work in swell up to 2 m high and usually operates 24 h a day when possible. Beach-quality sand from the borrow area is pumped in a pipeline to the beach. Because the dredge might operate up to 2 km offshore, the pipeline is floated on the surface near the dredge for ease of maneuverability and submerged near the shore for safety of boaters and beach users

Beach Nourishment,

Fig. 5 Small-scale beach renourishment at Alegre Beach, Santa Catarina, Brazil. Local erosion of a beach on the downdrift side of an inlet required sand renourishment for shore protection, recreation, and marine fisheries (beach launching sites for local fisherman). (a) The small coastal dredge pumps sand from the seabed offshore to the beach via a submerged pipeline. (b) Sediment is pumped up onto the upper beachface in a slurry, as shown in the subaerial plume in the photo center. Extra pipe is stored on the berm for future use as the dredge moves about in the offshore borrow area



retreat that dramatically exceeds background rates of erosion, as described by Finkl and Kerwin (1997) that develop during storms may require only a few thousand cubic meters of sediment until more thorough corrective action can be initiated. Beach nourishment the world over is based on the application of natural sediments, mostly beach-quality sands derived from offshore dredging. Many developed nations now recycle glass products and volumes of recycled glass cullet are increasing yearly. The State of Florida, for example, annually produces in excess of 130,000 tonnes of surplus glass cullet that could be made available for beach nourishment (Finkl and Kerwin 1997). Glass cullet, a chemically inert form of silica, can be graded (mechanically ground) to desired colors and grain sizes to perfectly match native beach sands. For small projects, costs per cubic meter of placed cullet are usually less than beachfill sands.

The design process specifies the quantity, configuration, and timing of sediment distribution along a specific coastal segment to emulate natural storm protection or recreational area, or both. The design must consider rates of long-term (background) erosion as well as temporal impacts of storms and wave climate to adequately address variables associated

with the quantity, quality, and placement of beachfill along the shore. As a general rule, sediment comprising the nourished beach is anticipated to erode at least as fast as the background rate of the pre-nourished coastline. It is usually observed in practice that sediment volume loss rates and coastline retreat for artificial beaches are significantly greater than historical rates for the natural beach (e.g., Dean 2000), even when differences in grain size and sorting are taken into account (e.g., Ashley et al. 1987). Although an allowance for continued erosion of beachfill is part of the design assessment, the purpose of beachfill design is to maximize the longevity of artificial beaches. The designs can only be optimized by changing the morphological configuration of the beachfill or by the choice of the fill material. The grain-size of borrow material was traditionally considered to be the most important factor for optimizing beachfill. Studies by Eitner (1996), however, indicate that grain-size has little effect on beachfill longevity because grain-size influences the critical threshold stress to a lesser extent than does the grain density. Only a coarser material such as gravel, which also has a significantly higher critical threshold stress, may effectively extend beachfill longevity. In general, however, most researchers



Beach Nourishment, Fig. 6 Restoration of Captiva Beach in 1996 on Sanibel Island, southwestern Florida, showing beach-quality sand pumped from offshore via a submerged pipeline to the shore. The sand-water slurry debauches from the pipe end (lower center foreground) to form an alluvial fan that builds up along the shore. Shell hunters often congregate about the proximal part of the sediment fan hoping to find collector's specimens. Shore birds search the distal portions of the fan for crustaceans brought up from the seafloor in the offshore borrow area. Road graders and bulldozers redistribute the pumped sediment into the

beachfill design shape that often incorporates overfill that will be sacrificed to the littoral system as the beach equilibrates to the ambient wave climate. The beachfill is sometimes initially somewhat darker than the native beach sand, as shown here, due to a small percentage of organic content. Organic matter in the fill bleaches out within a few weeks and the native-colored sands imperceptibly grade into the fill material. Pipe extensions lie at the foot of the foredunes (photo, lower right) and mark the approximate position of the back berm (photo: Courtesy of Coastal Planning and Engineering, Inc., Boca Raton, FL)

agree that coarser grain sizes produce steeper, more stable, and longer-lived fills (e.g., Bruun 1990; Pilkey 1990; Smith 1990; Dean 1991). Controversy does, however, surround the kinds of methods used to estimate erosion rates. Houston (1990), for example, emphasizes that designer extrapolations or predictions of replenished beach life of one to several decades is not advisable because beach conditions are too variable and especially vulnerable to cycles of storminess.

There are various methods of *beach nourishment design* that are complimentary in the overall process of optimizing *project performance*. Potential designs are initially evaluated at a preliminary level in which the anticipated project performance is predicted using simple and relatively inexpensive methods. After the performance characteristics are compared with project objectives, the design is refined until the performance predictions confirm an optimal design. For sites without complex boundaries (i.e., straight beaches without terminal groins, inlets, or headlands), prediction tools correctly estimate the time required for renourishment to within approximately 30% of actual project performance (NRC 1995). Subsequent to establishing the preliminary design, more sophisticated predictive methods are used to optimize the design. This bimodal approach checks preliminary and advanced methods of design, facilitates a rapid and efficient convergence to final design, and provides a clear perspective of how well the design parameters fit project requirements. If the predicted volumetric losses, based on preliminary and advanced methods, differ by more than 50%, the essential

elements of the design procedure are reviewed and revised, where necessary.

The Design Beach

The *design profile* is the shore-normal cross section that an *equilibrated beach* is anticipated to assume. The best estimate of this profile is obtained by the seaward transfer of the natural beach profile by the amount of beach widening that is required (USACE 1992). Estimates of beachfill volumes are generally increased if the borrow material is finer grained than the native sand. The *construction profile* is the cross section that the contractor is required to achieve. Because the constructed beach, which contains design fill and the advanced-fill volumes, is often steeper than the design cross section due to construction limitations, it is also usually significantly wider than the design profile. Wave action adjusts the construction cross section to a flatter dynamically equilibrated slope within the first few months to a year after placement of the beachfill (cf. Fig. 7). Because the dynamically adjusted profile contains *design and advanced fills*, it is wider than the design width during the *nourishment interval* (the time elapsed between nourishment episodes). At the time of renourishment, the design and equilibrium profile are theoretically equal (NRC 1995).

Mechanical deposition of sediment along a beach nourishment site, during initial construction or renourishment, may not correspond to the natural profile of the beach at the time of placement (Fig. 8). In the United States, use of a construction



Beach Nourishment, Fig. 7 Example of a large beach-replenishment project involving more than six million m^3 of sand that was pumped from offshore in southeast Florida near Port Everglades (Fort Lauderdale). As shown in the photo, the beachfill has been eroded back to the native sand beach (left) and the built beach is perched about 2 m above the eroded

beach. The beach scarp that developed during erosion of the fill was graded by dragging a heavy I-beam across the sand. The more gradual slope facilitated access to the water by reducing the hazardous vertical face of the scarp

profile rather than a natural profile is the normal placement practice. It is customary for nourishment designs in the United States to establish uniform beach width along a project's length. It is also standard practice to provide sufficient sand to nourish the entire profile from the dune to the depth of significant sand removal, the so-called *depth of closure* (DoC). The DoC is a term used by engineers to define the *depth of active sediment* movement on the seabed (see also entry on ► [Depth of Closure on Sandy Coasts](#)). Other terms that are related to this critical concept include profile pinch-off depth, critical depth, depth of active profile, maximum depth of beach erosion, seaward limit of nearshore eroding wave processes, and seaward limit of constructive wave processes. The DoC in beachfill design is defined as "The depth of closure for a given or characteristic time interval is the most landward depth seaward of which there is no significant change in bottom elevation and no significant net sediment transport between the nearshore and the offshore" (Kraus et al. 1998). This definition applies to the open coast where nearshore waves and wave-induced currents are the dominant sediment-transporting mechanisms. The definition of the DoC infers or stipulates that: (1) the landward water depth at which no sediment change occurs can be reliably identified, (2) there is an estimate of no significant change in bottom elevation and no significant net cross-shore sediment

transport, (3) a time frame is related to the renourishment interval or design life of the project, and (4) at the DoC cross-shore transport processes are effectively decoupled from transport processes occurring farther offshore.

Estimates of fill requirements are based on the geometric transfer of the active cross-shore profile seaward by the design amount. If the beachfill grain size matches the native sand and there are no rock outcrops, seawalls, or groins, the design profile (shore-normal cross section) at each alongshore range marker (permanent locations of cross-shore survey sites are typically spaced every 330 m along Atlantic and Gulf coast beaches in Florida) should ideally match the dynamically stable shape of the native beach profile. Cross sections may be more closely spaced in beach nourishment project areas for better engineering control. Enough sediment is included in the design to nourish the entire profile (Hanson and Lillycrop 1988).

The total sediment volume is independent of the cross-shore profile because the shape of the *renourished profile* is parallel and similar to the existing natural profile. Extra fill is required, however, in front of seawalls in order to achieve the proposed berm elevation. After these seawall volumes are calculated, estimates of nourishment fill volumes are based on seaward translocation of the entire profile. It is emphasized that sand is usually needed along the entire profile, both above



Beach Nourishment, Fig. 8 Emergency beach renourishment at Gravatá Beach, Santa Catarina, Brazil. A coastal highway and commercial infrastructure was threatened by beach erosion during a strong southeaster that brought heavy surf conditions to the coast during the

Southern Hemisphere winter of 1999. Removal of the beach by wave and current action, undermined part of the coastal highway interrupting coastal access. Although of finer grain size than the native beach sand, emergency fill was trucked to the site for immediate shore protection

and below the water because the beach, by definition, retains subaerial (berm) and submarine (beachface) sections. Placement of the required extra fill volume in front of seawalls typically moves the high-tide shoreline farther seaward than adjoining non-seawall segments. This design requirement, however, causes alongshore gradients in littoral drift that

tend to become erosional hot spots. An alternative to providing the additional seawall volumes is to build narrower berms in front of seawalls. Narrower berms are advantageous because they reduce littoral drift gradients that are set up by overly wide sections of nourished beaches in front of seawalls. Similar levels of storm protection (for uplands) are

provided by narrower berms when they are backed by seawalls compared to wider berms without them. In many instances, however, beach nourishment in front of seawalls can become problematic, especially where coastline retreat extends landward along coastal segments adjacent to the seawall and where there is deep wave scour in front of the wall (see discussions in Kraus and Pilkey 1988).

Coastal engineers attempting to predict beach washout and profile response seaward of seawalls often employ beach and dune recession models. Commonly used approaches include EDUNE (Kriebel 1986), SBEACH (Larson and Kraus 1990), and GENESIS (Hanson 1989; Hanson and Kraus 1989). The numerical models predict the evolution of the cross-shore beach profile toward the so-called equilibrium storm profile (NRC 1995). Both models are based on principles related to the disparity between actual and equilibrium (theoretical) wave-energy dissipation per unit volume of water within the surf zone. For convenience of calculations, the models assume that sand eroded from the upper beach deposits offshore, with no loss or gain of material to the profile. It is well-known, however, that beach sediment is often transported offshore and is lost from the littoral drift system (e.g., Pilarczyk 1990). Estimates of storm surge used in coastline recession models, and for calculating run up, are based on USACE (1984, 1986, 1989) engineering manuals. Storm-surge frequencies and extents of coastal flooding are also deployed by the Federal Emergency Management Agency (FEMA) and the National Oceanographic and Atmospheric Administration (NOAA). Storm hydrographs are thus obtained from FEMA, NOAA, and universities to generate probabilities of storm-induced shoreline recession. Wave statistics can be obtained from wave gauge records, published summaries of observations, or wave hindcast estimates such as the Wave Information Study (USACE 1989). Other methods such as the Empirical Simulation Technique (EST) (Borgman et al. 1992) develop joint probability relationships between various multiple parameters contained in historical data records. Using historical storm records as input, the EST statistically develops a much larger storm-response database while maintaining statistical similarities to original data and it is thus possible to achieve estimates of storm-induced beach recession (Howard and Creed 2000).

Protocols for Overfill on Design Beaches

Beachfill is usually dispersed out of the nourishment area (i.e., away from the replenished or artificial beach) to adjacent shores or deeper water. The process leading to a decrease in beachfill volume is referred to as “loss,” although this sand still temporarily contributes to the stability of the shore in general, but not at the original location. From the point of view of sediment transport, the sand is not “lost” because it is partly retained in the littoral system. From the perspective of the beach manager, however, migration of sediment away

from the beachfill represents a tangible decrease of dry beach area.

Erosion of nourished beaches feature two distinct components: (1) the linear regression of the volume of sand in the coastal profile and (2) additional erosion arising from the newly nourished shoreline which becomes more exposed (lying more seaward) than adjacent shore up-and down-drift (Verhagen 1996). Sediment losses alongshore as well as adjustment or equilibration of the constructed cross-shore profile are responsible for the so-called “additional erosion.” In cases where the volume loss associated with the coastal erosion is large compared to the quantity of the beachfill and where the previous rate of erosion is known, a multiplier is used to compensate for all sand loss. Verhagen (1996) suggests a value of 40% extra fill. The presence of structures such as seawalls, due to their interaction with coastal processes, may also require additional fill.

The term *advanced fill* refers to the eroded part of the beach profile before nourishment becomes necessary. The volume and areal distribution of advanced fill is estimated from analysis of the historical rate of erosion and shoreline change. The potential impact of *project fill* on coastal processes is an additional consideration that is taken in account. Procedures used to make these estimates include the historical coastline change method (USACE 1991) or numerical methods (Hanson and Kraus 1989). The *historical shoreline change method* assumes that the nourished beach will erode at the same rate as the pre-nourished beach. This method is commonly employed by beach designers (based on survey results) but can yield a significant underestimate of nourishment requirements (NRC 1995).

Most long-term erosion (as opposed to episodic storm erosion or development of erosional hot spots) of a nourished beach is initiated by increased gradients of littoral drift along the project length. Major littoral drift gradients affecting the stability of nourished beaches are the preexisting background (regional) rates or historical erosion of the pre-nourished shore and stresses associated with the high-tide shoreline salient that was advanced seaward by the project fill. These perturbations of normal coastal processes are the cause of *end losses* and *spreading* of the fill. All of these littoral drift gradients interact with the nourished beach causing a progressive loss of fill. Exclusive consideration of the background erosion rate neglects end (and spreading) losses, which causes an underestimate of nourishment volume and overestimate of project life. Although losses from the project due to spreading cause accretion on adjacent beaches, they must be included in the advanced-fill design in order to achieve performance objectives (NRC 1995).

Nourishment Profiles

Models of beachfill placement depend on renourishment design schemes that are selected by considering static and

dynamic peculiarities of the site, fill requirements, temporal and spatial distribution of natural habitats, and costs. Some of the more common approaches include: (1) placing all of the sand in a dune behind the active beach, (2) using the nourished sand to build a wider and higher berm above mean water level, (3) distributing fill material over the entire beach profile (above and below water), or (4) placing sand offshore in an artificial bar (NRC 1995). The approach taken partly depends on the location of the source material and the method of delivery to the beach. If the borrow site is a quarry on land and the sand is transported by trucks to the beach (cf. Fig. 5), placement on the berm or in a dune is generally most economical. If the material is pumped shoreward from offshore ocean-going dredges (cf. Fig. 3), it is usually more practical to place the sand directly on the beach, in the near-shore zone, or to build an artificial bar. If pumped onshore in a sand-water slurry, the sand is subsequently redistributed by grader or bulldozer across the shore to form a more natural beach profile (Fig. 6).

The use of large dunes (i.e., man-made dikes) fronted by renourished beaches as an effective coastal protection measure has long been recognized in The Netherlands (Verhagen 1990; Watson and Finkl 1990). These constructed dune-beach systems are designed to withstand the 1-in-10,000-year condition of wave intensity and storm surge flooding. This extreme level of protection is justified because entire cities lie behind the coastal defenses.

Bruun (1988) advocates nourishing the entire beach profile, which he terms *profile nourishment*. The main advantage of this approach is that the sand is placed in approximately the same configuration as the existing profile, so that drastic initial adjustments are mostly avoided, especially the rapid erosion of the nourished berm. When wave action undermines the newly constructed berm, a beach scarp frequently forms along the length of the project fill. These scarps can pose hazards to beachgoers trying to gain access to the water from the berm (Fig. 7). In some cases, foot traffic across the scarp tramples the steep slope to a flatter one that cuts into the beachfill. These cuts or beach tracks can provide ingressive pathways for surge and run-up which in turn can accelerate erosion of the project fill volume.

Artificial nourishment of eroded beaches can be indirectly achieved by placing dredged sand in the offshore zone (McLellan 1990). Dredged material is deposited in shallow water, typically using a split-hull barge either as a mound or shaped as a long liner ridge that simulates a shore-parallel sand bar. It is anticipated that the sand deposited in the offshore mound or artificial bar will migrate onto the beach. Prior to welding onto the beachface, the bar causes waves to break farther offshore, a process that reduces the wave energy on the beach in the lee of the bar. The disposal depth of the offshore nourishment should be shallower than the seaward boundary of active sediment transport (as defined by normal

to moderately elevated energy conditions) so that sediment quickly moves onto the subaerial beach.

Mechanical Bypass Systems

Shore protection structures (*q.v.*) or other coastal construction works can interrupt littoral drift flow patterns and trap sediment near structures, within navigational entrances to port and harbors, and in flood- or ebb-tidal deltas. Sediment trapping by littoral drift blockers causes downdrift beach erosion (e.g., Finkl 1993; Bruun 1995). In order to mitigate the downdrift effects of sand starvation along the coast, it is necessary to move sand around barriers in order to supply beaches with sediment. Due to losses of sediment offshore (see previous discussion), the quantity needed for downdrift beach nourishment may be greater than the trapped sediment volume. Some bypassing systems that are geared for normal use may be overwhelmed during large storms while others function best during or immediately after storms when sediment is brought to the *sand trap* (a dredge pit that collects sediment).

Fixed bypassing systems generally are less effective and more expensive to run than mobile floating systems (Bruun 1993). Most bypassing plants work at less than 50% efficiency, and some at 30%, which means that less than half of the drift is bypassed to the downdrift beaches. The combination of periodic beach replenishment and innovative bypassing techniques is an option that can restore longshore sediment transport and greatly reduce beach erosion (Bruun 1996). Suggested new alternatives include mobilization of the bypass intakes on rails or cranes, implementation of jet pumps, or seabed fluidizers (Bruun and Willekes 1992).

Several different kinds of mechanical bypassing systems are used effectively in a variety of coastal settings: (1) mobile dredges in the harbor and or entrance (e.g., Santa Cruz, CA), (2) movable dredge in the lee of a detached breakwater that forms an updrift sand trap (e.g., Channel Islands and Port Hueneme, CA), (3) floating dredge within an entrance using a weir jetty on the updrift side (e.g., Hillsboro Inlet, FL; Boca Raton, FL; Masonboro Inlet, NC; Perdido Pass, AL), (4) fixed pump with dredge mounted on a movable boom (Lake Worth Entrance, FL; South Lake Worth Inlet, FL), (5) jet pumps (eductor) mounted on a movable crane, with main water supply and booster pumps in a fixed building (e.g., Indian River Inlet, Delaware) (NRC 1995). These, and other installations, and their operational performances are described in engineering and design manuals (e.g., USACE 1991) which provide guidance for the design and evaluation of sand bypassing systems.

Veneer Beachfills

Veneer fills are beach-quality sands that are placed over a relatively large volume of material that is generally not suitable for beach nourishment. The unsatisfactory materials,

which may be either grossly coarser or finer than normal beach sand, remain as an underlayer beneath the thin surface sand veneer. Veneer beachfills are thus used in situations where beach-quality sand is not available in sufficient quantities to economically undertake a nourishment project. The usual reason for placing a veneer fill is based in economics where the cost is prohibitive if a cross section is totally built of beach-quality sand. Veneer fills are of two basic types: (1) fills where the underlying materials are coarser than typical beach sands (e.g., boulders, coral, rocks) and (2) fills where the underlying materials are finer than typical beach sands (e.g., silts or silty sands where the median grain size is much smaller than native sand). In the United States, veneer beachfills have been used in Corpus Christi, TX; Key West, FL; and Grand Isle, LA (NRC 1995). A fundamental design problem associated with veneer fills involves selection of a veneer that is thick-enough, so that it will not erode away and expose the underlayer during storms or before scheduled replenishment. Although variable, depending on the local conditions, the thickness of the veneer must provide a sedimentary envelope that incorporates profile variations without compromise.

Pros and Cons

In the United States, there are two schools of thought regarding beach nourishment; the larger group favors artificial placement of sand on eroding beaches as part of shore protection measures while the other smaller group discourages the practice on the grounds of environmental, economic, sociological, or political grounds. A recent study by the US National Research Council (NRC) examined the diversity of viewpoints about the success or failure of nourishment projects (Table 1). It was concluded that the factors involved include a large number of interest groups who

have different "... viewpoints, objectives, needs, and ideas ..." (NRC 1995, p. 41).

In the 1980s and early 1990s, there was much debate about the pros and cons of beach nourishment with many illuminating facts coming from both sides of the issue (Pilkey 1990). As persuasive as many arguments were, the end result was that federal support for new beach nourishment was largely withdrawn when the US Congress removed the USACE from many new projects by reducing or eliminating funding (e.g., Finkl 1996). If beach renourishment is to continue as a shore-protection measure in the United States, local funds will have to support the practice along many stretches of the coast.

Proponents of beach nourishment favor continuance of the engineering practice for many reasons, the most important of which feature shore protection (mainly flood control against storm surge) and economic value in terms of income from recreational use. The arguments for beach nourishment are legion and include those factors already indicated as part of the needs for shore protection.

Antagonistic to views of beach nourishment are concepts that focus on environmental impacts of offshore dredging (e.g., Nelson 1993), especially near sensitive environments such as coral reefs and sea-grass beds, and placement of sand along the shore which buries meiofauna and infauna, or which may adversely impact rare species such as sabellariid worm reefs. Divergent opinions also focus on expenditures of public funds for coastal segments that do not provide public access to the beach or interpretations of coastal management practices that call for retreat from the coast. Other issues that are sometimes raised concern beachfill performance (i.e., durability, longevity, half-lives of replenishments) which is keyed into the design life of renourished beaches (e.g., Houston 1991; Davis et al. 1993; Farrell 1995). The USACE, for example, often estimated and advertised life spans of a decade

Beach Nourishment, Table 1 Evaluation of beach nourishment projects, based on objectives, interpretation of criteria for success, and various measures of performance (Modified from NRC 1995)

Objective	Criteria for success	Measures of performance
Create, improve, or maintain a recreational beach	A viable recreational asset (acceptable dry-beach width and carrying capacity) during the tourist season	Periodic surveys of beach width using quantitative techniques. Assessment of beach visits and carrying capacity via such methods as aerial photography
Protect coastal infrastructure from wave attack and flooding by storm surge	Sufficient sand, gravel, or cobbles remaining in a configuration that blocks or dissipates wave energy and surge. Hard structures may be included in the solution	Evaluation of structural flooding damage following storms that do not exceed the project design
Maintain an intact dune seawall system	No overtopping during a storm that does not exceed the design water level and wave height limits	Verification of stabilized shore line position
Create, restore, or maintain beach habitat	Episodic erosional extremes do not exceed the design profile. Structures, if allowed, remain intact. Postfill erosion rates comparable to historical values	Profile surveys showing that the sedimentary volume and configuration meet or exceed the design profile
Protect the environment	Sediment volume, areal extent, and condition plus vegetation of the back beach or dune meeting environmental needs	Observation and survey of habitat characteristics and conditions
Avoid long-term ecological changes in affected habitats	Return to pre-nourishment (native beach) conditions within an acceptable time frame	Periodic monitoring of assemblages of critical concern

or so for many proposed projects. Studies subsequently found that, on average, the life spans of renourished beaches were usually less than anticipated. Durabilities of renourished Atlantic beaches, for example, were found to have a half-life of about 4–5 years. (Pilkey and Clayton 1989). The percentage of renourished beaches lasting more than 5 years along Atlantic coastal barriers averaged about 65% while those on Gulf and Pacific coasts, respectively, averaged about 75% and 65% (Leonard et al. 1990b; Dixon and Pilkey 1991; Trembanis and Pilkey 1998).

Conclusions

Among the different approaches to beach nourishment the world over, are various techniques that essentially boil down to methods of placing suitable sediment along the shore to: (1) maintain an existing but eroding beach, (2) create a new beach where none existed before, or (3) to improve a seriously degraded former beach. No matter the actual method of beach nourishment design that involved acquisition of suitable sediment and placement along the shore, this soft engineering approach to erosion mitigation must be regarded as a temporary solution to a chronic problem. In spite of the fact that in the United States, for example, there has been more than a half-century of experience, beach nourishment remains a procedure with unclear universal application. Experience has shown that there are no simple rules that work everywhere because it is now widely appreciated that local site characteristics must be important criteria in successful design. Peculiarities of local conditions related to bathymetry, sediment grain size or shape and composition, exposure and orientation of the beach to prevailing and storm wind patterns, wave climate, and para- and diabathic sediment flux pathways can all affect shore erosion and beach stability. Intricacies of shore processes and their interactions with engineering works such as jetties, dredged channels, groins, and breakwaters, for example, can exacerbate natural shore erosion. Fortunately, it is now realized that many shore protection structures are themselves the main causes of accelerated beach erosion. In southeast Florida, for example, stabilized navigational entrances (i.e., jettied tidal inlets) are responsible for about 90% of the beach erosion problem (e.g., Finkl and Esteves 1998). As formidable as this figure may seem, it is now evident that improved sediment bypassing at littoral drift blockers, as described by Bruun (1995), can significantly enhance beach nourishment efforts by prolonging what are relatively short life spans of placed sediments. Using a combination weir jetty, interior sand trap, and floating suction cutterhead dredge, the Hillsboro Inlet in Broward County (southeast Atlantic coast of Florida), is able to annually bypass 100% of the estimated net littoral drift (Finkl 1993). Thus, in many instances, improved sand bypassing at inlets

can maintain sediment transport alongshore to supply down-drift beaches with incremental sand.

Shore protection via beach nourishment comes, however, with a high price tag but there often are few options that are practical. Retreat from the shore in highly developed coastal infrastructures is not possible nor is a passive approach where structures or facilities are threatened by beach erosion or coastal flooding. The Dutch, for example, have taken an aggressive approach by actually reclaiming land from the seabed by diking and poldering (see Walker 1988). Elsewhere, most of the world's developed shores thus face the promise of attempting to maintain present coastlines via beach nourishment.

Even though beach nourishment is the shore protection measure of choice for many coastal managers, the future of the procedure in the short-term (less than 50 years beyond today) may seem bright but in the long term (more than 100 years from today) the prognosis would be poor. If the natural rise in mean sea level (see entry on ► [Changing Sea Levels](#), ► [Sea-Level and Climate Change](#)) continues to be exacerbated by human action to the point that relative sea level continues to increase, many coastal areas will experience inability to locate suitable beachfill materials in sufficient quantity to support artificial nourishment. As beachfill materials become scarcer due to increased demands, project costs will escalate, but cost/benefit ratios will probably be favorably maintained because of higher property values per length of coastal segment. Further, if the general rise in eustatic sea level accelerates as some researchers predict, renourished and constructed beaches will be no match for increased vulnerabilities from erosion and storm surges. The problem is, unfortunately, growing as more and more stretches of shore are developed.

Recommendations to improve performance (i.e., life span, durability of beach nourishment projects) include mapping of the shore zone (both subaerial and submarine features) to better understand the topographic features and sediment transport pathways that are related to coastal stability in the local area. Post-project monitoring is another important step that can help assimilate factors that are related to the degradation of beachfill project life. For now, beach nourishment projects meet the needs of many coastal communities that require protection of beaches.

Cross-References

- [Beach Erosion](#)
- [Beach Processes](#)
- [Coastal Management Practices](#)
- [Cross-Shore Sediment Transport](#)
- [Dredging of Coastal Environments](#)
- [Erosion Processes](#)

- Natural Hazards
- Net Transport
- Sandy Coasts
- Sediment Budget
- Shore Protection Structures
- Shoreline Response to Littoral Drift Barriers
- Storm Surge

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Beach Processes

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The continuous changes taking place in the coastal zone constitute *beach processes*. A beach is one part of the *coastal zone*, which is the transitional area between terrestrial and marine environments. The coastal zone comprises the beach; an underwater region that extends seaward to the depth where waves no longer effect the sea bottom; and inland to features such as *sea cliffs*, *dune fields*, and *estuaries* (*q.v.*). Beach studies focus on understanding spatial and temporal changes in alongshore and cross-shore geomorphic features of the beach (*Beach Features: q.v.*) and in the size and composition of beach sediment. Over time, the coastal zone changes in character in response to changes in wave climate and other physical processes.

Comprehensive knowledge of beach processes is crucial to society because the majority of the world's coastlines are eroding (Thornton et al. 2000). Moreover, *sea-level rise* (*q.v.*) from *global warming* could accelerate coastal land loss, resulting in an increasing rate of loss of coastal habitat and structures. Coastal land loss has a large negative societal impact because approximately two-thirds of the world's population lives adjacent to the ocean or large inland bodies of water. In the United States, more than half of the population

lives within 50 miles of the shore, while about 85% of the sandy shore is eroding from a combination of damming of rivers, inlet improvements, sand mining in the coastal zone, sea-level rise, and large storms (*q.v.*). Understanding the processes that cause land loss on sandy coasts is particularly important because beaches are a popular recreational area; are essential to commerce; and protect coastal cliffs, dunes, and structures.

Beaches continually change in response to changes in wave conditions. The changes occur over both short- and long-terms, reflecting both subtle changes associated with daily or weekly variations in tidal level or wave climate and gross changes associated with seasonal variations in wave climate. Beaches usually shift back and forth between a wide, built-up berm with a barless nearshore zone and a small-to-absent berm with one or more well-developed bars in the nearshore. These fluctuations often occur on top of an equilibrium profile that exhibits no net long-term change.

Textbooks that discuss beaches, the coastal zone, and the processes that affect them include Johnson (1919), Guilcher (1958), Shepard (1963), Shepard and Wanless (1971), Bascom (1980), Komar (1983, 1998), Bird (1984), Carter (1988), Carter and Woodroffe (1994), and van Rijn (1998). A report by Thornton et al. (2000) summarizes the state of coastal processes research as of 1999.

Beaches

Many classifications exist to describe different types of coast. Van Rijn's (1998) classification consists of mud, sand, gravel/shingle, and rocky coasts. In this classification, approximately 10% of the coasts are mud and 15% are terrigenous sand and carbonate sand coasts. Because of their large societal importance, most studies of coastal change have been conducted on sand coasts, which typically are wave-dominated environments (*q.v.*).

The beach is the most prominent visual feature of sand coasts. It is the area of unconsolidated material that extends landward from the low-water line to the place where there is a definite change in material or physiographic form (e.g., a cliff or dune), or to the line of permanent vegetation (*Coastal Boundaries: q.v.*). Beach material can be any combination of sand (grain size = 0.0625–2 mm), granules (2–4 mm), pebbles (4–64 mm), cobbles (64–256 mm), and boulders (>256 mm) (*Sediment Classification: q.v.*). The grain size on most beaches is in the sand range.

Grain Size and Composition

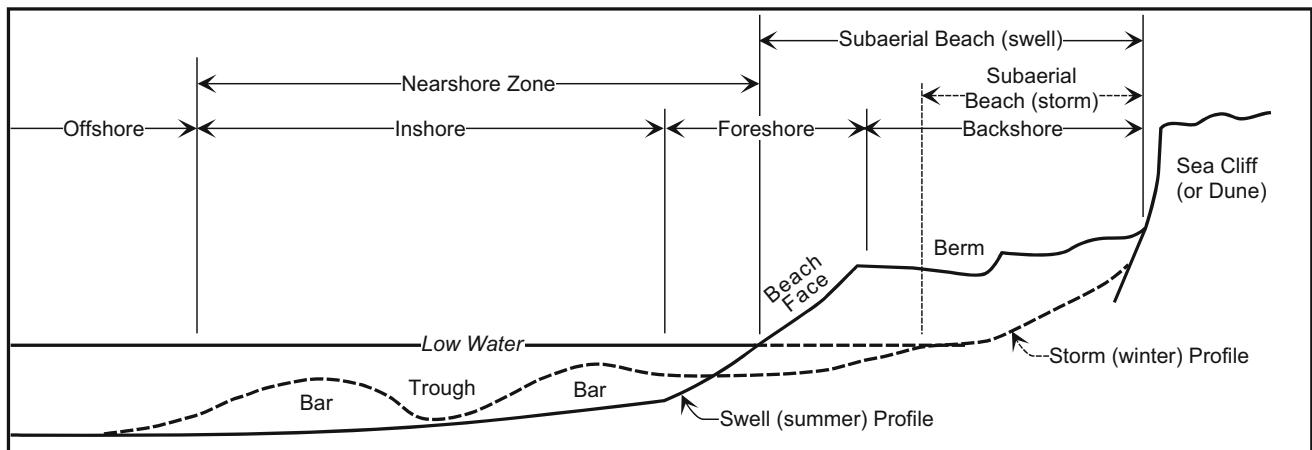
In temperate climates, beaches typically consist of quartz and feldspar grains derived from the weathering of terrestrial

rocks. Commonly, denser minerals (*heavy minerals*) that are specific to also occur in small percentages. On volcanic islands, the sand frequently includes lava fragments and associated minerals. Worldwide, many beaches have a calcium carbonate fraction from the breakup of shells, concentrations of foraminiferans, and nearby coral reefs (*q.v.*). For example, on the island of Hawaii, beach sand can be black, green, or white. The black sand comes from the erosion of lava beds and decomposition of hot lava flowing into the ocean; the green sand comes from the mineral olivine, which crystallizes when magma cools; and the white sand consists of calcium carbonate. Although calcium carbonate beaches are usually associated with the tropics, beaches in other climes also can have a large shell fraction.

Because the heavy-mineral fraction in beach sand is indicative of the provenance of that sand, heavy minerals can be used to trace the movement of sand along the beach. For example, Trask (1952) showed that the sand reaching the harbor at Santa Barbara, CA comes from the area around Morro Bay, CA, more than 160 km up coast. He established this by using the mineral augite as a tracer. Along the stretch of coast between the two cities, augite occurs in all beach sands, but the only source area is in the vicinity of Morro Bay. The concentration of augite decreases between Morro Bay and Santa Barbara as non-augite sources (e.g., local streams and cliffs) add to the sand moving along the coast.

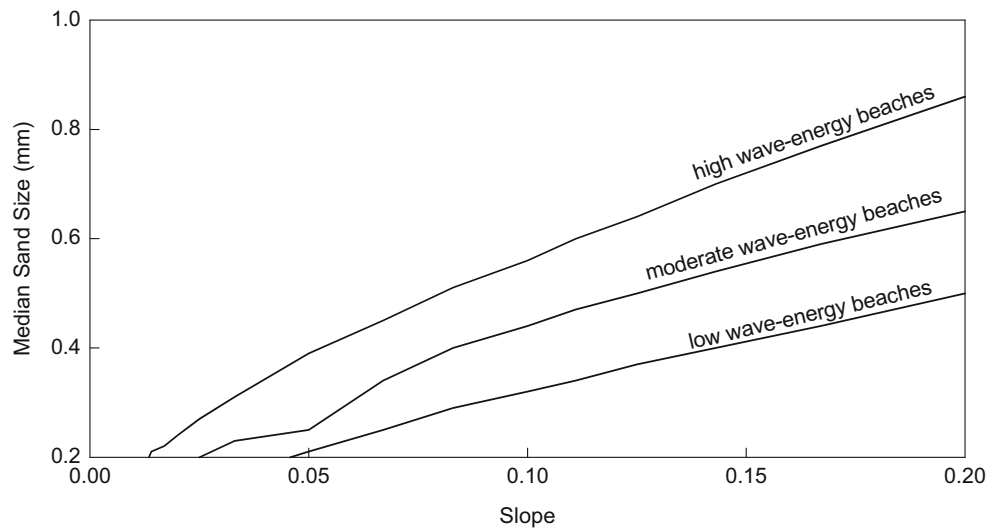
Beach Profile

Figure 1 shows a typical shore-normal cross section, or *profile*, of the coastal zone. The coastal zone can be divided into four major subzones: *backshore*, *foreshore*, *inshore*, and *offshore*. The locations of the beach subzones depend on whether the beach has accreted or eroded (*Beach Erosion: q.v.*). The offshore and inshore are seaward of the low-water line. The former is a relatively flat part of the profile seaward of the breaker zone. The latter includes the breaker and surf zones (*q.v.*). The foreshore is the sloping portion of the profile (typically 2–10°) and encompasses the normal intertidal part of the beach. On an eroding beach, the foreshore could cover the entire beach. The *beach face* is the upper portion of the foreshore that is normally exposed to wave *uprush*; it is often synonymous with the foreshore. On an accreting beach, the *berm crest* is the transitional area between the foreshore and backshore; often it is a striking feature several meters above low water, especially on cobble and boulder beaches (Guilcher 1958). The backshore extends landward from the berm crest to the edge of the beach; the nearly horizontal part is the *berm*, which forms from the deposition of lower-foreshore sediment that is transported over the berm crest. There can be more than one berm on a backshore. A nearly



Beach Processes, Fig. 1 Terminology for the coastal zone along a shore-normal profile

Beach Processes, Fig. 2 Beach slope as a function of sand size and wave energy



vertical escarpment (*scarp*) caused by erosion of an earlier berm crest can separate berms on multi-berm beaches.

The loss of beach sand usually corresponds to a gain of sand in the nearshore, and *vice versa*. Beach sand also can be lost to sanddunes, estuaries, and submarine canyons. Human activities can result in both beach loss and gain. *Pocket beaches* can be an exception where the sand often shifts alongshore with changes in wave climate. Visually, the result is a shift in the profile of the beach in both the horizontal and vertical. Although such changes are commonly associated with winter (erosion) and summer (accretion), in reality they occur any time of year in response to stormy and fair weather. *Storm waves* erode the berm and shift the shoreline (defined as the high-water line or wet-dry boundary) landward. *Swell waves* build up the berm and shift the shoreline seaward.

For sandy beaches, Wiegand (1964) developed a relationship between median grain size, slope of the beach face, and

wave climate using data from many US beaches. Grain size and beach slope were measured at mean tide level where the correlation was best (Bascom 1951). The relationship shows that the slope depends on two factors – grain size and wave exposure (Fig. 2). For any given wave climate, slope increases with increasing grain size. Correspondingly for a given grain size, the smaller the waves, the steeper the beach face. Thus, a beach face becomes flatter when eroding (larger, storm waves) and steeper when accreting (smaller, swell waves).

Beach Processes

A complete understanding of beach change requires an understanding of the processes active throughout the coastal zone. Accordingly, beach processes is a subset of *coastal processes*, or *coastal morphodynamics* (morpho-: form,

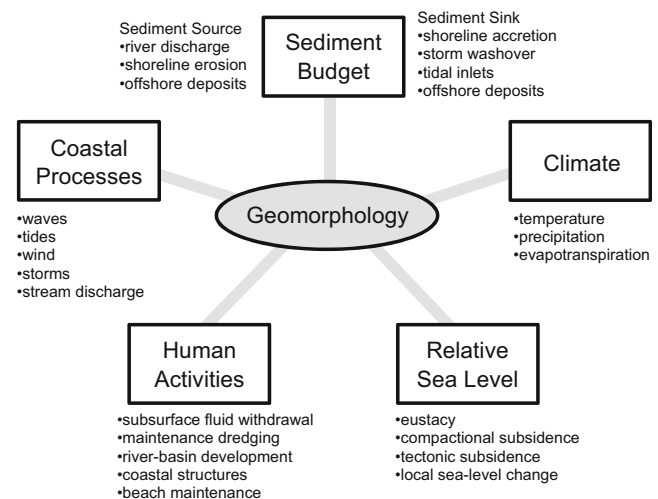
structure; dynamics: motivating or driving forces). Thus, coastal processes involve investigations of the interactions of coastal-zone features and hydrologic, meteorologic, and fluvial forces by means of sediment transport ($q.v.$). Coastal geomorphology and fluid dynamics couple at a continuum of temporal and spatial scales such that the fluid dynamics produces sediment transport, which produces geomorphic change. Progressive modification of the geomorphic features in turn alters boundary conditions for the fluid dynamics, which evolve to produce further changes in sediment-transport patterns and consequently the geomorphic features (Cowell and Thom 1994).

Coastal processes happen over a wide range of spatial and temporal scales; the upper limits are generally set at 10 km and 1 year, respectively. For example, the properties of waves entering the coastal zone from deep water and interacting with the nearshore profile determine the overall characteristics of nearshore waves and flows. However, small-scale processes control the turbulent dissipation of breaking waves, bottom boundary layer, and bedform processes that determine the local sediment flux. Cross- and longshore variations in waves, currents, and bottom slope cause spatial gradients in sediment fluxes resulting in changes due to erosion or accretion. Traditionally, the study of coastal processes has been restricted to small and intermediate scales (Thornton et al. 2000); making it but one of several influences on the coastal zone (Fig. 3).

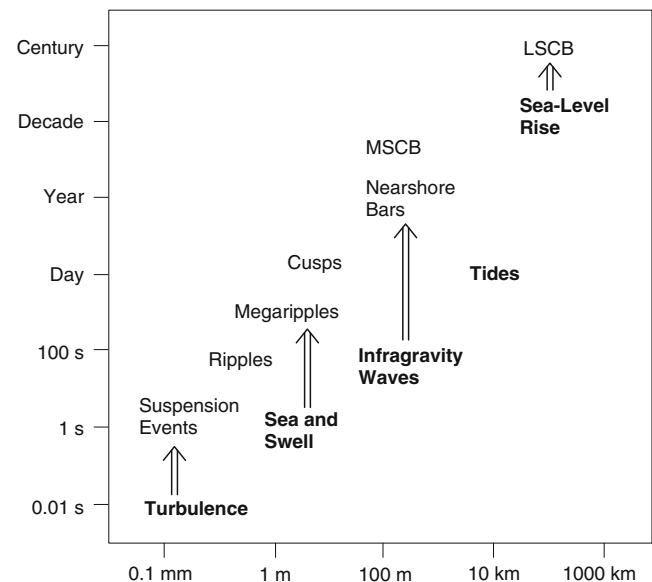
The rate of response of geomorphic features to the fluid dynamics depends on scale; larger features take relatively longer to change (temporal scale in Fig. 4). Hence, equilibrium is almost instantaneous for small-scale processes, and quasi-equilibrium becomes more noticeable as the geomorphic scale increases. For example, bedform scale and forcing history link the rate of transition between bedform types inside and just outside of the surf zone. Under large waves, significant changes in small-scale bedforms can occur within a single wave cycle, but changes in large-scale bedforms can exhibit significant hysteresis.

Wind, waves, tides, storms, and stream discharge are important driving forces in the coastal zone. Streams transport sediment from the hinterland to the coastal zone for the other forces to distribute. Wind directly transports beach sand and generates waves (*Meteorological Effects: q.v.*). Waves produce currents that transport sand cross-shore and longshore. Tides play a supporting role by exposing different parts of the beach face to waves and currents. Storms produce strong wind, waves, and elevated sea level, which can cause extensive coastal erosion; movement of sand to the nearshore or across barrier islands ($q.v.$) and spits ($q.v.$); and coastal flooding, as well as intensifying small- and intermediate-scale processes.

Waves are the major source of energy driving beach change in the nearshore. As a wave approaches the coast,



Beach Processes, Fig. 3 Processes that influence the geomorphology of the coastal zone (Thornton et al. 2000)



Beach Processes, Fig. 4 Temporal and spatial scales for coastal processes (Ruessink 1998). Fluid forces are boldfaced

it reaches a water depth where it begins to interact with the bottom (*shoal*). That depth occurs when the ratio of water depth to deep-water wavelength is less than 0.5, which is commonly in the depth range of 10–20 m. *Wave base* is the term Bradley (1958) gave to the depth at which normal wave erosion begins (also see Dietz 1963). *Depth of Closure* ($q.v.$) is the maximum depth of extreme bottom changes; it is a function of the nearshore storm-wave height exceeded 12 h per year and the associated wave period (Hallermeier 1981). Shoreward, a wave undergoes a systematic transformation (*shoaling*) where wavelength (L) decreases and height (H) increases (H/L is the wave *steepness*). When the ratio of the

wave height to water depth reaches 0.73–1.03 (Galvin 1972), the wave breaks, producing a beachward flow (*bore*). The result is an upward (*run-up*) and downward (*backwash*) flow of water on the beach face (the *swash zone*). If run-up reaches the back of the beach, it can erode cliffs, dunes, and structures.

Sand transport begins soon after waves begin to shoal. Transport volume and velocity increase shoreward in proportion to increasing intensity of the wave-bottom interaction. Whether there is beach erosion or beach accretion depends on wave height and period. Storms raise sea level by piling up water against the coast (*storm surge*) ($q.v.$), and greatly increase wave height and steepness (*storm waves*). Such waves tend to produce beach face erosion with the sediment being moved to the nearshore. Lower, less-steep waves (*swell waves*) produce accretion by moving nearshore sand onto the beach face. Storm and swell waves can occur any time of the year, although the former are more common in the winter and the latter in the summer. Consequently, the belief that erosion is a winter phenomenon (season of storm waves) and accretion a summer phenomenon (season of swell waves) is not completely correct. Basically, the wave climate continuously varies, and the beach face never reaches an equilibrium state where the volume of sand moving up the beach face equals that moving down.

Breaking waves create a circulation system where the water driven shoreward across the surf zone returns to the offshore via a strong, narrow flow called a *rip current* (Fig. 5) whose spacing ranges from tens to hundreds of meters. Velocities in rip currents are often strong enough to carry sediment and swimmers bathers alongshore through the breakers; in those cases, a distinct watercolor demarcates the rip current. Often the velocity is too strong to swim against, so people caught in a rip current must swim parallel to shore to escape.

If a wave enters the coastal zone at an angle to the bathymetric contours, its crest bends to align with those contours (*refraction*). If the wave breaks at an angle to the beach, a longshore current develops. Because sand grains move with

the flow, longshore sediment transport ($q.v.$) occurs (*littoral drift*). Because the sand grains are subject to both run-up and littoral drift, they follow a zigzag path along the beach face.

Geomorphic Features

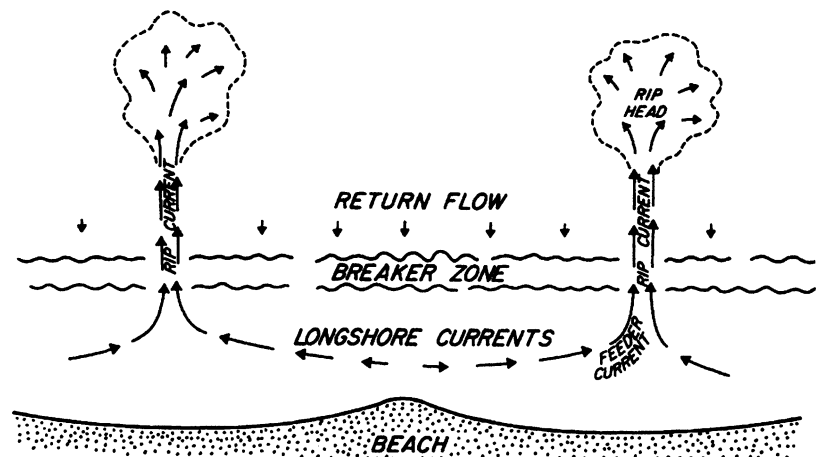
Based on the temporal and spatial scales of sediment transport and geomorphic change, the coastal zone can be divided into two cross-shore subzones – the upper and middle shoreface – and three longshore subzones – micro, meso, and macro cell. In the upper shoreface, breaking waves and bores generate active sand transport and rapid geomorphic response. In the middle shoreface, slow sand transport rates result in slow geomorphic change. Micro cells include smaller geomorphic features such as ripples and small beach-face features that change in times of a day or less. Meso cells include geomorphic features such as sand bars ($q.v.$), beach profiles, and beach cusps that change in a year or less. Macro cells extend kilometers and include large coastal geomorphic features (Fig. 4).

Micro-scale Beach Features

When waves begin to shoal, there is a back-and-forth water motion at the seafloor with onshore and offshore excursions being equal. The sediment moves the same way, and symmetric *oscillation ripples* form normal to the direction of wave advance. Nearer to shore, the water motion at the seafloor becomes asymmetric because the onshore component of the wave orbits is larger than the offshore one. These *current ripples* have a gentle seaward facing slope and a steep onshore one. Near the breaker zone and in much of surf zone, the bed is flat because of intense water motion. However, current ripples can form in the seaward flowing *rip currents*.

At the upper swash limit, deposition of debris forms scalloped, marginal lines known as *swash marks*. Backwash flowing around small obstacles form seaward opening “V”s,

Beach Processes, Fig. 5 The nearshore circulation cell. Breaking waves create a circulation system where the water-driven shoreward across the surf zone travels alongshore (*longshore current*) and returns to the offshore via a strong, narrow flow (*rip current*). (After Komar 1998)



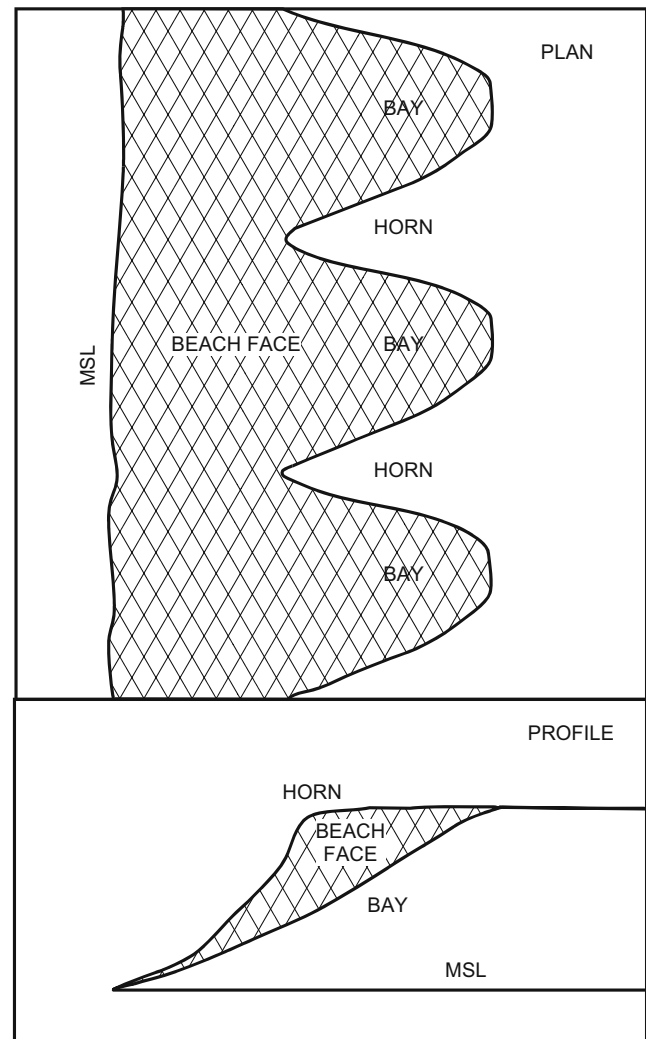
and in some cases, rhombic patterns develop as a result of the minor currents generated in the backwash. Seepage of interstitial water down the foreshore slope at low tide cuts miniature channels termed *rill marks*.

Meso-and Macro-scale Beach Features

Generally, sediment eroded from the foreshore and transported offshore forms one or more longshore bars with a trough shoreward of each one. The bars can extend alongshore for kilometers except for breaks caused by rip currents. Studies by Evans (1940), Keulegan (1948), King and Williams (1949), and Shepard (1950) concluded that there is a strong relationship between the bars and breaking waves. Keulegan, who studied bars formed in a laboratory wave channel, found that wave height and steepness govern the bar position. An increase in wave height moves the bar seaward (deeper water). Holding the wave height constant and reducing the wave period moves the bar shoreward. Starting with a smooth profile, the bar initially forms just shoreward of the breaker position. As it grows, both the bar and breakers migrate landward. When the bar is fully developed, it modifies energy transfer in the surf zone by reducing the amount of energy that the breaking waves can impart to the waves that reform in the surf zone. Shepard (1950) found that there can be no bar and trough for short-period storm waves because there is not a well-defined breaker zone.

Beaches can be two- or three-dimensional based on the linearity of the berm and beach face. On two-dimensional beaches, the berm crest and foreshore contours are straight and parallel. On three-dimensional beaches, the berm crest and foreshore have rhythmic, crescentic features (*Beach Cusps: q.v.*) that vary greatly in height and length. All cusps are characterized by seaward facing ridges or *horns* separated by *embayments* or *bays* (Fig. 6). Attempts to classify these features by size (e.g., Dolan and Fern 1968; Dolan et al. 1974) have been unsuccessful because there is a large overlap in sizes of cusps formed by different mechanisms. Komar (1983) developed a classification scheme with four types of cusps whose origins can be attributed to different processes of waves and currents within the nearshore (Fig. 7). These are reflective beach cusps, rip current embayment-cusp systems, crescentic bar-cusp systems, and transverse and oblique bars (see Fig. 7 for wavelengths).

There has been much disagreement as to the origin of reflective beach cusps. One recent theory is that *edge waves*, waves trapped at the shore with net motion in the longshore direction and decreasing amplitude offshore (Holman 1983), play a role in the formation of beach cusps. In a laboratory wave tank, Guza and Inman (1975) found that beach cusps developed in response to edge waves. In the field, Huntley and Bowen (1978) observed the formation of cusps in the presence of edge waves. Werner and Fink (1993) proposed

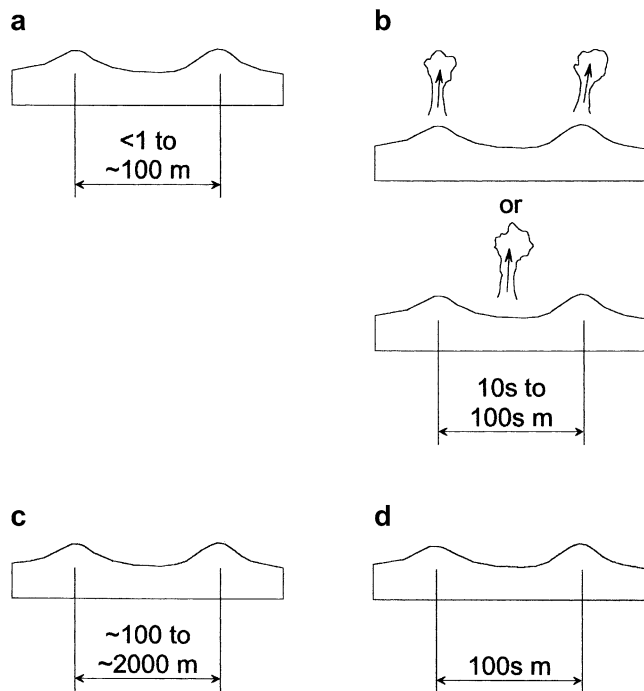


Beach Processes, Fig. 6 Plan and profile views of a cusped beach

that beach cusps form through strong, positive feedback between wave run-up and beachface topography.

A rip current embayment-cusp system develops when rip currents erode the beach face creating embayments. The cusps are midway between the embayments (Bowen and Inman 1969). The cusps correspond to positions of zero longshore sediment transport produced by waves combining with the feeder currents that flow alongshore toward the rip currents (Fig. 5). In some cases, rip current erosion can be so extensive that the embayments cut across the beach, exposing fore-dunes and cliffs to wave attack (Komar and Rea 1976; Komar and McDougal 1988). Beach sediment can be deposited in the lee of a rip current so that the cusps correspond to the rip locations. This appears to occur most commonly on steep beaches under relatively low wave conditions (Komar 1971).

Crescentic bars are rhythmic lunate features with uniform spacing primarily found underwater, commonly on long



Beach Processes, Fig. 7 Classification of rhythmic shoreline forms: (a) reflective beach cusps; (b) rip current embayment cusps; (c) crescentic bay cusps; (d) transverse and oblique bars. (After Komar 1998)

straight beaches (Shepard 1952; Homma and Sonu 1963). They appear to be confined to regions of small- to medium-tidal range (Bowen and Inman 1971) and form best where the beach slope is low. The generally accepted mechanism for their formation is by edge waves in the infragravity range (Bowen and Inman 1971).

Transverse and oblique sandbars are nearshore features that do not parallel the beach, and they are welded to the beach face. There are various explanations for their formation that include the rotation of rip-current segmented bars (Sonu 1972), processes akin to the migration of river bars or the development of river meanders (Bruun 1954; Sonu 1969; Dolan 1971), and the superposition of edge waves (Holman and Bowen 1982).

Beach Change

Beach Cycles

Various cycles affect the beach, depending essentially on changes in wave steepness and effective sea level. In regions of tidal action, the swash-backwash zone and breaker zone shift landward and seaward with the flood and ebb tide. The range of the tides and the slope of the foreshore determine the distance over which the shift takes place. With other parameters remaining constant, breaker height will be greater at high than at low tide. At high tide, the prevailing waves approach less hindered over the relatively deeper water of the

nearshore, whereas at low tide, approaching waves are modified by the shoaler depths of the gently sloping nearshore bottom. Under this changing regime, scour may be slightly increased at high tide.

At approximately 7.5-day intervals, the tidal range changes from minimum (neap) to maximum (spring). The change from neap to spring tides can produce upper-foreshore erosion with lower-foreshore and nearshore deposition (LaFond 1938; Inman and Filloux 1960). LaFond and Rao (1954) postulate the cumulative result of higher effective sea level, higher waves, and a time lag in recovery during the low spring tide cause the redistribution of sand. The opposite sand movement can occur during neap tides.

Effects of Beach Erosion

Erosion strips sediment from the beach face and moves it to the nearshore or redistributes it alongshore. When enough sand is removed, there is no longer a high, wide beach, and waves can attack coastal features such as cliffs, dunes, and anthropogenic structures. In some places, storms transport beach sand across spits and barrier islands and deposit the sand in adjacent lagoons. If the sand returned to the beach is less than the volume eroded, the beach narrows, and if possible, the shoreline shifts landward.

Shoreline retreat is a natural process that is of little or no concern in unpopulated areas. However, in populated areas, shoreline retreat is a major issue. Several years can pass between storms severe enough to cause significant damage to a stretch of coast. Consequently, many people build or buy homes and other facilities on the coast with the idea that the adjacent beach is permanent. Later they watch storm waves remove the beach sand and directly attack their property or the coastal cliffs and dunes that protect them. Then, affected communities quickly want to know how to save their beaches and protect their homes and facilities. Although beaches usually rebuild after storms, a beach does not always return to its pre-storm position, and the community must take remedial measures to reverse long-term shoreline retreat.

Because storm waves threaten coastal facilities when fronting beaches lack sand, post-storm beach accretion is essential to minimize economic loss in the coastal zone. In areas where there is insufficient beach to protect coastal structures, there are several procedures to prevent or mitigate shoreline retreat. Traditionally, these procedures require building a protective structure (*q.v.*) on the beach or at the landward edge of the backshore. These include *seawalls*, *revetments*, *groins*, and *breakwaters*. While these structures often protect the property behind them, the fronting beach typically does not return because increased water turbulence at the structure prevents sand deposition during swell conditions (see Dean 1999 for examples). The result is a section of coast with no beach, and if longshore sand transport is not properly taken into account, the shoreline downdrift of a

structure also can lose its sand and retreat. Furthermore, structures often fail if improperly designed, allowing coastal retreat to resume.

Beach nourishment (*q.v.*) is another technique used to prevent shoreline retreat by augmenting the native beach sand with sand imported from other areas. Although beach nourishment creates wide beaches, this technique may not provide a long-term solution to beach loss especially where erosion rates are high or there is a persistent problem. A major problem is that the cost of importing sand can be high, especially since the sand should be similar in character to the native sand and because more sand is frequently needed after a storm season. When this technique is successful, there will be a year-around beach for public use and shoreline protection.

Other techniques include relocating coastal structures to allow for shoreline retreat and defining setback lines for coastal development. Shoreline retreat permits nature to take its course, but often is infeasible in populated areas. Setback lines, which are based on historical shoreline retreat rates, need to be implemented before coastal development begins.

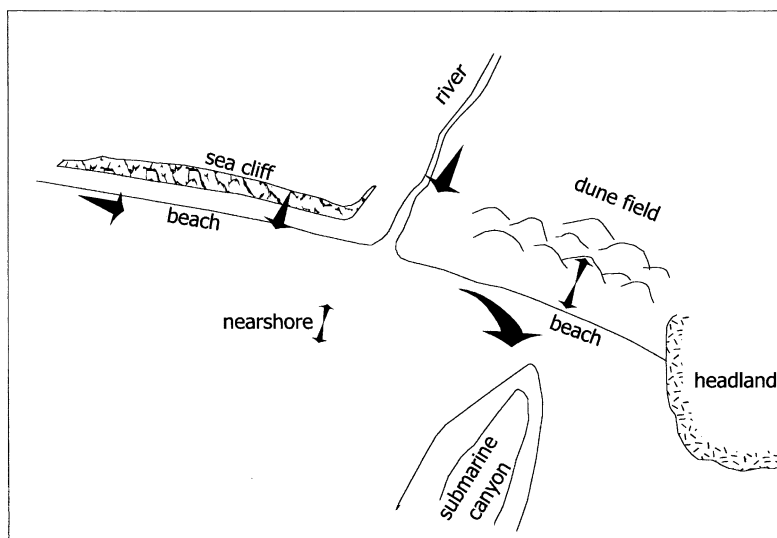
Sediment Budget and Littoral Cell

The budget of littoral sediments is simply an application of the principle of conservation of mass to the littoral sediments – the time rate of change of sand within the system depends upon the rate at which sand enters the system versus the rate at which it leaves. An analysis, therefore, involves evaluations of the relative importance of various sediment sources and losses to the nearshore zone, and a comparison of the net gain or loss with the observed rate of beach erosion or accretion.

The budget of littoral sediment involves making assessments of the sedimentary contributions (credits) and losses (debits) and equating these to the net gain or loss (balance of sediments) in a given sedimentary compartment or littoral cell (Bowen and Inman 1966; Komar 1996). The balance of sediments between the credits and debits should be approximately equivalent to the local beach erosion or accretion. Table 1 summarizes the possible credits and debits of sand for a littoral sedimentary budget, while some of the more important components are diagrammed in Fig. 8. In general, the longshore movement of sand into a littoral compartment, river transport, and sea-cliff erosion provide the major natural credits; longshore movement out of the compartment, off-shore transport (especially through submarine canyons), transport into estuaries, and wind transport shoreward to form sand dunes are the major debits. Included in Table 1 are the major human-induced credits and debits, including beach nourishment, which is increasingly used to rebuild lost beaches, and mining (*q.v.*), which directly removes sediment from the nearshore.

Research Techniques

More is known about the geomorphology of the coastal zone than about the processes that modify the geomorphology. For example, it has only been during the past decade that the coupling between waves, circulation, and changes in near-shore bathymetry has begun to be observed and modeled. In addition, research is now focusing on fluid velocities and particle flux profiles in the bottom boundary layer and in the surface boundary layer under breaking waves.



Beach Processes, Fig. 8 Principle constituents of a littoral zone sediment budget in a natural setting. Size of arrows is a rough indication of relative importance, though they can vary for a given situation. Not shown are anthropogenic impacts such as sand mining. (After Komar 1998)

Beach Processes, Table 1 The budget of littoral sediments

Credit	Debit
Longshore transport into the area	Longshore transport out of the area
River transport	Wind transport away from the beach
Sea cliff erosion	Offshore transport
Biogenous deposition	Solution and abrasion
Hydrogenous deposition	Mining
Wind transport onto beach	
Beach nourishment	

Studies of velocity and sediment concentration measurements in the swash zone are moving forward with the development of new instruments.

At present, it is not possible to forecast the effect of an upcoming storm season on a section of coast. However, it is possible to ascertain both the ultimate storm profile and the rate at which a beach returns to its original profile or shifts to a new equilibrium profile. The principle method for obtaining quantitative data on beach change is to repeatedly survey a beach. Comparing the resulting profiles will give erosion and accretion rates for the time encompassed by the surveys.

Geomorphology

Various surveying techniques are available to determine the geomorphic character of the coastal zone. Offshore, a boat-mounted depth sounder can be used to measure the bottom profile (*Bathymetric Surveys: q.v.*), and a side-scan sonar can be used to collect oblique views of the bottom. Both of these techniques are limited to depths where boats can safely operate, which does not include the breaker and surf zones. Furthermore, their accuracy is limited because of boat movement by waves and currents.

Vehicles have been developed to measure beach profiles across the nearshore. These include remotely controlled tractors (Seymour et al. 1978; Dally et al. 1994), sleds (Sallenger et al. 1983), and an 11-m high motorized tripod that can drive across the beach and into the nearshore (Birkemeier and Mason 1978). MacMahan (2001) mounted an echosounder and a *global positioning system* (GPS) (*q.v.*) unit onto a waverider to profile from deep water through the surf zone.

Surveying with a rod and transit is the most common method used to obtain beach profiles. Techniques range from the “Emery Board” procedure that uses two wooden rods separated by a rope, to sophisticated instruments that use light beams to measure the distance to a prism. The vertical accuracy of the latter instruments can be less than 1 cm. With all these instruments, however, a closely spaced grid of points can be difficult to achieve except in small beach areas. When a beach is two-dimensional, a single cross-shore survey is sufficient to characterize it, but a beach with cusps and other three-dimensional features requires multiple cross-

shore and, in some cases, alongshore surveys. With temporal surveys along a fixed shore-normal line, the existence of cusps can cause errors in comparing beach volume and beach-face location. At present, kinematic GPS units mounted on survey rods and various kinds of vehicles are being used to rapidly survey large sections of beach (Kaminsky et al. 1998).

Remote sensing techniques can be used to study large sections of the subaerial part of the coastal zone. Some of the techniques also can be used to look at geomorphic features in the nearshore. Air photos provide a qualitative look at the geomorphology, and quantitative measurements are possible with images that can be orthorectified. Such images are especially useful for measuring cliff retreat. Plant and Holman (1997) measured intertidal beach shape using a combination of time exposures and differential GPS. Lippmann and Holman (1989) investigated the locations and forms of offshore bars by taking several-minute time exposures from a camera mounted high above the beach and looking longshore. Recently, LIDAR (light detection and ranging) (*q.v.*), an airborne scanning instrument, has been used to map the subaerial part of the coastal zone (Brock et al. 1999) and shallow parts of the nearshore (Irish and Lillycrop 1999). These instruments are capable of rapidly estimating elevations to within a couple of centimeters approximately every 3 m² over regional scales.

Physical Processes

Many instruments are available to measure water and sediment motion at all scales throughout the coastal zone. Many of them are sturdy enough to withstand the forces generated in the breaker and surf zones. Pressure sensors and bidirectional current meters have a sampling rate fast enough to measure water depth and wave-generated currents, respectively. Optical sensors measure sediment concentration. Sonic devices measure rapid changes in bottom elevation and, in some cases, the height of sediment suspension.

Future Research in Coastal Processes

The goal of future research in coastal processes includes developing predictive models for:

- bedload and suspended sediment transport under combined wave and current forcing;
- turbulent wave/current boundary layers over 3-D small-scale morphology;
- effects of moving sediment on boundary layer;
- contribution to sediment transport by bedform migration;
- effects of grain-size distribution on sediment transport (Thornton et al. 2000).

Morphology and its variability are important end products of predictive models. However, because sediment transport is not well understood, prediction of morphological change is inadequate at all scales. For example, at smaller scales, ripples and megaripples are observed to be ubiquitous, but have not been incorporated into models even though their effect on the flow field (as roughness elements) and sediment transport may be significant. Complex patterns in long-term, large-scale morphology have also been observed. However, models for morphology change have predictive skill only over the short term, whereas long-term, large-scale predictions are not yet possible. Research issues include:

- predicting morphology across the spectrum of length scales;
- free versus forced large-scale morphology models;
- understanding feedback between morphology and the flow field;
- coupling between length scales.

Cross-References

- [Barrier Island Landforms](#)
- [Bars](#)
- [Beach Erosion](#)
- [Beach Features](#)
- [Beach Nourishment](#)
- [Beach Profile](#)
- [Changing Sea Levels](#)
- [Cliffed Coasts](#)
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- [Wave-Dominated Coasts](#)
- [Waves](#)

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Beach Profile

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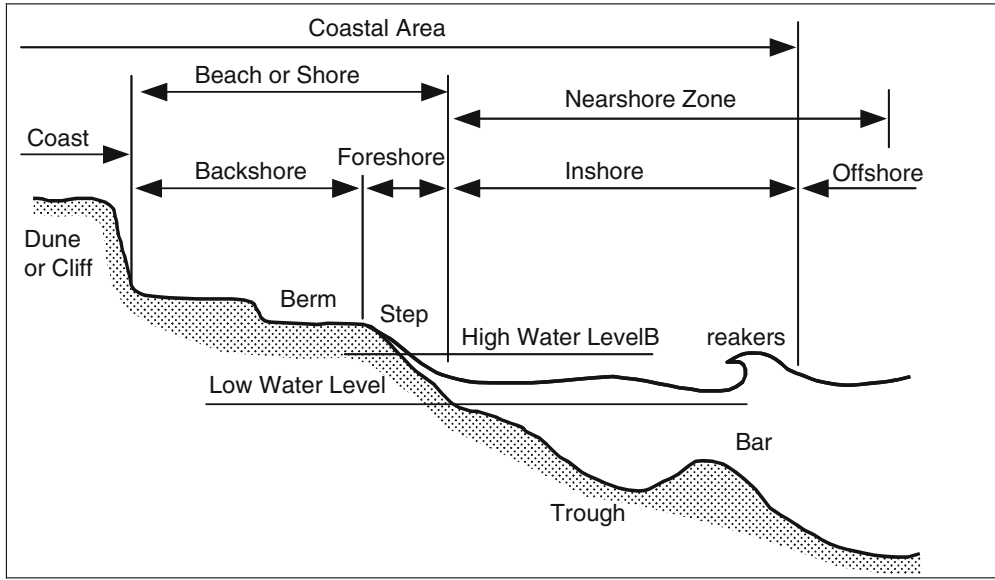
The beach profile is one the most studied features of coastal morphology. The shape of the beach profile determines the vulnerability of the coast to storms, the extent of usable beach for habitat and recreation, and the legal boundary distinguishing public and private ownership of land (Shalowitz 1962, 1964; Anders and Byrnes 1991). The first modern studies of the beach profile were motivated to understand its shape and variability in support of amphibious operations during World War II, when personnel and supply boats had to cross the beach profile from offshore to the dry beach (Bascom 1980).

Beach Profile Terminology

The term “beach profile” refers to a cross-sectional trace of the beach perpendicular to the high-tide shoreline and extends from the backshore cliff or dune to the inner continental shelf or a location where waves and currents do not transport sediment to and from the beach. The profile shape is variable, depending on the time of year within the annual beach cycle and, also, the elapsed time after a storm. Waves, water level, and sediment grain size are the main controlling factors of beach profile shape.

Terminology associated with the beach profile is shown in Fig. 1. The backshore runs from the seaward-most dune or the cliff to the land and water intersection. One or more berms

Nicholas C. Kraus: deceased.



Beach Profile, Fig. 1 Terminology associated with the beach profile

may appear on a beach, depending on seasonal changes in water level. Berms are flat areas created during times of accretionary wave conditions, typically during summer. The beach intersects the water at the foreshore, and the foreshore is typically a plane slope that extends over a water level range from low tide to high tide. During a storm, a vertical step or scarp may form on the berm. The inshore covers the surf zone from the seaward end of the foreshore to past the seaward-most longshore sand bar, joining to the offshore. Several bars and associated troughs may appear on the beach profile.

Approximations of Beach Profile Shape

As a first approximation, it is often possible to represent the profile of a gravel, pebble, or sandy beach (Here, the profile is assumed to have an unlimited supply of sand and that no “hard bottom” is present such as limestone reefs, coral reefs, and other non-erodable (hard) substances.) by a straight line of constant slope as,

$$h = x \tan \beta, \quad (1)$$

where h is the still-water depth, x the distance from shoreline, and $\tan \beta$ is the beach slope. This expression, defining a “plane beach” or planar beach slope has convenience for making simple calculations. Typically, however, the foreshore is the only area of the beach profile well represented by a straight line.

A more realistic representative profile shape was introduced by Bruun (1954) and studied extensively by Dean

(1977, 1991). This profile is called the “equilibrium” or “x to the two-thirds profile” and is given as,

$$h = Ax^{2/3}, \quad (2)$$

where A is the shape parameter and will be discussed below. For water with temperature of about 20 °C and typical sand sizes with sediment all speed varying between about 1 and 10 cm s⁻¹, Kriebel et al. (1991) found that A could be related to fall speed w by,

$$A = 2.25 \left(\frac{w^2}{g} \right)^{1/3}, \quad (3)$$

where g is the acceleration due to gravity. Moore (1982) was the first to study the functional dependence of A and found it to be an increasing function of the median grain size d_{50} for a wide range of materials. The empirical curve can be approximated by,

$$\begin{aligned} A &= 0.41(d_{50})^{0.94}, & d_{50} < 0.4 \\ A &= 0.23(d_{50})^{0.32}, & 0.4 \leq d_{50} \leq 10 \\ A &= 0.23(d_{50})^{0.28}, & 10 \leq d_{50} \leq 40 \\ A &= 0.46(d_{50})^{0.11}, & 40 \leq d_{50} \end{aligned} \quad (4)$$

for which d_{50} is expressed in millimeters. The equilibrium profile can encompass a large range in grain size, as seen by the values in Eq. 4. Because the shape parameter A increases with increasing sediment grain size or fall speed, finer-grained beaches have gentle slopes and coarser-grained beaches are steeper, in accord with observations (Bascom 1980).

Bars and Troughs

Bars and troughs, also called longshore bars and longshore troughs, are the major perturbations from the equilibrium profile. Typically, areas with small tide range possess the most prominent bars because the wave breaker line remains in one position longer. Multiple-barred beaches are common on the Great Lakes, bays, and the Gulf of Mexico coast of the United States, for example, where the tide range is small. Some of these bars are formed by various predominant waves, such as typical waves and storm waves. Likewise, if the wave conditions occupy a relatively narrow range of height and period, such as on the north shore of Long Island, New York (facing the Long Island Sound), bars tend to be more prominent as compared to bars on the south shore of Long Island, because the Atlantic Ocean has a much more variable wave climate and smears out such bottom features.

Larson and Kraus (1989) analyzed large-scale laboratory data for breaking waves and sandy beach beaches and found that the depth over the crest of a bar h_c was related to the breaking wave height H_b as

$$h_c = 0.66 H_b. \quad (5)$$

Conversely, if the depth over the crest of a well-established bar is measured, the breaking wave height that created the bar may be inferred from Eq. 5 to be $H_b = 1.5 h_c$.

The beach profile at North Padre Island, TX, located along the Gulf of Mexico, was surveyed in the mid-1970s in one of the earliest applications of a sea sled, and then again in the mid-1990s with a sled, assuring high accuracy (see “► Beach Profile”). Figure 2 plots a time series of surveys made at the same location on the beach. In both eras, profile elevation was referenced to mean sea level

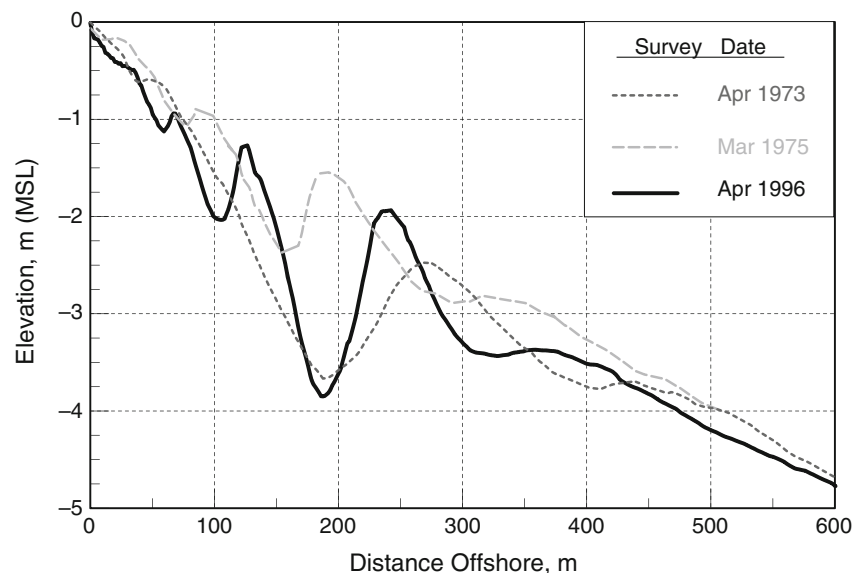
(MSL) at a local ocean tide gauge. Although the beach may have advanced or receded during the two decades, the shape of the profile can be compared because of the common vertical datum.

Figure 2 indicates that one to three (occasionally four) bars can appear on the profile at North Padre Island. It can be estimated through Eq. 5 that these bars are related to different classes of waves as outer bar-severe storm waves; middle bar-typical storm waves; and inner bar-waves under normal Gulf of Mexico conditions. In Fig. 3, the average of 18 beach profile surveys taken alongshore at North Padre Island in 1996 is plotted together with the equilibrium ($x^{2/3}$) profile, Eq. 2, determined by the median grain size in the surf zone (0.18 mm). Sediment sampling performed during the surveys demonstrated a decrease in grain size with distance offshore, and such a decrease in size of surficial sediments is typical along the beach profile, with the coarsest material located at the beach face and bars, and the finer material located offshore. Coarse material is also found at the landward sides of longshore bars, because the finer material is transported away from this hydrodynamically energetic area.

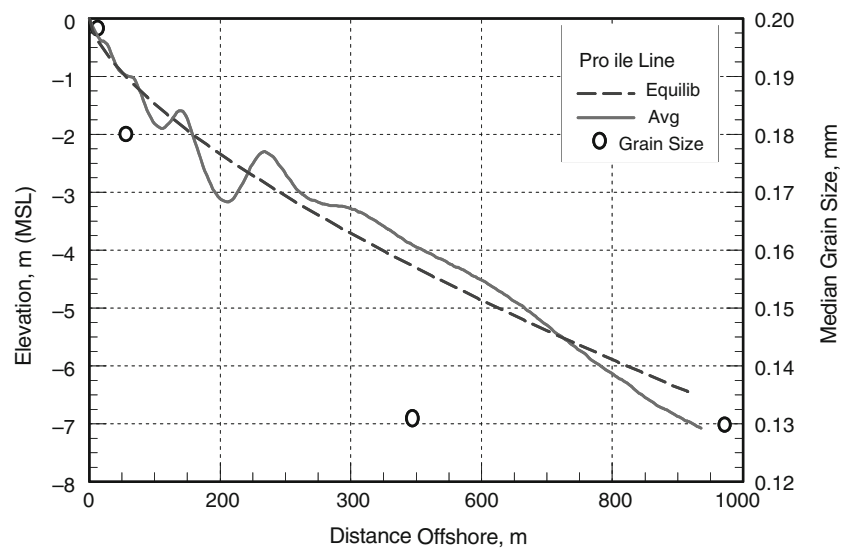
Seasonal Characteristics of the Beach Profile

During winter and the occurrence of seasonal storms (periodic northeasters, tropical storms, hurricanes), waves and cross-shore currents remove material from the beach and deposit it in bars far offshore. Large volumes, for example, $30\text{--}100 \text{ m}^3 \text{ m}^{-1}$ width of beach, can be removed from the beach berm and dune in a single large storm. Whether a beach will erode or accrete and the bar move onshore or offshore can be estimated with a dimensionless parameter called the Dean number formed as $N = H/wT$, where H is

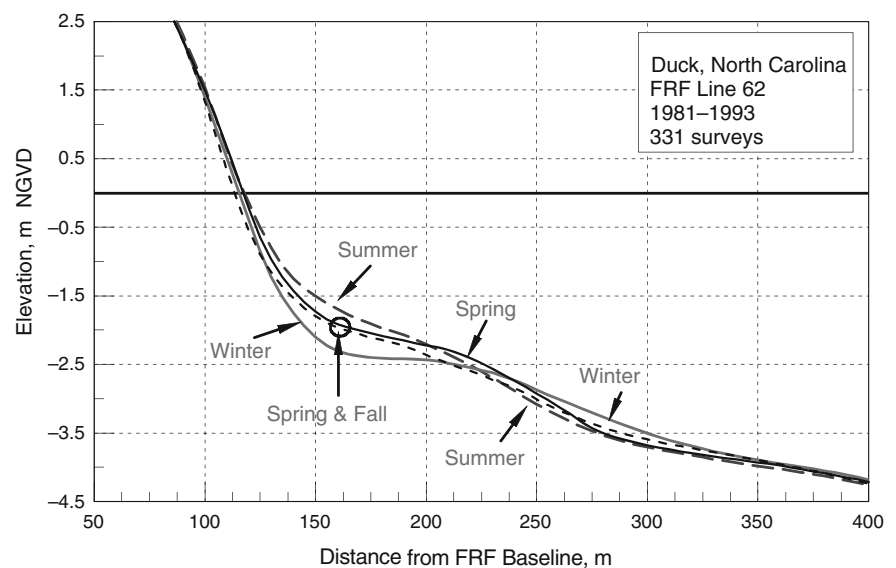
Beach Profile, Fig. 2 Beach profile surveys taken by sled two decades apart, North Padre Island, TX



Beach Profile, Fig. 3 Average beach profile and equilibrium profile, North Padre Island, TX, 1996



Beach Profile, Fig. 4 Average seasonal profiles, Duck, NC



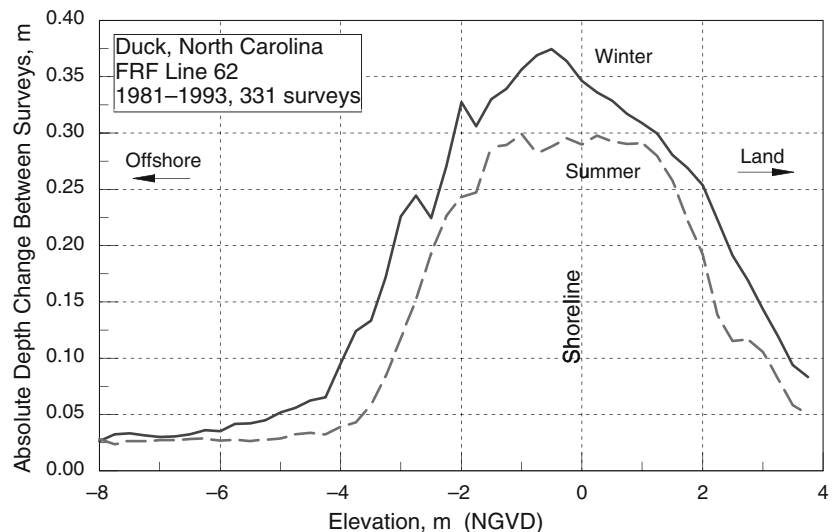
the wave height in deeper water, w is the beach sediment fall speed, and T is the wave period. For $N > 3.2$, erosion is probable, whereas for $N < 3.2$, accretion is probable (Kraus et al. 1991). Large values of the Dean number tend to occur during storms, when wave heights are large, and small values tend to occur in the summer, when wave heights are small or after storms, when the wave period of the swell waves becomes long. During the following milder accretionary waves of summer, material gradually moves onshore and returns to the beach, deflating the bar(s) and building the berm. Wind-blown sand then gradually rebuilds the dunes.

Sometimes, storms can be so severe that the beach does not recover on the timescale of human lifetime or engineering projects. Such is probably the case on the south shore of Long Island for the “Great New England Hurricane” of September 1938, which weakened the barrier islands and caused many

breachings or cutting of temporary inlets. Sand taken offshore by such strong storms lies in such deep water that summer or “recovery” waves cannot readily transport it back to the beach.

The seasonal averages of a large number of surveys made on the same cross-shore transect (Line 62) are plotted in Fig. 4. The surveys were made at the US Army Corps of Engineers’ Field Research Facility (FRF), in Duck, NC, located on the “Outer Banks” barrier island chain. The profile is surveyed every 2 weeks as routine monitoring or more frequently for specific research goals by means of a large motorized tripod, estimated to have a vertical accuracy of ± 2 cm. The National Geodetic Vertical Datum (NGVD) is close to MSL at the FRF. Bars are absent from the plots because the average is taken over a large number of surveys. Sand moves offshore in winter (arbitrarily defined as the

Beach Profile, Fig. 5 Average absolute depth change for summer and winter surveys, Duck, NC



interval January to March) and returns in summer (June–August). A broad hump in the winter average at about 2–4 m depth indicates the presence of storm bars during that season. During summer months, the steep profile in shallower water created by the winter waves is gradually replenished and becomes shallower.

One property of the beach profile observed in Fig. 4 is that the spring (April–June) and fall (October–December) average profiles almost plot on top of one another, and in between the two terminal states of summer and winter (Larson and Kraus 1994). The regularity in profile response to waves indicates that the processes should be predictable with relatively simple techniques. Although not shown, the average of all profiles corresponds well with the equilibrium profile with a median grain size of 0.2 mm.

The seasonal response of the profile is shown in another way in Fig. 5, which plots the average change in depth irrespective of sign (absolute value) as a function of average depth for the winter and summer profiles. The change in depth is less in summer than in winter, which is intuitively reasonable because waves are smaller in summer. The average maximum depth change occurs in winter, near the shoreline, and is about 0.5 m (1.5 ft). The location where the profile changes little through time, delineating the area where sediment in the offshore is no longer exchanged with the beach, is called the depth of closure (Kraus et al. 1999). Knowledge of how much the profile elevation changes is required for modeling coastal processes, for designing beach fills, for placing pipes, outfalls, and cables, across the surf zone, and for placement of instruments so that they do not become buried.

Cross-References

- [Accretion and Erosion Waves on Beaches](#)
- [Bars](#)

- [Coastal Warfare](#)
- [Depth of Closure on Sandy Coasts](#)
- [Meteorological Effects on Coasts](#)
- [Monitoring Coastal Geomorphology](#)
- [Storm Surge](#)
- [Surf Zone Processes](#)
- [Tidal Datums](#)

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Beach Ridges

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Definitions

Beach ridge is a relict, no more active wave-, often (also) wind-constructed shore ridge, with its long axis generally parallel with the coastline and if part of a strandplain, (also spelled as strand plain or strand-plain) with landward-adjacent similar ridges (Fig. 1). It is built up by berm-overlapping swash. Frequently, it is surmounted by a foredune. Regardless of their dimensions, shapes, and origin, active shore ridges, before becoming inactive (relict) beach ridges, continuously impacted and modified by wave processes, are excluded from the beach ridge designation (Otvos 2000). Once a ridge is isolated from the foreshore zone, often protected and preserved by dune cover or by its coarse sandy-to-gravelly nature, it is separated from daily beach processes by shore progradation and becomes a beach ridge. Shore-parallel dune ridge sets in eolian backshore plains may also form from a line of embryonic dunes behind a still active foredune. Although relict foredunes do not necessarily mark previous shoreline positions, their beach ridge designation also appears to be appropriate. Beach ridge plains develop in response to “normal” (sediment supply-related shore progradation) or “forced” (sea-level drop induced) regression.

Johnson (1919) described beach ridges as depositional features constructed by waves at successive shore positions. Reineck and Singh (1975) characterized beach ridges as having been formed at high-tide level of “rather coarse sediment” and related to storms or exceptionally high water stages. Bates and Jackson (1980) designated beach ridges as low mounds of beach and beach-and-dune material, heaped up by waves over the backshore beyond the present limit of storm waves or ordinary tides, but there is a risk of confusing wave-deposited and wind-deposited ridges. Referring to relict strandplain dunes as beach ridges, Carter (1986) used the term very broadly to cover “all large constructional forms of the upper

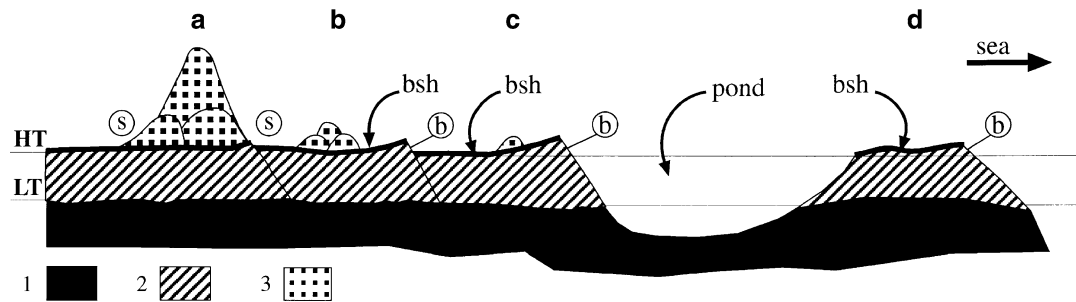
beach, capable of preservation,” applying it also to landward-shifting offshore bars, already welded swash bars, and prograded “berm ridges.” Davis et al. (1972), Fraser and Hester (1977), Carter (1986, 1988), and others included onshore-migrating swash bars and/or their stranded end products in the beach ridge category. For some time after welding to the shore, the prograding sandy berm ridges may continue to be impacted by active shore processes (Fig. 2).

Associated Landforms: Berm and Berm Ridges Versus Beach Ridges

“Berm,” not to be mistaken for beach ridge, originally was defined as a narrow, scarp-backed, and wave-cut horizontal surface in the beach foreshore (e.g., Komar 1976). Subsequently, berm came to mean a wedge-shaped ridge, between an upper foreshore slope and a landward-inclined berm top surface (King 1972). Its base is the horizontal plane that intersects the foreshore slope at the level of the backshore plain. Hine (1979) described berm as a shore-parallel linear body of triangular cross section with a horizontal to slightly landward-dipping surface (berm top) and a steeper seaward-dipping slope (beach face). Formed on mainland or island beaches or on shore-parallel sand spits, berm ridges are bracketed by the foreshore and the landward (lagoonward) margin of the backshore. On the landward side they may follow the shoreline of an elongated lagoon or beach pond (“cat’s eye pond”; Coastal Research Group 1969; Hayes 1979), enclosed by an ephemeral sand spit (Fig. 1b). Erosional berms replaced by several fair-weather accretional berms during wave run-up on mixed sandy-gravel beaches.

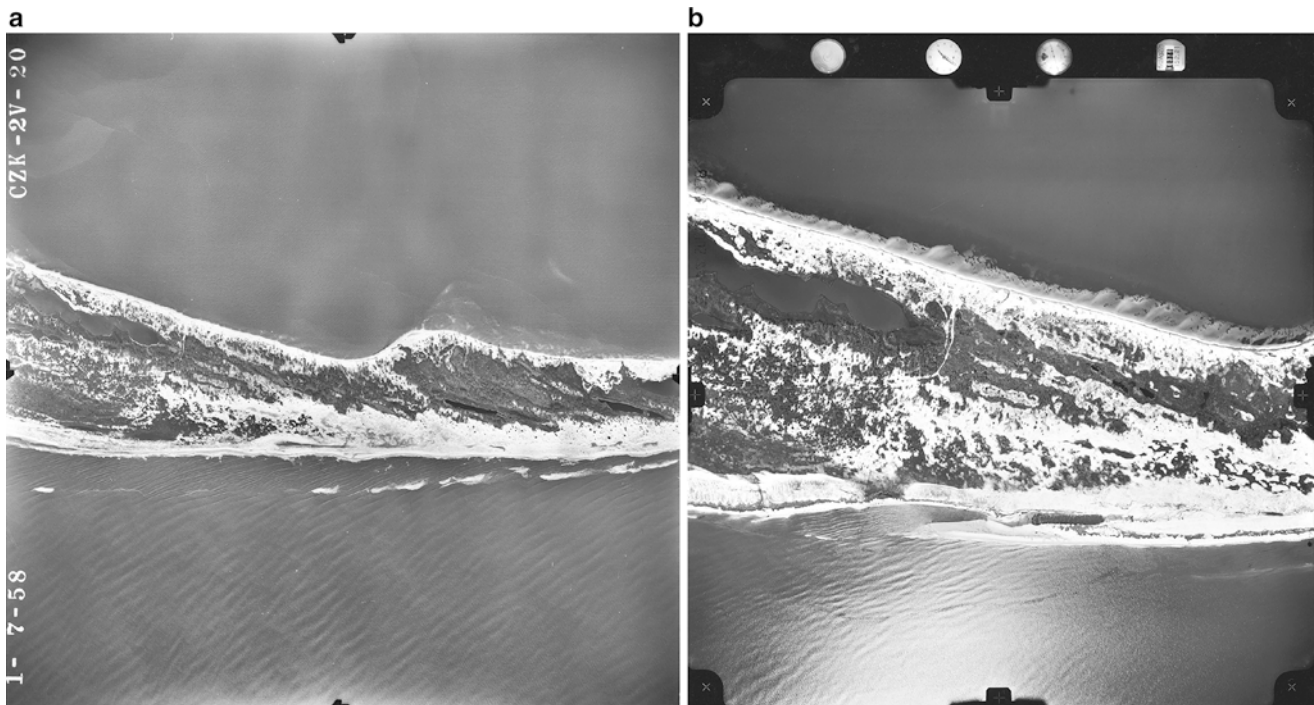
Generally, berms are short-lived and frequently reforming landforms. Swash currents deposit sediment that builds the landward-sloping high-tidal berm above the level of the adjacent backshore. Ephemeral high-tidal fine sandy berms are of aggradational origin, with secondary indications of erosional scarping. Increased onshore winds during falling tides briefly stabilize the water level. Intermediate-level berms with vertical scarplets may form during stillstands. “Berm ridges,” occasionally sizable and more permanent than berm surfaces, are wave-built intertidal-supratidal landforms. They are composed of intertidal and high-tidal (swash-overwash) deposits, bounded by the backshore plane and the berm surface along its foreshore margin (Fig. 2).

Sets of prograding “berm ridges” may form on beaches of limited sand supply. Short et al. (1989) reported on a nearly exclusively swash-built beach ridge plain in Australia. In Egypt, Goodfriend and Stanley (1999) described a shelly sand ridge plain, composed of 20–30 cm high, wave-built ridges without eolian cover. Most North American and Australian authors considered stabilized onshore beach ridges as either of predominantly wave-built or of wave- and wind-constructed, composite



Beach Ridges, Fig. 1 Beach ridge-associated depositional facies and landforms on a prograding strandplain. Depositional facies: (1) subtidal; (2) wave-built, intertidal-to-supra-high tidal; (3) eolian. Landforms: *b* berm, *bsh* backshore plain, *s* swale, (a) foredune; (b) accreting embryonic dunes on backshore plain in the early foreshore stage; (c) single

embryonic dune on berm ridge (backshore berm) surface; pond, spit-growth-enclosed beach pond; (d) pond-isolated berm ridge (intertidal-supratidal sand spit interval), without embryonic dunes. *HT* high-tide level, *LT* low-tide level (Otvos 2000) (Reprinted with permission from Elsevier Science)



Beach Ridges, Fig. 2 Prograded strandplain, southeastern Horn Island, MS. (a) Inter-swale ponds between old beach ridge sets in wooded interior. Elongated shore-parallel ponds of different orientation isolated from Gulf of Mexico (south) by barren narrow strip of backshore and berm ridges (USDA aerial photo, January, 1958. Width of image: 4.56 km). (b) Eleven years later: western half of previous image.

Shore-parallel westward spit and spit-platform growth about to form new cat's eye pond (bottom). Remnant of isolated, mostly sand-filled old "cat's eye" beach foreshore pond visible to the east. Eroding (white) and forested (dark) old interior beach ridge sets, with swale ponds in the island interior (USCGS aerial photo, October, 1969. Width of image: 2.3 km)

origin (e.g., Price 1982). Hesp (1984, 1985), Mason (1990), and Mason et al. (1997) thus distinguished between low profile "smooth, terrestrial," squat, small berm ridges and steep dune ridges that often overlie and bury them..."

The presence of berms, often absent from beaches, especially from dissipative and high-energy beaches (e.g., Short 1984), clearly is not indispensable precondition for foredune development (Hesp 1984, 1985). Berm formation by wave action on the Tabasco shore of the Gulf of Mexico has been attributed to alternating "cut-and-fill" cycles of erosion and

aggradation (Psuty 1966). This berm-shaping process, however, appears to be a localized and ephemeral phenomenon without bearing on long-term strandplain development. The adjacent, 20-km wide, ~6.5 ka old Tabasco-Grijalva strandplain, for example, has been receiving abundant fluvial and eolian sand supply and represents a foredune ridge plain, underlain by backshore and berm ridge deposits (Nooren et al. 2017).

After isolation from the active foreshore by progradation, berms that became inactive may convert into beach ridges.

Influenced by seasonal water level fluctuations, a miniature swash-built coarse to medium sandy strandplain sequence, driven also by occasional minor storms, formed rapidly on the Caspian seashore (Storms and Kroonenberg 2007). Hesp et al. (2005) restricted the term “beach ridge” to wave-built structures of sand, less frequently of coarse clastics, at or somewhat above normal spring tide level. However, unless composed of coarser particles such as granule or gravel, including coral shingles, or protected by foredune cover, low and ephemeral “berm ridges,” composed of fine-to-medium grained sand, usually possess no preservation potential on active beaches, let alone along relict shorelines. On the other hand, one strandplain, just as several others located in tropical Australia, was formed over a time span of six millennia. It illustrates the survivability of a coarse sand ridge that apparently was constructed by unusually powerful cyclonic storm waves and surge along a slowly seaward-shifting shoreline. Deposition during repeated overwash events was the likely cause for ridge aggradation (Nott et al. 2009). The strandplain composed of 3.5–6.0 m high beach ridges is capped by layers of fine eolian sand.

Shore-parallel inter-ridge swales bracket each ridge. Several authors regarded high-tidal sand berms as incipient (wave-built) beach ridges during the Australian “berm debate” (Davies 1957; Bird 1960; Hails 1969). Later, Bird and Jones (1988) proposed that if a berm survived a 15-day tidal cycle, it becomes a beach ridge.

Alternating rapid berm development and elimination processes under storm, respectively, fair-weather accretional conditions, influenced by variable total run-up values, were monitored in great detail in microtidal mixed sand-gravel beaches (Bergillos et al. 2016). Berms were credited with providing a foundation for the development of embryonic foredunes that develop into full-sized ones (Davies 1957; Bird and Jones 1988). However, dune ridges, parallel with and behind foredune ridges, often form from rows of embryonic dunes on backshore sand plains as well. They are stabilized by grassy vegetation.

Strandplains (beach ridge plains) consist of more than two beach and/or relict foredune ridges with the intervening swales. The landforms are oriented approximately parallel with the shoreline. Along the passive continental margin of SE Australia, mid-to-late Holocene strandplains prograded dominantly by processes of forced regression during sea-level fall and by alongshore and onshore shoreface sand supply (Kinsela et al. 2016).

Beach Ridge Categories

Shell Ridges Along Beach Foreshores and Landward of Macrotidal Mudflat

Seashells eroded and winnowed from shallow marine and intertidal mudflat deposits occasionally form low, elongated

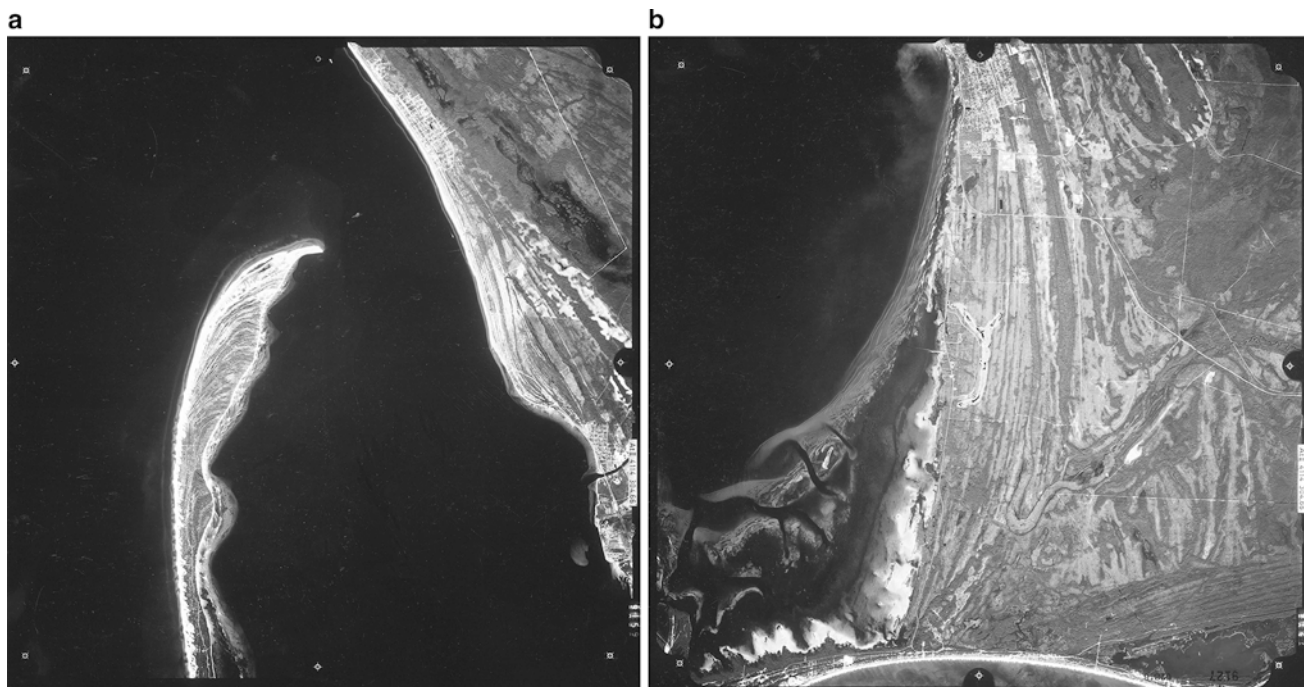
ridges and mounds on berm surfaces with minor sand content along high-tidal shorelines. Such landward-advancing shell concentrates that display a distinct morphology and mislabeled “chenier” (q.v.) by Neal et al. (2002) have overridden and buried the seaward margins of high-tidal salt marshes in southeastern England. Identical in content and form to microtidal Sanibel Island, Florida, beach landforms, these frequently overwashed elongated shell ridges display wave-molded landward-facing crest and berm slopes, respectively, seaward-inclined foreshore faces and internal sedimentary structures.

Truncation Lines That Separate Individual Beach Ridge Sets

Where beach ridge progradation has been abruptly terminated by shore erosion, then followed by renewed beach growth, there are crosscutting truncation lines that separate generations of beach ridges in mainland and island barriers of Quaternary age. This process is well documented in all Mississippi barrier island strandplains (Otvos 1981). The St. Joseph Bay area barrier spit and small mainland strandplains in NW Florida provide good illustrations (Fig. 3). (see article ► “Gulf Shorelines, Last Eustatic Cycle”)

Gravel-boulder (“storm”) ridges. Storm-associated high tides and waves build gravel ramparts as high as 6 m (Clapperton 1990). Gravel-boulder ridges, associated with storm surges, therefore rise well over their associated sea (lake) levels. Coarse clastic sediments, including large seashells, resist backwash erosion and become stranded on these shore ridges. Permanent shingle emplacement at superelevated tide levels is aided by backwash percolation into the permeable gravelly substrate (Carter 1988). For a given still-water level, the height of wave-built ridges formed during winter storms may vary by as much as 2–2.6 m (Adams and Wesnousky 1998). These “storm ridges” are common on glaciated and bedrock shores and also on tectonically or isostatically raised marine and lacustrine terraces. Examples abound in Canada’s Maritime Provinces, New England, and on high Pacific shores between Alaska and Mexico.

Coarse clastic beach ridges or bedrock terrace veneers accompany raised strandlines of pluvial and glacial lakes in the North American interior basins (Fulton 1989; Morrison 1991). Coarse clastic sediments were delivered by high-gradient streams, alluvial fans, mass wasting, and, occasionally even, fluvio-glacial processes. Adjacent bedrock areas that served as sediment sources have undergone intensive physical weathering under periglacial and cold-temperate conditions. Pluvial Lake Bonneville and its successor, early stage Great Salt Lake in Utah and Nevada, as well as Lake Lahontan in Nevada and California, provide the best examples (Morrison 1965, 1991; Currey 1980; Adams and Wesnousky 1998, 1999). Gravel-boulder ridges, deposited



Beach Ridges, Fig. 3 (a) Generations of narrow Late Holocene strandplain fans, northern St. Joseph Bay, Apalachicola Coast, NW Florida. *Left*: north tip of St. Joseph Peninsula (barrier spit). *Right*: small Holocene Palm Point mainland strandplain. *Upper right corner*: wide Late Pleistocene (Sangamonian) marine highstand beach ridges and partially filled swales represent erosion-impacted strandplain sectors

(q.v., Last Interglacial entry). (b) St. Joseph Bay, NW Florida. Quaternary strandplains. Wide beach ridges of N–W-trending Late Pleistocene (Sangamonian) strandplain (*center field of photo*) again contrast with narrow, crisply outlined Late Holocene strandplain ridges (*lower left*) (USGS aerial photo, October, 1978. Width of image: ca. 15 km) (Otvos 2005, 2015)

on wave-cut bedrock terraces, form discontinuous tabular and tabular cross stratified bodies, several meters thick (Adams and Wesnousky 1998). Playa beach ridges that contain carbonate nodules include paleosols and incorporate secondarily calcreted and gypsum-creted grit in the arid Lake Eyre basin, Australia, one of the world's largest internally drained regions (Nanson et al. 1998).

Bouldery-gravelly clastic deposits that often dominated ice-dammed glacial lake shorelines along the fluctuating glacial margin in North America originated from reworked moraine till, fluvio-glacial delta, periglacial colluvium, and, in particular, ice-contact (esker, kame, outwash delta) deposits. Major examples include relict shore features on glacial lakes Agassiz, Algonquin, and Ojibway (Fulton 1989, pp. 144–145, 257, 343, 362–364). Due to sand scarcity, the short life span of a given strandline, and erosive wave regimes, instead of regular beach ridges, gravelly bouldery shore zone lags frequently veneer wave-cut lake terrace surfaces.

Sediment types and forms in sandy beach ridges. Depending on the wide range of wave and current conditions and source sediments on a given marine or lake shore sector, the upper beach deposits (intertidal-supratidal intervals) may be represented by fine-to-coarse, even gravelly sands in the berm ridges. Whereas sorting tends to be good in uniformly sandy beaches, it becomes moderately sorted when coarser

clastic fractions are also included (e.g., Thompson 1992). Chappell and Grindrod (1984) described a rare transition between sandy ridges of a regular beach ridge plain and a small adjacent chenier plain (q.v.), composed of shell-rich ridges. Reflecting the low relief and gentle seaward and landward slopes of the berm ridges, basinward-dipping parallel laminae and low-angle (3–5°) cross-lamination tend to characterize the upper foreshore slope. Subhorizontal or gently landward-dipping (overwash) laminations occur on the landward beach ridge surfaces. The highly variable beach ridge dimensions, the height above still-water level, and slope angles depend on wave conditions, local tidal, or lake level ranges, including wind-induced rise in sea and lake levels along given shore sectors.

Pebble-armored ridges. Pebble sheets plastered onto sand dunes during storms were designated as gravelly ramp barriers (Orford and Carter 1982; Mason 1990). The hydraulic ratios and shapes of shell bioclasts result in higher transport and dispersal rates than with regard to larger, denser silicate rock clasts. Whereas gravel/boulder ridges accumulate during direct storms, their impact tends to flatten and disperse already existing ridges, composed of sand and lighter, platy shell clasts (Greensmith and Tucker 1969; Rhodes 1982, p. 217). Higher-energy events enable accumulation even of bioclastic rudites (Woodroffe et al. 1983; Meldahl 1993).

Strandplain Versus Shore Terrace Development

Strandplain formation may be a continuous process with grain-by-grain addition of sand to the widening foreshore. Continuous progradation of the neap berm at mid-to-high-tide level results in a gently undulating, almost level beach plain. On mesotidal foreshores where neap high tide remains below the highest foreshore levels, continuously accreting neap berms are uninterrupted by inter-berm swales (Hine 1979; Fig. 17a). Increased sand supply along the low-microtidal Gulf of Mexico beaches leads to steady outbuilding of the foreshore, accompanied by consequent progradation of narrow, closely spaced beach ridges. Discontinuous beach ridge progradation involved either the stranding or remolding of landward-migrated mesotidal swash bars on the foreshore (Hine 1979; Carter 1986; Wilson 1996) or spit growth from and downdrift reattachment to the beach in microtidal settings (Otvos 1981).

Both processes isolate elongated ponds. Fronted seaward (lakeward) by newly formed active foredunes, and slowly filled by eolian and washover sands, such ponds may gradually become wide supratidal inter-ridge swales (Figs. 1 and 2).

A comparison of Late Pleistocene gulf coastal strandplain ridges with the sharply outlined, narrow Late Holocene beach ridges illustrates the fact that prolonged infilling and erosional modification of Late Pleistocene Sangamonian (Last Interglacial (q.v. Last Interglacial entry) strandplain ridges (Otvos 2005, 2015) result in more subdued, more gently sloping beach ridges, separated by wider swales (Fig. 3). Instead of strandplains, gently undulating, nearly level eolian sand terraces form when sand supply and beach progradation do not keep pace with rapidly growing and sand-trapping beach vegetation. Beach progradation and/or eolian sand supply rates under these conditions are relatively low (e.g., Ruz and Allard 1994).

Rates of Beach Ridge Development. Beach Ridges as Event Markers

Depending on ridge dimensions, antecedent topography, vegetative conditions, sediment volume and supply rates, beach ridge development may proceed slowly or quickly. Thus, Nanson et al. (1998) report on a beach ridge that formed in less than 1 year along an Australian playa lake during a high lake stage. Development rates in a number of other calculated examples from worldwide locations ranged between 1.8 and 3.3 year/ridge, at other sites the rates were as low as 30–150 year/ridge (in Otvos 2000, pp. 90–91). Calculation yields an even lower, 295 year/ridge rate in the earlier cited Cowley Beach strandplain in which ridge accumulation probably marks rarely recurring high-energy storm-related

aggradation events (Nott et al. 2009). Among the largest and oldest Holocene beach ridge plains, the ~300 km long Grijalva-Usumacinta strandplain complex on the south shore of the Gulf of Mexico began ~6.5 ka ago. The plain includes ~500 ridges; growth rates ranged between 6–19 yr/ridge (Psuty 1966; Nooren et al. 2017). Archiving alternating wet and dry Holocene climate phases, Dillenburg et al. (2017) discussed the record-setting (18-km) progradation of the large South Brazilian Cassino Strandplain.

Beach Ridges: Ancient Marine and Lake Shores, Water Levels

Former sea (lake) levels may be identified when a horizontal interface is recognizable between the wave-built foreshore and the overlying eolian lithosome in a given ridge. This was the case in Lake Michigan strandplain ridges (Fraser and Hester 1977; Thompson 1992) where low-angle sand and gravel cross beds and trough-cross-bedded lacustrine sands of wave-built origin underlie land snail-bearing, cross-bedded, in part massive, structureless dune sands. On pure sand beaches that lack granule and pebble clasts due to the very short transport distance from immediately adjacent sediment sources, distinctions between wave- and wind-deposited lithosomes often are difficult or impossible to make on granulometric grounds alone.

At times of superelevated lake and sea levels, associated with storm-related temporary rise of the water levels, wave-built sandy, shelly, or gravelly beach ridges may aggrade significantly above normal high-tide levels (e.g., Mason 1990; Mason and Jordan 1993; Otvos 2018). Precise reconstruction of former sea levels from beach ridges therefore may be problematical. However, based on GPR surveys, berm, and foredune morphology, from a Portuguese barrier spit sequence Costas et al. (2016) did establish a 6.5 ka sea-level curve within a vertical error range of ~0.8 m.

Similar to beach ridge summits, coarse clasts can be useful markers of reference surfaces to document postdepositional tectonic and isostatic changes, including vertical displacement, tilting, and warping of the land surface. In their absence, wave-cut bedrock terraces and lag clast-veneered strandlines may also serve this purpose. Strandline staircases were documented on isostatically uplifted marine and glacial lake shores in subarctic North America and Scandinavia (e.g., Hudson Bay, Tyrell Sea; Fulton 1989). Flights of raised beaches characterize former pluvial/playa lake shores in western North America, Australia, and other presently arid and semiarid regions. Bounding surfaces of lower foreshore *Lepidophthalmus* (formerly *Callianassa*) ghost shrimp-burrowed barrier ridge deposits and correlative adjacent lagoonal-saltmarsh surfaces in six Late Pliocene-Pleistocene

coastal terrace sequences were similarly utilized on the Georgia-northeast Florida seaboard (Hoyt 1969; Otvos 2018) (q.v. Last Interglacial entry).

Shore Ridges Reflect Tectonic Events

Beach ridges may contribute to critical modern and paleoseismic data in tectonically active regions. On the Kenai Peninsula coast, Alaska, coseismic subsidence-induced multiple beach ridge erosion events were related to two subduction zone megathrust earthquakes (Kelsey et al. 2015). From another quake-impacted setting in Indonesia, Monecke et al. (2015, 2017) reported comparable phenomena. These involved a strandplain, fronted by a high shore ridge, destroyed by the monumental Sumatra Tsunami of December, 2004. Later, a new, higher ridge formed further landward.

Summary

Beach ridge is a relict, no more active wave-, often wind-constructed shore ridge, generally parallel with the coastline, usually as part of a ridge sequence, with landward-adjacent similar ridges. The ridges are composed of sand, occasionally gravel, and shell concentrate. Sets of beach ridges form strandplains, also named beach ridge plains. Beach ridges often form along prograding seashores, in prograding barrier islands, and on lake shores when abundant longshore (littoral drift) and/or onshore-directed sediment supply is available. The rate of ridge plain progradation depends on local sediment supply and hydroclimate (wave and erosion) conditions. It varies greatly between different coastal localities. Seashells eroded and winnowed from shallow marine and intertidal mudflat deposits occasionally form low, elongated ridges and mounds on berm surfaces, with minor sand content along high-tidal shorelines. Use of beach ridges as sea-level or storm event markers is often problematic and requires several lines of confirming evidence.

Cross-References

- Barrier
- Barrier Island Landforms
- Beach Processes
- Changing Sea Levels
- Cheniers
- Dune Ridges
- Ingression, Regression, and Transgression
- Last Interglacial
- Meteorological Effects on Coasts
- Pleistocene Epoch

- Rock Coast Processes
- Tectonics and Neotectonics
- Wave Climate

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Beach Safety Research

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Definition

Each year, hundreds of people drown and tens of thousands more are injured or rescued on beaches globally. A recent and growing field of coastal science involves studies specifically related to beach hazards and beachgoer safety.

This multidisciplinary beach safety research has improved our understanding of the physical and social factors behind common beach hazards and has aided water safety practitioners, governments, and tourism agencies in developing education and awareness strategies for beach user safety.

Introduction

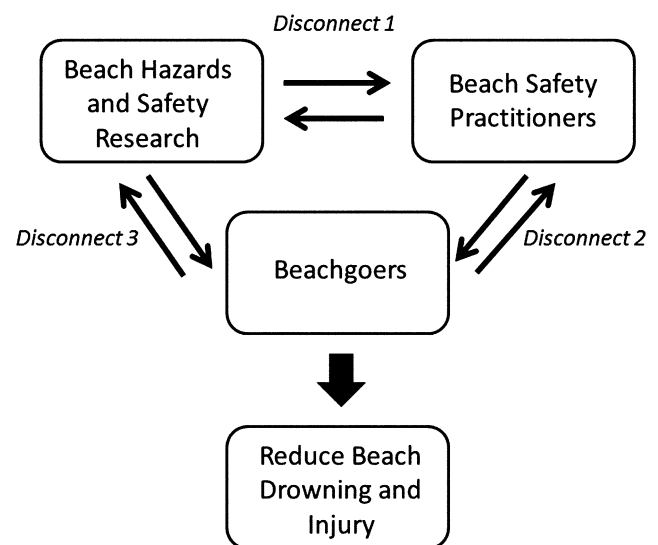
Beaches along both oceanic and lacustrine shorelines are a major recreational destination for millions of people worldwide each year. However, due to a range of hazards and physical and social factors, there is always some level of risk of drowning or injury for those who enter the water for recreational purposes. Due to an overall lack of consistent, accurate, and reliable incident reporting, the total numbers of beach drownings, injuries, and rescues that occur globally each year are unknown, but are clearly substantial. In both the United States and Australia alone, approximately 100 fatal coastal drownings and tens of thousands of rescues occur on average each year according to United States Lifesaving Association (USLA) and Surf Life Saving Australia (SLSA) statistics. Collectively, beach-related drowning deaths, near-miss drownings, injuries, and trauma represent a serious global public health issue with significant personal, societal, and economic costs.

Common physical beach hazards affecting beach user safety include (i) offshore flowing rip currents (see “► [Rip Currents](#)”), which are considered the greatest hazard to swimmers, as they can transport people seaward through the surf zone and various distances beyond, often against their will (e.g., McCarroll et al. 2014); (ii) sudden changes in water depth caused by beach step and bar/trough morphology leading to bathers finding themselves out of their depth (Short and Hogan 1994); (iii) dangerous breaking waves at the shoreline, or across sand bars, leading to surf zone injuries (e.g., Puleo et al. 2016); and (iv) large wave activity and/or steep beach profile gradients creating dangerous swimming conditions and strong swash activity capable of knocking people off their feet and/or transporting them offshore into deeper water.

Social factors which influence beach user safety include (i) the presence or absence of lifeguards (Houser et al. 2017), (ii) beach user swimming ability (e.g., Drozdowski et al. 2015), (iii) frequency of beach visitation and exposure (e.g., Morgan et al. 2008); (iv) familiarity and understanding of physical beach and surf conditions including hazards themselves (e.g., Brannstrom et al. 2014), (iv) awareness and understanding of beach flags and warning signage (e.g. Matthews et al. 2014), and (v) beach usage and behavior (e.g., Sherker et al. 2010). Of note, while a relatively sound scientific understanding of physical beach environments and hazards exists, much less has been documented about the social aspects of beach safety, including demographic and cultural variability.

Based on a model proposed by Brander and MacMahan (2011), the continued occurrence of beach drowning and injury is partially due to several key information disconnects involving the communication of scientific knowledge of beach hazards and beach safety information between scientists, beach safety practitioners, and the general beachgoing public (Fig. 1). The first disconnect involves a lack of direct collaboration between coastal scientific researchers, who have an understanding of beach hazards, and beach safety practitioners, organizations, and local governments responsible for developing and communicating beach safety information to the public. The second relates to a lack of evidence-based beach safety education strategies and programs aimed at a range of demographic beach users as well as a lack of understanding of their effectiveness and how better to make beach users aware of these strategies. The third disconnect relates to the fact that the general public has little direct access to scientific journals, coastal scientists do not often engage directly with the public, and few scientific studies have examined social aspects of beach hazards and safety. Brander and MacMahan (2011) argued that reducing these disconnects is crucial to reducing beach drowning fatalities (Fig. 1).

A recent increase in beach safety research has helped address each of these disconnects. In particular, there have been excellent research-based collaborations between coastal scientists in the United States and the National Oceanographic and Atmospheric Administration (NOAA), National Weather Service (NWS), and the USLA. In the United Kingdom and Australia, scientists have partnered with the Royal National Lifeboat Institute (RNLI) and SLSA, respectively, and similar collaborations are increasingly occurring worldwide. These collaborations have involved a range of research involving



Beach Safety Research, Fig. 1 Improving information disconnects between beach safety research, beach safety practitioners, and beach users may help reduce beach-related drowning and injury

beach hazard rating and forecasting, epidemiological studies, and social science and socio-physical approaches to common beach hazards that have had important implications toward managing future beachgoer safety.

Hazard Rating and Forecasting

Several attempts have been made to construct beach hazard ratings based on the prevailing character of beach type; morphology; quantification of surf zone processes, such as wave height and tide range; and the type and presence of nearshore currents and structures (Short and Brander 2014). The most notable example is the Australian Beach Safety and Management Program (ABSAMP), which provides a hazard rating for each beach in Australia (Short and Hogan 1994) based on a scale of 1 (low hazard) to 10 (high hazard) and has been adapted for use in other countries. However, it does not take into account the presence of lifeguards, rescue statistics, or beach user information, which is often difficult to obtain.

In terms of specific beach hazards, a beach risk assessment program in the United Kingdom involving beach type, wave and tide conditions, beach lifeguard rescue statistics, and field measurements of rip currents (Scott et al. 2014) has shown that high-risk, high-exposure scenarios for bathers being caught in rip currents occur around mean low water on days with low wave heights (<1 m), long wave periods (>10 s), a shore-normal wave approach, and light winds (<5 ms^{-1}). An operational rip current prediction tool was developed that showed 64% of recorded lifeguard incidents occurred under high-risk conditions, and 36% occurred during medium-risk conditions.

In the United States, efforts at rip current risk forecasting have primarily relied on correlative predictions to relate rip current occurrence with meteorological and oceanographic conditions. Rip current outlooks based on a low-/moderate-/high-risk scale are issued by the US National Weather Service as part of daily weather forecasts on television news and weather stations (Moulton et al. 2017). In some cases these outlooks are assisted by observations provided by lifeguard services. However, given the variability of wave climate, geology, and morphology between and within different regions and beaches, any type of beach hazard rating based on empirical predictive indices should be calibrated locally and applied elsewhere with caution.

Epidemiology

The first step at identifying the extent and nature of drowning and injury on beaches is conducting evidence-based research focusing on the epidemiology of the problem. Typically these

studies are often country- or region-focused and rely on available statistics of drowning or injury that occur due to a range of beach hazards or specific hazards. Often these studies provide a basic description of the number of incidents over a given time period, such as Brighton et al. (2013) who established that rip currents are the most serious hazard on Australian beaches being responsible for 21 confirmed drowning fatalities on average each year and 44% of all beach-related drowning deaths. Other studies provide further detail involving demographic characteristics, location, timing, presence/absence of lifeguards, and relationships to basic environmental factors such as weather, wave and tidal conditions, and beach morphology (e.g., Gensini and Ashley 2010; Arozarena et al. 2015). For example, in a study of surf zone injuries (SZI) on the east coast of the United States, Puleo et al. (2016) found that men and tourists were twice and six times as likely to be injured relative to women and local residents, respectively, with highest injury rates associated with moderate wave heights.

Other studies have attempted to use resident population, beach visitation numbers, and/or bathing observations to establish crude surf beach drowning rates per both number and demographic type of population. Using available databases, Morgan et al. (2008) found that between 2001 and 2005, the annual average crude surf beach drowning rate in Australia was 0.28 per 2.36 per 100,000 population for males and international tourists, respectively. However, the primary challenge of conducting epidemiological studies is the dependence on data availability and quality, which varies globally, and for many areas is extremely limited.

Social Science

There has been a recent increase in interest in the social dimension of beach hazards and safety that has been multidisciplinary in nature, involving elements of social, behavioral and public health sciences, human geography, and tourism studies. Much of this research has involved gathering data via survey questionnaires administered either in hard copy form to beach users at both beach and non-beach locations or via online surveys, sometimes distributed through social media channels, to the general public. Often studies focus on specific demographic or “at-risk” groups, such as young males, tourists, and international students (e.g., Bal-lantyne et al. 2005; Williamson et al. 2012; Houser et al. 2016). To date, this type of research has primarily occurred in Australia and the United States with a particular focus on the rip current hazard.

One of the key factors leading to drowning fatalities, injuries, and rescues at beaches is a lack of public awareness and understanding of beach safety practices and hazards. Matthews et al. (2014) found that less than half of surveyed

beach users observed beach warning signs when present and those that did were more likely to notice hazard symbols rather than textual or diagrammatic information. Brannstrom et al. (2015) examined rip current safety signs in the United States finding that the core graphic used was not particularly successful in improving users' ability to identify rip currents. This is problematic given that numerous studies have shown that most beachgoers are unable to identify rip currents, even when shown a picture of one (Sherker et al. 2010; Brannstrom et al. 2014). However, studies in Australia and the United States have shown that improved understanding, awareness, and identification of rip currents and beach safety can be achieved through dedicated public safety campaigns (Hatfield et al. 2012; Houser et al. 2017) and in some cases may help reduce the number of drowning fatalities (Klein et al. 2003).

An important component of social-based beach safety research involves examining beach user behavior, often between different demographic groups such as rural residents or international tourists, in relation to proximity or presence of beach lifeguards and associated beach flag systems. The reasons why people choose to swim away from lifeguard patrol have been related to beach crowds, distance to nearest lifeguard service, proximity to beach access points, and beach user characteristics such as gender, age, beach familiarity and exposure, self-rated swimming ability, and lack of knowledge (Klein et al. 2003; Ballantyne et al. 2005; Sherker et al. 2010; White and Hyde 2010). Results of these studies have shown that improving beach safety requires more refined strategies rather than a "one-size-fits-all" approach (Williamson et al. 2012).

Socio-Physical Approaches

Beach safety research also involves relating physical field measurements of waves, currents, and nearshore morphology to social science dimensions. Scott et al. (2014) examined the occurrence of individual and mass rescues conducted by lifeguards in the United Kingdom on macrotidal beaches with bar/rip channel surf zone morphology and found them to be related to attractive weather and wave conditions and large numbers of bathers, but also to characteristics of rip current flow. Most rescues were linked to water levels 1–2 h either side of low tide when enhanced wave breaking caused rip current flow to suddenly "turn on and off," particularly during spring tides when rip flow activation was maximized. These findings have helped RNLI lifeguards in operationalizing their beach surveillance strategies on these beaches.

Australian-based field studies using GPS attached to Lagrangian drifters and swimmers examined rip current circulation patterns and the efficacy of different rip escape strategies, specifically staying afloat and swimming parallel

to the beach, in different types of rip currents (McCarroll et al. 2014). Staying afloat was found to be a longer duration and more variable escape strategy than swimming parallel, but both strategies failed in some scenarios. Subsequent field and modelling studies (Castelle et al. 2016) also clearly showed that no single-escape strategy action is appropriate in all rip current flow scenarios. These studies were done in conjunction with online surveys and interviews of rip current survivors to learn about firsthand experiences of being caught in rip currents (Drozdowski et al. 2015). Swimming parallel was the most recalled rip current escape advice, but regardless of swimming ability and understanding of rip currents, panic tended to override any recalled advice, and most respondents attempted to swim against the rip current back to shore. These findings were directly incorporated into SLISA public education material which focusses on "knowing your options" when caught in a rip current.

Summary

Beach safety research involves a multidisciplinary approach to understanding the socio-physical factors related to both the incidence and prevention of drowning fatalities and injury on beaches. It has had significant practical applications by influencing public education and awareness interventions implemented by beach safety practitioners. While most of this research has been limited to specific geographic regions and primarily focused on the rip current hazard, there is significant potential for application to other beach hazards and beach user activities globally. More research is needed to better understand demographic, cultural, behavioral, logistical, and knowledge-based factors involving decisions to bathe or swim and to evaluate the effectiveness of both existing and future public beach safety strategies.

Cross-References

- [Beach Use and Behaviors](#)
- [Lifesaving and Beach Safety](#)
- [Rating Beaches](#)
- [Rip Currents](#)
- [Sandy Coasts](#)
- [Surf Zone Processes](#)

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Beach Sediment Characteristics

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Beach sediments are derived from a wide variety of sources, including cliff erosion, rivers, glaciers, volcanoes, coral reefs, sea shells, the Holocene rise in sea level, and the cannibalization of ancient coastal deposits. The nature of the source and the type and intensity of the erosional, transportational, and depositional processes in a coastal region determine the type of material that makes up a beach. In turn, the characteristics of the sediments strongly influence beach morphology and the processes that operate on it (Trenhaile 1997).

Grain Size

The grain size of pebbles and other large clastic material can be measured with callipers, and sieves are used for sand and other coarse beach sediments. A number of techniques are used to determine the size of finer sediments including Coulter Counters, pipettes, hydrometers, optical settling instruments, and electron microscopes. The grain size can be expressed using the Wentworth scale, which is based on classes that are separated by factors of two, so that each is twice the size of the one below. A $\log_2$ transform can be used to provide integers for each of the Wentworth class limits:

$$D_\phi = -\log_2(D_{\text{mm}})$$

where D_ϕ is the grain diameter in phi units (ϕ) and D_{mm} is the corresponding diameter in millimeters (Table 1). Unfortunately, the term “grain diameter” can refer to several different things (Sleath 1984):

1. the mesh size of the sieve through which the grains are just able to pass;
2. the diameter of a sphere of the same volume;
3. the length of the long, short, or intermediate axes of the grain, or some combination of these lengths; or
4. the diameter of a smooth sphere of the same density and settling velocity as the grains.

Beach Sediment Characteristics, Table 1 Sediment grain size classification

Type	ϕ units	Wentworth (mm)
Boulder	>-8	>256
Cobble	-8 to -6	256–64
Pebble	-6 to -2	64–4
Granule	-2 to -1	4–2
Sand		
Very coarse sand	-1 to 0	2–1
Coarse sand	0 – 1	1–0.5
Medium sand	1 – 2	0.5–0.25
Fine sand	2 – 3	0.25–0.125
Very fine sand	3 – 4	0.125–0.0625
Silt		
Coarse silt	4 – 5	0.0625–0.0312
Medium silt	5 – 6	0.0312–0.0156
Fine silt	6 – 7	0.0156–0.0078
Very fine silt	7 – 8	0.0078–0.0039
Clay		
Coarse clay	8 – 9	0.0039–0.00195
Medium clay	9 – 10	0.00195–0.00098

The weight-percentages of the sediment can be plotted against the diameter in phi units in the form of histograms or frequency curves. Grain-size distributions are most frequently represented, however, by plotting the grain size data on a probability, cumulative percentage ordinate, and the phi scale on an arithmetic abscissa. The percentiles on the cumulative size distribution can be used to estimate the mean, standard deviation, and other simple descriptive statistical measures, although the calculations can also be made by computer. For comparative purposes, sediment samples can be represented by the mean or median grain size, or by the size of the grain that is coarser than some percentage of the sample.

There have been many attempts to identify the transportation processes and the depositional origin of sediments based on their sediment-size distributions. The grain-size distributions of beach sediments often consist of three straight-line segments, rather than the single straight line of a normal distribution plotted on a Gaussian probability axis. The three segments have been variously interpreted as representing: coarse bed load, fine suspended load, and intermediate-sized grains that move in intermittent suspension; the effect of packing controls on a grain matrix, the larger grains being a lag deposit, with the finest grains resting in the spaces between grains of median size; and different laminae in the beach, representing several depositional episodes. A further possible explanation is that the segmentation of grain-size distributions on log-normal cumulative probability paper may reflect the use of an inappropriate probability model. The log-normal model poorly represents the extremes of natural grain-size distributions, which may

conform much better to a hyperbolic probability function (Trenhaile 1997). Some workers believe that the four parameters of a logarithmic hyperbolic distribution are more sensitive to sedimentary environments and dynamics than the statistical moments of the normal probability function, but others have found that there is little difference (Sutherland and Lee 1994). Grain sizes may also be fitted to a skew log-Laplace model, a limiting form of the log-hyperbolic distribution which is essentially described by two straight lines, and is defined by three parameters (Fieller et al. 1984).

Grain Shape

The shape of beach grains can be expressed in various ways. The roundness of a grain, which refers to the smoothness of its surface, has been defined as the ratio of the radius of curvature at its corners to the radius of curvature of the largest inscribed circle. Grain sphericity describes the degree to which its shape approaches that of a sphere with three equal orthogonal axes. The shape of a grain can range from spherical, to plate, to rod-like forms, according to the relationship between the three axes, which can be depicted in the form of a ternary diagram. Grain shape can be measured and defined using a variety of indices. They include the E shape factor (ESF):

$$ESF = D_s \left[\frac{D_s^2 + D_i^2 + D_l^2}{3} \right]^{-0.5}$$

and the Corey shape factor (CSF):

$$CSF = \frac{D_s}{(D_i D_l)^{0.5}}$$

where D_1 , D_s , and D_i are the long, short, and intermediate axes of the grain, respectively.

The shape of coarse clasts can be determined fairly easily by direct measurement, but this is usually impossible or too time-consuming for sand and other small grains. Therefore, the roundness and sphericity of sand grains has often been estimated by visual comparison with a set of standard grain images of known roundness, although Fourier analysis is increasingly being used (Powers 1953; Thomas et al. 1995). Winkelmolen's (1971) "rollability" index, the time taken for a grain to roll down the inside of a revolving, slightly inclined cylinder, is easier to measure than other shape parameters, and the shape distribution factors, obtained by plotting grain rollability against grain size, may be more characteristic and indicative of the mode of origin of coastal sediments.

Grain Density

The density of a grain is determined by its mineralogy (Table 2). In temperate regions, most beach sediment originated from the granitic rocks of continents, and they largely consist of quartz and, to a much lesser extent, feldspar grains, but carbonates may dominate in the tropics, especially where there are coral reefs. The sediments in pocket beaches enclosed between prominent headlands, and in beaches derived from other restricted source areas, however, can be strongly influenced by the mineralogy of the local geological outcrops, or by the accumulation of shelly carbonate material. Beaches can consist almost entirely of heavy minerals in volcanic areas, and the usually small amounts of heavy minerals in continental beach sediments, such as magnetite, hornblende, and garnet, help to identify the source rocks, their relative importance, and the direction of longshore transport.

Bulk Density and Packing

Bulk density reflects the way the grains are arranged or packed together. Spherical grains of uniform size can be packed in four ways. The centers of grains in unstable cubic packing describe the corners of a cube, whereas a tetragonal arrangement is formed by moving the upper layer of grains so that they occupy the hollows between the grains below. With orthorhombic packing, the centers of the lower layer of grains form a diamond pattern, with the centers of the grains in the upper layer directly above. A rhombohedral arrangement is created by moving the upper layer of grains into the hollows created by the lower layer. The porosity of the sediments is 48%, 30%, 40%, and 26% with cubic, tetragonal, orthorhombic, and rhombohedral packing of spherical grains, respectively.

Beach Sediment Characteristics, Table 2 The mean density of some minerals found in beach sands

Mineral	Density (kg m ⁻³)
Aragonite	2,930
Augite	3,400
Calcite	2,710
Foraminifera shells	1,500
Garnet	3,950
Hornblende	3,200
Magnetite	5,200
Microcline	2,560
Muscovite	2,850
Orthoclase	2,550
Plagioclase	2,690
Quartz	2,650
Rutile	4,400
Zircon	4,600

The shape of the grains exerts an important influence on the bulk properties of a sediment, including its packing geometry, stability, porosity, and permeability. Small cavities are created in a deposit by shell fragments and other flat, flaky, or plate-like particles, which greatly increase its porosity. Differences in the size of the grains also affect packing density and porosity. Smaller grains occupy the spaces between larger grains, increasing the packing density and decreasing the porosity. Grains that are less than about one-seventh the size of the larger grains can pass down through the voids between the larger grains. Packing is also influenced by deposition rates. Cubic arrangements develop when there are high depositional rates and grain collisions, and rhombohedral packing, when slow deposition allows grains to settle into their optimum positions. Grains settling onto the bed with high fall velocities jostle and vibrate the underlying layers, increasing the packing density and reducing the porosity. Suspended grains settling out in still water are also less densely packed than those that are deposited by waves and currents.

Grain Sorting

Grains are sorted or separated according to their shape, size, and density (Table 3). Beach sediments are generally better sorted than river sediments, but less well than dunes. Beach grain-size distributions are occasionally positively skewed, but the skew is generally negative. Although the presence of a tail of coarse grains has been attributed to the removal of fine

Beach Sediment Characteristics, Table 3 Factors controlling sediment sorting. (Steidtmann 1982; with permission of Blackwell Science)

Rate of sediment accumulation
Slow – allows reworking of grains
Rapid – allows little or no reworking of grains
None – scour
Nature of the sediment surface
Size distribution of grains
Packing and arrangement of grains
Type of bedforms present
Style of grain motion
Traction, including sliding and rolling
Saltation
Suspension
Fluid characteristics
Velocity or shear velocity
Turbulence
Depth
Grain characteristics
Size
Shape
Density

grains, or the addition of coarse clasts or shells, skewness can also arise from a single sedimentary event, and it is not necessarily symptomatic of the mixing of two or more sediment populations (McLaren 1981).

Cross-shore and longshore changes in beach sediment characteristics can result from mechanical and chemical breakdown, differential transport of grains according to their size, longshore variations in wave energy, the addition or loss of sediment, or the mixing of two or more distinct sediment populations. Sorting occurs through selection, breaking, and mixing (Carter 1988). Rejection and acceptance phenomena play an important role in the selection process, and in perpetuating sorted grain distributions on beaches. Rejection accelerates the transport of coarse grains over finer grains, whereas shielding impedes the movement of fine grains over coarser grains. Grains moving over material of similar size have a high probability of being assimilated or accepted by the underlying material.

Erosion of a source material produces a lag deposit that is coarser, better sorted, and more positively skewed than the original sediment. If all the transported sediment is deposited, the deposit will be finer, better sorted, and more negatively skewed than the source. If the transported sediment is only selectively deposited, the deposit will be better sorted and more positively skewed than the source. The deposit will be finer than the source if only material finer than the mean size of the source is eroded, but it may be coarser if sediment larger than the mean size is removed from the original deposit (McLaren 1981).

The mean grain size of beach sediments depends on the characteristics of the source and the nature of the sedimentary processes. Mean grain size varies according to differences in wave energy along beaches and on the exposed and sheltered sides of islands, and it also changes through time as gently sloping, storm-eroded beaches recover to their steeper, fully accreted states. In the cross-shore direction, the coarsest sediments are generally found on a beach at the plunge point of the breaking waves, and the grains tend to become finer seawards and landwards of this point. There are often coarser sediments on the upper part of the beach, however, which could either have been stranded over the berm crest by large swash events, or it could be a deflation lag deposit, resulting from the aolian removal of finer grains to form dunes. It is not known whether larger or smaller grains move most easily alongshore, and therefore whether examples of beach sediments becoming coarser downdrift represent anomalous or normal situations. In any case, whereas there is often longshore grading on beaches in essentially closed embayments, it is generally lacking or poorly developed where there are large amounts of sediment moving alongshore, or where active sediment throughput does not allow enough time for it to develop (Carter 1988). The degree of grain-size sorting normal to the beach is also a contentious issue. Some workers

have found that the poorest sorting occurs in the breaker and surf zones and the best in the swash zone, whereas others have found that the degree of sorting declines on either side of the breaker zone.

Beach sediments are also sorted according to grain density, and particularly to the abundance and mineralogy of the heavy mineral component. Small heavy mineral grains occupy the spaces between the larger and less dense quartz and feldspar grains, shielding them from the flow so that they are less easily entrained. The lighter quartz grains are transported alongshore more rapidly than the heavy minerals, even when both types of grains have the same settling velocity – presumably because the smaller size of the heavy mineral grains inhibits entrainment during each brief suspension episode. Selective longshore transport of quartz grains may therefore result in heavy minerals becoming concentrated in erosive lag deposits.

There are often concentrations of heavy minerals on beaches in the form of bands or streaks near the high tide or upper swash zones, in the troughs of ripples, or where there are shells, coarse clasts, or other flow obstructions. The upper swash zone may consist of dark layers of fine, heavy mineral grains grading upwards into light colored layers of coarser, quartz–feldspar grains. The alternating layers are between about 1 and 25 mm in thickness, and they typically extend along the beach for a few tens of meters (Clifton 1969). The formation of swash laminae has been attributed to shear sorting in the downrush, which causes the coarser grains to migrate upwards into the zone of lower shear, while the finer and heavier grains move downwards, into the zone of maximum shear at the bed. An alternate explanation is that the smaller particles tend to fall into the spaces between the larger grains, thereby displacing coarser grains toward the surface.

Heavy mineral concentrations in the cross-shore direction have either been attributed to wave asymmetry, the heavy minerals being carried onshore by high current velocities, but not by the weaker offshore flows, or to beach erosion and offshore transport of the quartz–feldspar grains. There may be poor separation under vigorous wave conditions, however, and the heavy and light minerals can be entrained and transported together.

There has been little research on the effect of sand grain shape on longshore and cross-shore sorting patterns. The proportion of angular grains increases in the direction of longshore transport between Delaware and Chesapeake Bays, possible because their lower settling velocities allow them to remain in suspension longer, so that they are carried further and at higher rates than more rounded grains. On Long Island, however, grains become rounder with longshore transport. In a laboratory and field study, grains of similar size and mineralogy (quartz) were differentially transported and sorted within the swash zone, with the more rounded grains being deposited near the top of the uprush (Trenhaile et al. 1996).

Cross-References

- [Cross-Shore Sediment Transport](#)
- [Cross-Shore Variation of Grain Size on Beaches](#)
- [Longshore Sediment Transport](#)
- [Sediment Suspension by Waves](#)
- [Surf Modeling](#)
- [Surf Zone Processes](#)

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Beach Stratigraphy

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Definition

A beach is the boundary between the land and water bodies such as oceans and lakes that develops on wave-dominated coasts. It is defined as a shore consisting mainly of unconsolidated sand and gravel extending from the low-water line to where marked changes in physiographic form and/or sand and

gravel are observed or to the permanent vegetation line. The zone between the low-water and high-water levels, which has a concave topography and slopes gradually seaward, is known as the foreshore. The area landward from the crest of the berm of a beach is called the backshore.

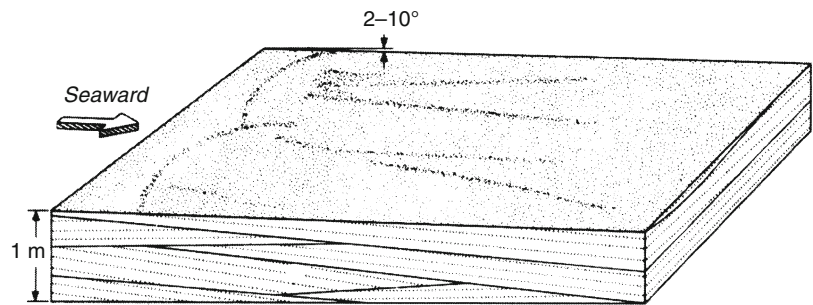
The slope gradient of the beach face varies according to material, particularly the grain size, and wave intensity (Carter 1988; Hardisty 1990). In general, beaches consisting of coarse-grained materials and with steeper waves have steeper slopes. Waves and currents transport and deposit sediments of beaches, resulting in characteristic stratigraphy (Harms et al. 1975; McCubbin 1981).

Succession of Coastal Sediments

At accumulating or progradational beaches, the coastal sediment is typically characterized by a shallowing-upward succession of lower shoreface, upper shoreface, foreshore, backshore, and dunes facies, in ascending order (Clifton 2006). The shoreface, located in the nearshore zone, has a concave topography formed by waves. The upper shoreface is a zone with bar and trough topography constantly influenced by waves and wave-induced currents. The migration of bars and dunes in trough results in the tabular and trough cross-stratification that characterize the upper shoreface sediments. Two-dimensional (2-D) and three-dimensional (3-D) wave ripple structures are also commonly found. The upper shoreface sediments overlie the lower shoreface sediments, which are characterized by swaley cross-stratification (SCS) or hummocky cross-stratification (HCS). HCS is characterized by low-angle ($<15^\circ$) erosional lower set boundaries with subparallel and undulatory laminae that systematically thicken laterally and by scattered laminae-dip directions (Harms et al. 1975). SCS is amalgamated HCS with abundant swaley erosional features. These sedimentary structures are thought to be formed by the oscillatory currents of storm waves with offshore-directed currents. During storms, beaches are eroded and longshore bars migrate seaward (Komar 1998). Strong oscillatory currents caused by storm waves agitate sea-bottom sediments at the shoreface. Some of the sediments are transported offshore by bottom currents caused by coastal set-up and gravity currents. Oscillatory currents related to calming storm waves produce HCS/SCS in the shoreface to inner shelf region overlain by wave ripple lamination. HCS/SCS is found only in sediments of coarse silt to fine sand. Because lower shoreface sediments are mainly deposited during storms, there is a sharp boundary formed by bar migration between upper and lower shoreface sediments (Tamura et al. 2007). The lower shoreface topography depends on inner-shelf topography. Because typical shoreface topography can form only on a gentle/flat basal surface, no clear shoreface topography can form in the steep shelf regions

Beach Stratigraphy, Fig. 1

Swash cross-stratification. Stratification and set boundaries are formed parallel to changing slope of beachface and dip generally seaward (after McCubbin 1981)



of active plate margins. Thus, sometimes only the upper shoreface is referred to as the shoreface.

The uppermost part of the upper shoreface facies represents a step zone characterized by slightly coarser sediments, which are overlain by foreshore sediments. The foreshore sediments are characterized by seaward-dipping ($2-10^\circ$), parallel lamination, and wedge-shaped set boundaries. This structure is called swash cross-stratification or wedge-shaped cross-stratification (Fig. 1). The angle of the dipping lamination reflects the slope of beach face as critically observed in ground-penetrating radar profiles (New Figure; Tamura 2012). The essential characteristics of this stratification are 1–30 cm-thick bedsets, low-angle dips of laminae and set contacts, an average dip direction toward the sea or lake, mostly erosional set contacts, and laminae lying parallel to set contacts.

Backshore sediments overlie foreshore sediments with a gradual contact and are characterized by low-angle landward-dipping parallel lamination, current ripples, plant remains such as rootlets, and heavy mineral concentrations. Light minerals are removed by winds and form eolian coastal dunes behind the backshore. Heavy mineral concentrations are also a characteristic feature of erosional beaches, where they are residuals of the eroded beach sediments.

The coastal succession and sedimentary facies reflect the intensity of current velocity under fair-weather and storm conditions and seaward-decreasing energy conditions. Under fair-weather conditions, from the foreshore to the upper and lower shoreface, the bedforms (sedimentary structures) found are upper plane beds (parallel lamination), 3-D and 2-D subaqueous dunes (trough and tabular cross-bedding), and 3-D and 2-D ripples (ripple lamination), respectively. On the other hand, under storm conditions, beaches are eroded and the lower shoreface resembles an upper flow regime resulting in the formation of HCS. Ripples are formed in shelf regions.

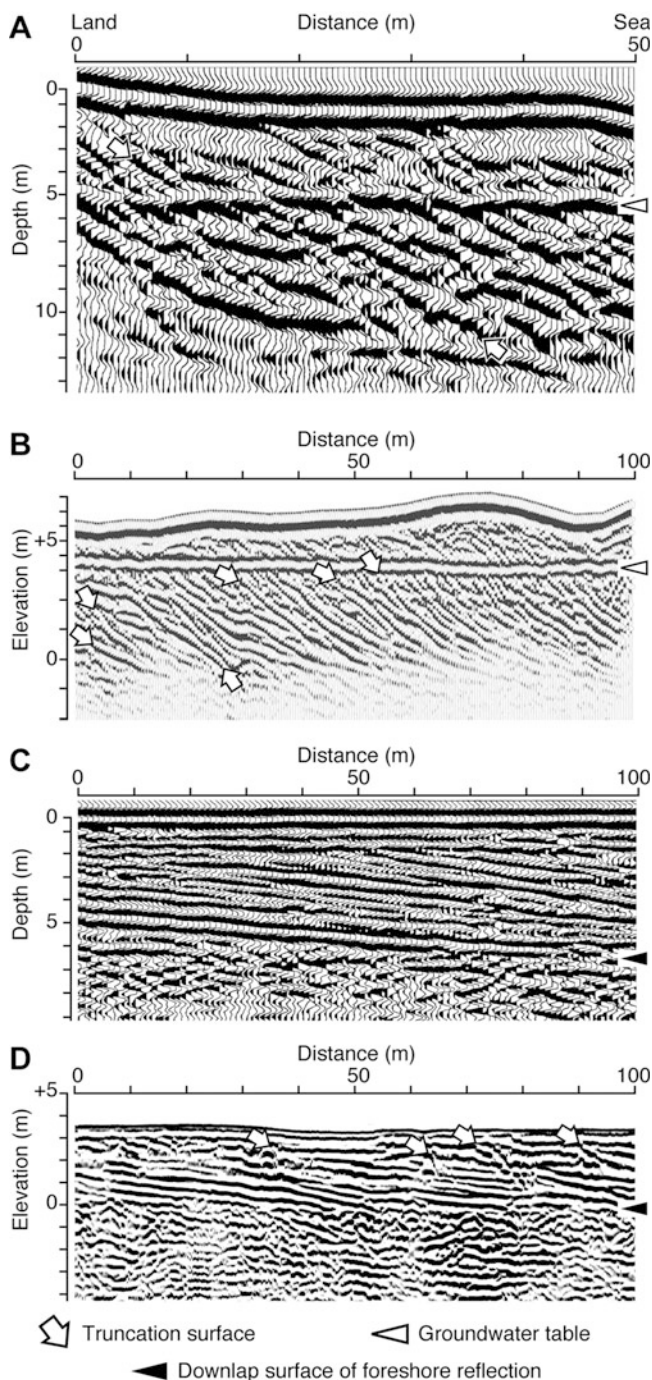
Foreshore Sediments

There are three hierarchies of foreshore sediments: lamination, tide-controlled structures, and storm wave/current-controlled structures.

Foreshore sediments are characterized by parallel lamination formed by the combined processes of wave swash (uprush) and backwash. Each lamina shows reverse grading from fine to coarse with thicknesses of a few millimeters to 2 cm related to each swash and backwash event as a result of either downward filtering of fine particles or Bagnoldian dispersive pressure resulting from shear between the grains in the flow (Clifton 1969; Allen 1984). The fabric of the foreshore sediments shows elongated grains that orient themselves normal to the shoreline, and both landward-imbricated and seaward-imbricated grains are reported. However, these imbricated structures are influenced by the combination of waves and tides.

Reversals of the imbrication dip are thought to result from a predominance of swash transport during the flood stage and backwash transport during the ebb stage. The tidal pattern also influences the depositional thickness of the foreshore sediments. The thick layers are deposited during cycles of higher tidal range, and the thin layers are deposited during cycles of smaller tidal range (Yokokawa and Masuda 1991). Grain size is also influenced by tides. Water-level changes by tides cause the breaker zone of waves and swash/backwash to shift. Allen (1984) showed that coarser sediments are deposited during flood stages, and finer sediments are deposited during ebb stages (Fig. 2).

Storm waves and storm-induced currents have an erosional impact on beaches. During subsequent waning and fair-weather conditions, beaches recover as a result of sediment accretion by waves. This cycle results in an upward-fining succession from a basal erosional surface with coarse-grained materials to finer sandy materials. The coarse deposits formed under high wave energy just after the storm show a remarkable dominance of seaward-dipping imbrication, independent of tidal cycles (Yokokawa and Masuda 1991). By regarding major erosional surfaces in beach sediments as sequence boundaries according to the sequence stratigraphic model, the depositional zone of foreshore sediments and their stacking pattern can be analyzed. A bedset with a thickness of tens of centimeters bounded by major erosional surfaces is regarded as a depositional sequence, and a lamina set with a thickness of several centimeters to ca. 20 cm is regarded as a parasequence. The depositional pattern of lamina sets shows a



Beach Stratigraphy, Fig. 2 Compilation of ground-penetrating radar profiles of beach deposits in different setting (Tamura 2012). (a) Boulder gravel beach (Jol et al. 1996), (b) fine to coarse sand beach (Bristow and Pucillo 2006), (c) fine sand mesotidal beach (Jol et al. 1996), and (d) fine sand microtidal beach (Tamura et al. 2008). All profiles were collected using 100-MHz antennae

landward shift of the depositional zone (onlap) in the lower part of the bedset and a seaward shift of the depositional zone (downlap/progradation) in the upper part. Moreover, bedsets also form a higher order sequence (Masuda et al. 1995).

The storm erosion surface is in some cases represented by concentrations of heavy minerals and causes strong radar wave reflections to be distinguished in ground-penetrating radar profiles (Buynevich et al. 2007).

Seasonal changes in wave direction and wave energy produce seasonal beaches. For example, high, strong waves may create high-level beaches with coarse sediments and a steep beachface, or occasionally erosional beaches with residual coarse sediments and heavy minerals, at the high-water level in winter; calm waves may form depositional beaches in summer, depending on the location of the beach.

Beaches are distributed not only along wave-dominated coasts but also along tide-dominated coasts influenced by waves. In general, tide-dominated coasts have muddy or sandy tidal flats in the intertidal zone. However, waves create narrow beaches in the upper part of the intertidal zone, occasionally with beach ridges landward from the beach. A typical example is the coast of the Mekong River delta, which is a meso-tidal coast with waves. Beaches and well-developed beach ridges are found in the upper part of the intertidal zone to the supratidal zone (Ta et al. 2002; Tamura et al. 2010).

Summary

Beach stratigraphy is characterized by a succession of sediment facies formed in response to wave and current transport and deposition. A typical succession of beach sediments is a shallowing-upward succession that results from sediment accumulation on or seaward progradation of beach. Sedimentary structures of beach facies reflect the sedimentary processes of individual environments under fair-weather and storm conditions; lower shoreface, upper shoreface, and foreshore and backshore facies in ascending order are characterized by swaley cross-stratification or hummocky cross-stratification, tabular or trough cross-stratification, and parallel lamination, respectively. Parallel lamination found in foreshore sediments dips seaward at an angle of the original foreshore slope. Storm waves and currents have erosional impacts on beaches, which are then recorded as erosional surface often with concentrations of heavy minerals in foreshore sediments. Ground-penetrating radar is useful for visualizing structures of beach sediments.

Cross-References

- [Beach Erosion](#)
- [Beach Features](#)
- [Beach Processes](#)

- Ground-Penetrating Radar
- Rhythmic Patterns
- Shelf Processes

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Beach Use and Behaviors

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Introduction

Beaches comprise only 9% of the total conterminous coastline in the United States. Unfortunately, while no national census of beach visits exists, several studies rank beach recreation as one of the most popular outdoor recreational activities in the United States. It is, therefore, surprising that so little research has been undertaken that addresses the socio-economic aspects of these activities. This anomaly is particularly noticeable when contrasting the volume of physical and biological research undertaken dealing with beaches and the nearshore environment. Historically, beach recreational activities have centered on the following three activities: bathing, shore-based fishing, and beachcombing. During the past 20 years, many new activities have emerged, several of which use beaches primarily as a staging area. Such activities include surfing, windsurfing, boogie boarding, and a host of shallowwater boating activities including kayaking, canoeing, personal water crafts (PWC), and surfboarding. Many of these activities are incompatible with the more traditional uses of the beach, resulting in user conflicts. Some of these have been managed through the introduction of local, state, and federal legislation, while others have been adjudicated in the courts. Finally, the absolute increase in the number of users of beaches, as well as the diversity of activities occurring, have resulted in a growing demand for both access and accessibility to the nation's beaches.

This entry begins with a historical overview of beach uses, followed by a discussion of three related concerns: increased beach density; the demands this has placed on beach access and accessibility; and how this problem has been addressed. The entry concludes with a discussion about the increasing threat to beachgoers from pollutants in coastal waters.

Overview of Beach Uses and the Factors Affecting Beach Activities

It is likely that beach recreation owes its origin to the perceived value of beaches as healthy environments capable of relieving serious medical conditions (Goodhead and Johnson 1996). In Britain, during the early part of the 1800s, many people visited beaches with the belief that immersion in, and the drinking of, seawater was healthy and would result in the relief of a number of physical ailments (Meyer-Arendt 1986).

Half a century later, these activities had evolved into resorts generally located within a day's travel of major European and North American cities. Newport, and to a lesser extent Narragansett, Rhode Island, became well-known resorts in New England and were connected by rail to both Boston and New York. In England, Brighton served the same function. More recently, the Hamptons on Long Island have become important beach destinations point for the wealthy. However, nearly all of the research dealing with the early history of beach recreation is descriptive.

One of the few examples where geographers have sought to move from purely descriptive studies to nomothetic research can be found in the extensive writings of Meyer-Arendt, who built on the early work of British geographers with an interest in beach recreation. These studies centered on the concomitant urbanization of coastal areas. Meyer-Arendt studied a number of beach resorts along the northeast Gulf coast, and, based on these efforts, developed the Coastal Resort Morphology Mode (Meyer-Arendt 1986). This is a spatiotemporal, five-stage recreational land use model. The initial stage is characterized by easy beach access that has attracted limited residential developments, which, in turn, support a small recreational business district (RBD). Toward the end of this first stage, increased day visitation takes place. This second phase is referred to as the "Take Off" stage and is characterized by increased recreational development, extending outward from the RBD. Most of this development is along the coastline on both sides of the RBD. Sometimes a recreational fishing pier is constructed, usually at the foot of the RBD. The third phase is dominated by further development and urban expansion. True central business district (CBD) land uses characterize the area immediately surrounding the RBD. Residential developments continue to expand outward, and most of the early structures located closest to the CBD undergo rapid demolition or conversion to more up-scale recreational developments. Most of the structures still cater to a seasonal clientele, but with a small core of year-round residents. If the resort is located on a barrier beach, developments will have reached the bay-shore. As a result, much of the wetlands located there will have been destroyed by canalization, or filled in, resulting in significant environmental impacts. The fourth stage is consolidating the developments characterized in the previous stage, except that condominium developments now cater to those who no longer can afford to buy (let alone build) single-family homes. The final phase is characterized by complete saturation, where lower income residents and those on fixed incomes are forced to sell out, in part because of high property values and property taxes. Dolan and his co-workers analyzed the rise and decline of religious sea camps during the nineteenth century, only vestiges of which exist today.

Following the end of World War II, beach visitation became one of the most popular outdoor recreational activities that cut across all socioeconomic groups, although

significant social and ethnic discrimination was still in evidence. Furthermore, the popularity of beach visitation continued to increase in concert with a general population migration from inland to coastal areas (Kimmelman et al. 1974). Perhaps a contributing factor to the popularity of beach recreation related to the low cost associated with bathing, where transportation often represented the only cost of engaging in the activity. During this period, the predominant activities were bathing, sunning, and socializing, attractions that are as popular today as they were then. The only difference is that today many beachgoers are experiencing significant competition due to other outdoor recreational activities that utilize the beach as a staging area for other water-based activities. These include shore-based fishing and the launching of a variety of light vessels that can be trailered or car-topped, including kayaks, canoes, surf and sailboards, and PWCs.

For many beachgoers, the beach represents a place on which a host of activities can be undertaken, including bathing, sunbathing, ball playing, and socializing. While some degree of specialization appears to take place on certain beaches, most beach visitors tend to participate in several different activities during a day-on-the-beach.

Most early studies concerned with beach uses sought to describe and classify beach users and beaches based on perceived preferences. Several of these studies emerged from the Chicago School of Geography; under the direction of Gilbert White (1973) and his students, where the resource users' perceptions of the environment were seen as the primary factors influencing behavior. The initial research thrust dealt with perceived flooding risks, but these studies soon expanded to include all kinds of perceived environmental factors influencing behavior, including those affecting beach visitation. Few studies have analyzed the activities and social interactions occurring on the beach (Gerlach 1987). Examples of these include Hecock who concluded that beachgoers were attracted to certain beaches based on their physical characteristics (Hecock 1966). This study suggested that younger beachgoers preferred beaches with a stronger wave environment where bodysurfing could be undertaken. Conversely, families with small children preferred beaches where the wave environment was more gentle and where the beach slope was less steep, allowing children to play safely in the shallows (Jubenville 1976).

One area that most coastal recreational planners and resource managers have addressed concerns the number of beach visitors that a given beach can accommodate. While no overall accepted standard exists on the number a given site can accommodate before the perceived value of a visit begins to decline, some efforts were made to address this issue nationally. The Outdoor Recreational Resources Review Commission suggested that 2000 bathers could be accommodated per one mile of beach (Rockefeller 1962). One problem with this measure is that no distinction is being made on the basis of the width of the beach. Jackson (1972), citing a

California study, suggested that each bather in lakes required a minimum of 50 square feet of water. Other factors play a role in the decision-making process leading to a person, family, or group deciding to visit a given beach. In a study conducted in the New York–New Jersey Metropolitan area, West (1973) found that access and especially accessibility were considered more important factors compared to water quality.

By far, most of the social studies conducted on beaches have dealt with density and crowding (Boots 1979). In this context, several authors identified “crowding” as a factor influencing beach use (Sowman 1987). De Ruyck et al. (1987) identified two types of densities, one of which defined overall density as a number of visitors per unit area. He also defined “patch densities,” a term he referred to as “social carrying capacity” on three beaches in South Africa. These researchers found that density tolerance was influenced by the size of the beach (the smaller the beach, the greater the willingness to accept more people (greater densities). He also found that “crowd attracting beach activities,” such as impromptu ball games and other sports events, resulted in higher crowding tolerance by the visitors.

In an unrelated study, West (1974) also found that beachgoers’ perception of beach density varied depending upon the respondent’s residence. Those beach visitors living in urban areas were willing to tolerate greater beach crowding compared with those living in suburban areas.

Beach Access and Accessibility

Access and accessibility are terms often used interchangeably, however, in this entry access refers to the ability to move from an existing “right-of-way,” such as a road or public parking lot, to a public beach. Accessibility refers to the obstacles that a beach visitor may encounter in traveling from his or her home to the beach. Such obstacles may include a lack of parking facilities, high entry fees, or in an urban context, a lack of public transportation to the beach.

Physical access to the shore is governed by two sets of law, one related to common law, the other by legislation. The common law principles concerning beach use originate from old Roman Law, which held that beach resources (seaweed, fish, and shellfish) were held by the sovereign, who then allowed the citizens to fish and collect seaweed from the shore. This principle was adopted in Britain during the Roman reign and eventually transferred to North America during the Colonial Period where, following independence, the concept of the “sovereign” was replaced with the general public. This meant that the government held the submerged lands seaward of the mean high water line (MHWL) in trust for the general public. In most US states, the legal definition of the public domain is seaward of the MHWL (Anon 1988). The MHWL, in turn, is defined on the basis of the location of the average high-tide shoreline during a full metonic cycle.

A legislative approach to increasing public access was initially implied in the Coastal Zone Management Act (1972), and in the subsequent amendments. The 1986 amendments were identified as an area of special interest. Most coastal states have made some efforts to increase public access to the nation’s beaches, although accomplishments vary widely. One of the aims of California’s and Oregon’s, and to a lesser extent Washington State’s Coastal Management Program has been to increase physical access at certain intervals along their respective coastlines. Along California’s rural coast, the aim is to provide coastal access every three miles. This goal is comparable to those formulated in Oregon and Washington. The objective of providing access to the shore at regular intervals has been more problematic along the Eastern Seaboard, in large part because of much higher population densities, less land in public ownership, and overall higher land prices. Together these factors have made eminent domain acquisition much more difficult and costly. Some states have attempted to increase coastal access using the principle of perfecting public right-of-ways. Rhode Island has undertaken a statewide search to identify existing and abandoned right-of-ways, largely through legal research. This effort has significantly increased public access to the state’s coastal areas.

The absence of physical access to the beach is only one of the many constraints that a potential beach visitor is likely to encounter. Lack of accessibility may at times be a greater hindrance to visiting the beach. Such factors may be deliberate attempts by local cities and towns to limit or outright prohibit out-of-town visitors on local beaches. In other instances, impeded accessibility is unintentional or unavoidable (Heatwole and West 1980).

Limiting or prohibiting beach access to out-of-town citizens on facilities owned and operated by local municipalities may vary from outright prohibition to charging unreasonably high entry or parking fees. Many of these instances have been adjudicated in the courts, which have generally ruled that where higher entrance fees have been levied against nonresidents, such fees may be permitted as long as the increased fees cover the additional costs resulting from accommodating the increased number of nonresidents. The courts have generally assumed that a portion of a resident’s property tax is designated to the operation of recreational facilities, including beaches, and that opening such beaches to nonresidents often means increased expenditures to insure the health and welfare of the visitors. This may mean higher costs to cover the costs of additional guards, beach patrols, cleanups, and other services. The courts have generally felt that such additional expenditures could be recovered by charging the nonresidents a higher fee compared with those levied on residents (*Neptune v Burrough of the City of Avon* 1972).

The popularity of beaches and beach uses has increased significantly during the latter part of the twentieth century, a development that is likely to increase for the foreseeable

future. This increased demand has raised two concerns: use conflict and water quality declines.

Conflict Resolution

As mentioned in the introduction, many additional beach uses now exist. Some of these are incompatible with traditional recreational beach activities. Examples include shore-based fishing, various boating activities, including water skiing, use of personal water crafts and surfing. Most of these conflicts have been dealt with on the local level, while a few have been adjudicated in a court of law. Of the management procedures that have been introduced on the local level, zoning procedures are probably the most common. Zoning, as it was first conceptualized in New York City in 1916 (Haar 1977), was originally intended to control building height. Zoning maps later followed with zoning ordinances specifying restricted or prohibited uses.

Recreational applications of zoning have been attempted both on land and on the water in an attempt to reduce conflicts between and among different recreational pursuits. On the water, zoning has been used by a number of municipalities to segregate swimmers and bathers from boaters—especially powerboaters, surfers, and PWCs. Two versions of zoning have been used: permanent zones and space/time zoning ordinances. In the case of permanent zones, a protected activity (e.g., swimming or bathing) is protected from all other activities by prohibiting those from entering the designated area. A less common practice is sometimes referred to as time zoning. In this instance, the competing uses are assigned different periods when each activity can take place, thereby eliminating any conflicts between competing uses. If a given beach is sought by both swimmers and surfers, the beach may be restricted to one user group while the other use may be permitted during different periods. A coastal municipality may allow surfers access to the beach during the early morning and again in the late afternoon. Swimmers and bathers may have exclusive use of the beach and adjacent nearshore during the period from mid-morning to late afternoon.

The same procedures may be utilized on land in areas where users compete for the same area. Sunbathing and shore-based fishing are both legitimate recreational activities that sometimes may compete for the same stretch of beach area. Shore-based fishing may be restricted to the early morning and late afternoon, while sunbathing may be permitted from mid-morning to late afternoon.

Environmental Impacts on Beach Use

Socioeconomic factors are not the only variables influencing beach recreation quantitatively as well as qualitatively. There

are at least two additional variables that increasingly have played a role in this nation's beach recreational activities. One concerns the increased outbreaks of algal blooms, in particular, those classified as harmful. The other concerns the impact that beach activities may have on endangered species and the restrictions imposed on beach visitors to protect threatened and endangered biological resources.

Algal blooms have occurred along the nation's coasts at least since the Spanish first settled Florida. However, there is growing evidence that these incidents are increasing quantitatively and qualitatively. The number of harmful algal bloom (HAB) incidents have increased significantly in recent years as have the impacts on marine life, swimmers, bathers, and people handling fish and shellfish affected by these incidents. While the cause for these events has not yet been determined, there is growing evidence that land-based pollution is partially responsible (Anon 2000). The effects of HAB events range from the discoloration of large patches of waters to fish kills, die-offs of manatees in Florida, and possibly the deaths of small marine mammals in the United States, Scandinavia, and the Mediterranean. So-called "red tide" incidents in Florida, have significantly affected swimmers and bathers. Toxins released from these HABs can also become airborne, resulting in respiratory irritation, coughing, and sneezing by people who are not even in direct contact with the affected waters (Luttenberg 2001).

The second factor that has influenced swimming and bathing in recent years is the potential conflict between the Endangered Species Act and all types of beach recreational activities. During the 1980s and 1990s, large stretches of barrier beaches on Cape Cod were closed to fishing, overland vehicular traffic, and bathing in an effort to protect the Piping Plover nests and fledglings from being trampled. In 1988, it was estimated that only 20 pairs of piping plovers were nesting within the Cape Cod National Seashore (Lopez 1998). Largely because of the severe restrictions placed on beach traffic (both pedestrian and vehicular), a substantial increase in nesting pairs has been noted throughout the seashore (Lopez 1998). These accomplishments, however, have not been made without impacts on beach recreation in general, shore-based fishing or bathing. Within the Cape Cod National Seashore, less than 10 miles of the Atlantic shore are now open to ORV traffic during the nesting season (from March through July). Similar restrictions have been imposed on bathing and beachcombing in piping plover nesting areas.

Conclusions

Beaching and bathing continue to be two of the most popular outdoor recreational activities both here and abroad, yet with few exceptions, not many studies have addressed the behavior and motivation of the beach-going public. Studies conducted

on or adjacent to beaches generally fall into two groups. The first has sought to analyze the reaction of the beach visiting public to deteriorating water quality. The second group of studies has concentrated on infrastructure changes that have taken place in the areas immediately inland from many popular bathing beaches. While these studies are important both socially and economically, it is suggested that many additional findings would enhance our understanding of the factors motivating the beach visitor and the management of beaches. Answers that are needed include studies dealing with crowding and density tolerances and better understanding of beach preferences by different user groups. Are some beaches attracting certain population groups simply because they are more accessible, or because the amenities found on the beach attract specific user groups interested in participating in activities (e.g., surfing) that may not be readily available on all beaches? The role of physical access is still an issue in many communities, notwithstanding that access along the shore is recognized by most states as a public right, Public beaches constitute less than 10% of all the beaches in the United States. This increasingly scarce resource may be better managed if we had a better understanding of the factors that attract and detract the public to certain beaches.

Cross-References

- [Beach Processes](#)
- [Coastal Boundaries](#)
- [Coastal Management Practices](#)
- [Developed Coasts](#)
- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Lifesaving and Beach Safety](#)
- [Rating Beaches](#)
- [Tourism and Coastal Development](#)
- [Tourism, Criteria for Coastal Sites](#)

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Beachrock

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Formation and Distribution of Beachrock

Beachrock is defined by Scoffin and Stoddart (1987, p. 401) as “the consolidated deposit that results from lithification by calcium carbonate of sediment in the intertidal and spray

Beachrock, Fig. 1 Multiple unit beachrock exposure at barrio Rio Grande de Aguada, Puerto Rico. The sculpted morphology, development of a nearly vertical landward edge, and dark staining of outer surface by cyanobacteria indicate that this beachrock has experienced extended exposure. Landward relief and imbricate morphology of beachrock units define shore-parallel runnels that impound seawater. (Photo: R. Turner)



zones of mainly tropical coasts.” Beachrock units form under a thin cover of sediment and generally overlie unconsolidated sand, although they may rest on any type of foundation. Maximum rates of subsurface beachrock cementation are thought to occur in the area of the beach that experiences the most wetting *and* drying—below the foreshore in the area of water table excursion between the neap low and high tide levels (Amieux et al. 1989; Higgins 1994). Figure 1 shows a beachrock formation displaying typical attributes.

There are a number of theories regarding the process of beach sand cementation. Different mechanisms of cementation appear to be responsible at different localities. The primary mechanisms proposed for the origin of beachrock cements are as follows:

1. Physicochemical precipitation of high-Mg calcite and aragonite from seawater as a result of high temperatures, CaCO_3 supersaturation, and/or evaporation (Ginsburg 1953; Stoddart and Cann 1965);
2. Physicochemical precipitation of low-Mg calcite and aragonite by mixing of meteoric and fresh groundwater with seawater (Schmalz 1971);
3. Physicochemical precipitation of high-Mg calcite and aragonite by degassing of CO_2 from beach sediment pore water (Thorstenson et al. 1972; Hanor 1978); and
4. Precipitation of micritic calcium carbonate as a byproduct of microbiological activity (Taylor and Illing 1969; Krumbein 1979; Strasser et al. 1989; Molenaar and Venmans 1993; Bernier et al. 1997).

Although most beachrock cement morphologies suggest an inorganic origin, physicochemical mechanisms operating

alone do not adequately account for the discontinuous distribution of beachrock formations. As Kaye (1959, p. 73) put it, “the problem hinges more on an adequate explanation for the absence of beachrock from many beaches than on its presence in others.” The discontinuity of beach cementation, along with the complex assemblage of cement types found in adjacent samples of beachrock led Taylor and Illing (1971) to propose that the microenvironment exerts a greater influence on the cementation process than does the macroenvironment.

Several beachrock researchers concur with this assessment and support the theory that *initial* cementation in beach sands is controlled by the distribution and metabolic activity of bacteria because: (1) dark, organic-rich micritic rims have been identified around cemented grains in most petrographic studies of beachrock (Krumbein 1979; Beier 1985); (2) microbially mediated precipitation of carbonates has been repeatedly demonstrated in both marine and terrestrial environments (Buczynski and Chafetz 1993); and (3) bacterial populations are particularly large and productive in the intertidal zone of water table fluctuation where beach lithification occurs. Once biologically mediated cryptocrystalline cements are established as nucleation sites, larger crystals precipitated via physicochemical processes can grow and bridge the sediment grains.

Rates of beachrock formation are undoubtedly variable but are generally believed to be quite rapid, on the scale of months to years (Frankel 1968). For example, Hopley (1986) reported that beachrock formed within six months on Magnetic Island near Townsville, Australia, while Moresby (1835) wrote that Indian Ocean natives made an annual harvest of beachrock for building stone and within a year they had a new lithified crop.

Several Pleistocene and older beachrock formations have been identified. However, the dynamic nature of sandy coastlines and a historically fluctuating sea level necessitate that most occurrences of *intertidal* beachrock are less than 2,000 years old. This is commonly supported by the incorporation of modern man-made artifacts in beachrock formations rather than by ^{14}C dates, as beachrock is poorly suited for radiocarbon dating.

The majority of recent beachrock is formed on beaches in the same regions that favor coral reef formation. This is generally below 25° latitude where there is a well-defined dry season and “the temperature of ground water at a depth of about 76 cm in beaches remains above 21°C for at least 8 months of the year” (Russell 1971, p. 2343). However, beachrock can also form at higher latitudes. For example, beachrock exposures are common throughout the Mediterranean and have been reported along portions of the coasts of Norway, Denmark, Poland, Japan, New Zealand, South Africa, the Black Sea, and the northern Gulf of Mexico. Beachrock formations have also been reported on lakeshores in Pennsylvania, Michigan, Africa, New Zealand, southeast Australia, and the Sinai Peninsula.

Subaerially exposed beachrock units constitute only a small proportion of the cemented sediments in the intertidal zone. For example, Emery and Cox (1956) found beachrock *exposures* on only 24% of the predominantly calcareous beaches of Oahu, Kauai, and Maui, whereas jet-probing conducted by Moberly (1968, p. 32) revealed that “exposed or covered beachrock appears to be present at all calcareous beaches in the state” of Hawaii. In the event of continued sea-level rise and human activities that exacerbate coastal erosion, much more beachrock will be exhumed.

Morphology of Beachrock Formations

Beachrock formations typically consist of multiple units, representing multiple episodes of cementation and exposure. Beachrock that forms below the foreshore has an upper surface slope that tends to mimic that of the seaward dipping ($4\text{--}10^\circ$) internal beach bedding. However, beach sand cementation has also been found to occur below the berm and foredune of a beach (Russell 1971; Hopley and Mackay 1978). Those authors found that the beachrock forming below the backshore had a nearly horizontal upper surface that corresponded to the groundwater table and truncated the original beach bedding.

Most intertidal beachrock formations are detached from subaerial and subtidal cemented sediments. Beachrock is laterally discontinuous as well, usually exposed for only short distances before disappearing under loose sand or

ending entirely. It is likely that the formation and preservation of beachrock on a given section of beach is negatively correlated to alongshore increase in wave energy and frequency of beach erosion.

The reported thickness of beachrock formations ranges from a few centimeters up to 5 m, with approximately 2 m being most common. Variations in degree of cementation within a beachrock unit can be controlled by variability in porosity, permeability, and composition of different sand layers (Molenaar and Venmans 1993). Generally, precipitation of cements is most rapid near the top of a beachrock unit. Accordingly, young beachrock units are better cemented at the top and noticeably less so near the base. This attribute makes them more susceptible to scour at their base upon exposure, commonly resulting in undercutting and slumping. It is this undercutting that fosters the development of nearly vertical landward edges on beachrock units. In areas where a chronic deficit in sand supply or erosive conditions have exhumed the seaward edge of a beachrock formation, it is frequently observed to be steep as well.

Long-term exposure of beachrock will radically change the ecology of a sandy shoreline by providing a hard substrate that can support an increased diversity of animal and plant life. The reader is referred to the papers of McLean (1974), Jones and Goodbody (1984), and Miller and Mason (1994) to learn more about the ecology and biophysical modification of intertidal beachrock exposures.

Beachrock and Coastal Evolution

Although beachrock, as defined, forms in the intertidal zone, it does not always remain there. On prograding coasts, a series of beachrock units may form at depth, leaving older units stranded well behind the active beach. On retreating coasts, outcrops of beachrock may be evident offshore where they may serve as a hard substrate for coralline reef growth. If the strike of the beach changes over time, then the strike of the beachrock units will reflect that change.

Armed with the knowledge that beachrock is formed in the intertidal zone, many geologists have related beachrock outcrops to changes in sea level for particular coasts. Semeniuk and Searle (1987) demonstrated that beachrock formation can keep pace with slow shore recession, resulting in a wide, continuous band of beachrock, but that rapid shore recessions (or periods of high wave energy and foreshore instability) are represented by gaps (unconsolidated sediment) in a sequence of beachrock units. Assuming a nearly constant rate of sea-level rise, these gaps may indicate that beachrock can temporarily stabilize the position of the shore under erosive conditions until sea level has risen enough to cause the shore

to jump back (Cooper 1991). Many other researchers have asserted that beachrock outcrops will protect a beach from erosion, as well as control the plan configuration of a coastline.

Research by Turner (1999) has demonstrated that the influence beachrock has on beach processes will largely depend on the extent and morphology of the exposure, both of which evolve over time. Cumulative exposure and erosion of a beachrock formation over a period of years to decades can foster a gradual increase in the landward and seaward relief of the beachrock units and the development of shore-parallel runnels and shore-perpendicular breaches in the beachrock. The high seaward relief of such a beachrock unit effectively attenuates incident wave energy and retards onshore sediment transport. The high landward relief of the beachrock unit can act as the seaward wall of a runnel that blocks offshore return of backwash and forces impounded seawater and entrained sand to flow laterally on the foreshore to low spots and shore-normal breaches in the beachrock formation. Beachrock breaches and runnels are erosionally enlarged over time, locally increasing onshore inputs of wave energy and longshore sediment transport rates on the foreshore.

On a beach on Puerto Rico's west coast, beach width and volume were found to be least stable where the seaward beachrock unit was breached and most stable away from the breaches behind high relief beachrock. Sections of foreshore most protected by a high relief beachrock ridge exhibited the lowest volumes of subaerial sand storage, unusually narrow beach widths, and the slowest beach erosion recovery rates. In short, a beach with a high relief intertidal beachrock exposure is more likely to be sediment deficient and out of synch with the wave regime. This puts the backshore of a beachrock beach at risk of catastrophic retreat following the development of a breach in the beachrock or in the event of a high energy wave event coupled with a storm surge or spring high tide.

Conclusions

The transformation of sandy beaches to rocky beachrock beaches is increasingly common in the tropics and subtropics. Where beachrock is exposed by erosion, it acts as a natural breakwater or revetment, decelerating further shoreline and backshore retreat. However, it also tends to retard beach buildup and is poorly suited to recreational use, both major issues in the tropics where tourism is often the primary source of income. The potential for beachrock to significantly alter the evolution of a coast justifies additional research on its influence on beach processes. In particular, the characteristics and effects of beachrock on dissipative beaches have received little attention and are likely to be significantly different than those observed on more reflective beaches.

Despite many petrographic investigations of beachrock cements, the processes responsible for beachrock formation are still poorly understood. Given the likelihood of cement diagenesis in the beach environment, there is a need to pursue other research methods. For example, the subsurface formation of beachrock should be tracked on a variety of beaches over an extended period. The processes affecting beach sand cementation should also be examined under controlled conditions in a laboratory setting. Preliminary experiments conducted by Turner (1995) indicate that the addition of dissolved nitrate or organic carbon to beach sand microcosms stimulates bacterial growth and the precipitation of intergranular calcium carbonate. This leads to the question as to whether coastal discharges of groundwater contaminated with fertilizers or human wastes are increasing the rate and geographic range of beachrock formation.

Cross-References

- [Beach Features](#)
- [Coral Reef Coasts](#)
- [Eolianite](#)
- [Rock Coast Processes](#)
- [Sea-Level Indicators, Geomorphic](#)

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Beaufort Wind Scale

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Definition

The Beaufort wind scale relates wind speed to the physical appearance of the sea surface and to the effects and damages on land. The sea appearance factors include apparent wave height, the prominence of breakers, whitecaps, foam, and spray. It is the oldest method of judging wind force effects on sea surface and land. The effects on land include damages and destruction of trees, homes, and utilities. Originally devised by Admiral Sir Francis Beaufort (1774–1857) of the British Navy in 1805 to simplify the signaling of wind and weather conditions between sailing vessels, it has since been repeatedly modified to make it more relevant to modern systems of weather warning and navigation.

An updated modern version of the Beaufort scale is shown in Table 1. This is adapted from British Admiralty (1952) and Thomson (1981). The Beaufort number and the corresponding lower and upper wind speeds are plotted in Fig. 1. A second-order polynomial fit yields the following equations:

$$U_U = 0.14B^2 + 1.44B, \quad (R^2 = 0.99) \quad (1)$$

$$U_L = 0.17B^2 + 0.78B, \quad (R^2 = 0.99) \quad (2)$$

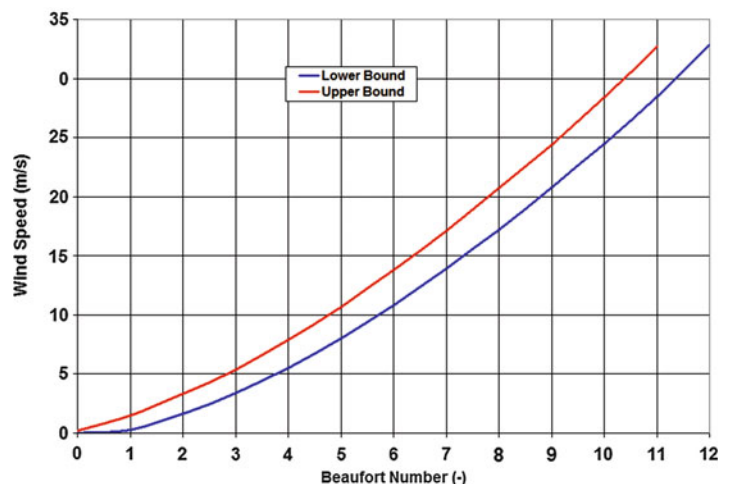
where U_U and U_L are upper- and lower-bound wind speeds (m/s), respectively, B is Beaufort number, and R is regression coefficient.

There is also an elaborate version of the scale proposed by Meyers et al. (1969) based on the works of British Admiralty (1952), McEwen and Lewis (1953), and Pierson et al. (1953). Wind speed measured at 36 ft above sea surface is usually applied to use the scale, which for all practical purposes is assumed to represent the wind at 10 m. The wave heights are approximate and represent fully arisen sea state.

As with any subjective judgmental method, the Beaufort scale is far from perfect. Similar subjective scales have been proposed to assess hurricane and tornado damages. The Saffir-Simpson Hurricane Scale (Saffir 1973; Simpson 1974) is used for rating the severity of damages caused by hurricane-scale storms. The Fujita Tornado Scale (F-scale) was proposed in 1971 (Fujita 1971) by Tetsuya Fujita of the University of Chicago for rating the damage severity of tornadoes. Both the scales have gone through multiple

Beaufort Wind Scale, Table 1 Beaufort wind scale

Beaufort No.	Name	Wind speed (knot)	Wind speed (m/s)	Effect of wind on sea surface	Significant wave height (m)	Effect of wind on land
0	Calm	<1	0.0–0.2	Like a mirror	0	Still, smoke rises vertically
1	Light air	1–3	0.3–1.5	Ripples form with the appearance of scales, but without foam crests	0.1–0.2	Smoke drifts, vanes remain motionless
2	Light breeze	4–6	1.6–3.3	Small wavelets, crests appear glossy but no breaking	0.3–0.5	Leaves rustle, vanes move, wind can be felt on face.
3	Gentle breeze	7–10	3.4–5.4	Larger wavelets begin to break, some scattered white horses	0.6–1.0	Constant movement of leaves and small twigs, flags begin to stream
4	Moderate breeze	11–16	5.5–7.9	Small waves predominant but fairly frequent white horses	1.5	Dust and loose paper are lifted, thin branches move
5	Fresh breeze	17–21	8.0–10.7	Moderate waves, distinctly elongated, many white horses, chance of spray	2.0	Small trees in leaf begin to sway
6	Strong breeze	22–27	10.8–13.8	Long waves with extensive white foam, breaking crests, spray likely	3.5	Large branches move, power lines whistle, stop lights sway, umbrellas difficult to control
7	Moderate gale	28–33	13.9–17.1	Sea heaps up and white foam from breaking waves begins to be blown in streaks, spindrift begins to be seen	5.0	Entire trees sway, some resistance to walkers, car feels force of wind
8	Fresh gale	34–40	17.2–20.7	Moderately high waves of greater lengths, edges of crests break into spindrift, foam is blown into well-marked streaks	7.5	Twigs break off trees, difficult walking against wind
9	Strong gale	41–47	20.8–24.4	High waves, rolling sea, dense streaks of foam, spray may affect visibility	9.5	Roof tiles lifted off, windows may be blown in, trees may topple
10	Whole gale	48–55	24.5–28.4	Very high waves with long overhanging crests, foam in great patches blown in dense white streaks downwind, heavy rolling sea causes ships to slam, visibility reduced by spray, sea surface takes on whitish appearance	12.0	Trees uprooted considerable structural damage to some buildings
11	Storm	56–66	28.5–32.7	Exceptionally high waves, sea covered with long white patches of foam blown downwind, wave crests blown into froth everywhere, visibility impeded by spray	15.0	Widespread damage, extensive flooding in low lying areas if wind is directed onshore
12	Hurricane	>66	>32.7	Air filled with foam and spray, sea completely white, visibility seriously impaired	>15.0	Severe structural damage to buildings, widespread devastation and flooding

Beaufort Wind Scale, Fig. 1
Upper- and lower-bound wind speeds corresponding to the Beaufort wind scale numbers

modifications and updates – the latest updates were implemented in the 2012 and 2007 for the Saffir-Simpson Scale and the Fujita Scale, respectively. More on the modified F-scale known as the Enhanced Fujita scale (EF-scale) can be found in <http://www.theweathernetwork.com/>

Cross-References

- [Climate Patterns in the Coastal Zone](#)
- [Coastal Climate](#)
- [Meteorological Effects on Coasts](#)
- [Nearshore Wave Measurement](#)
- [Wave Hindcasting](#)

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Bioconstruction

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The term, *bioconstruction*, usually refers to a bioconstructed limestone that has been built-up by colonial and sediment-binding organisms including algae, corals, bryozoans, and stromatoporoids. The term, bioconstructed limestone, was introduced by Carozzi and Zadnik (1959) in their study of the Silurian Wabash reef in southern Indiana. The word, bioconstructed, was used to distinguish the limestones and dolomites which were found in a reef from the dolomitic calcarinites preserved in the reef flanks and the dolomitic shales in the country rock (Carozzi and Zadnik 1959). The term, bioconstruction, was next applied to Devonian stromatoporoid reefs in the Beaverhill Lake Formation, Upper Devonian, Alberta Canada (Carozzi 1961).

European Use of the Word Bioconstruction

The word, *bioconstruction*, was widely accepted and used in European geologic journals, but has not appeared in any North American journals since 1961. The European use of the term, *bioconstruction*, includes what the North American geologists would refer to as reefs, bioherms, and biostromes. Based on living coral reefs, Ladd (1944) defined a reef as a rigid, wave-resistant framework constructed by large skeletal organisms. A broader definition of a reef as “a discrete carbonate structure formed by in-situ organic components that develops topographic relief upon the seafloor” has been proposed by Wood (1999, p. 5). Cumings (1930) defined a bioherm as a mound-like, dome-like, lenslike, or reef-like mass of rock built-up by sedentary organisms (such as corals, algae, foraminifera, mollusks, gastropods, and stromatoporoids), composed almost exclusively of their calcareous remains and enclosed or surrounded by rock of different lithology. A biostrome is defined as a distinctively bedded and widely extensive lenticular, blanket-like mass of rock built by and composed mainly of the remains of sedentary organisms and not swelling into a mound-like or lens-like form; an organic layer, such as a bed of shells, crinoids, or corals, or a modern reef in the course of formation, or even a coal seam (Cumings 1930).

Types of Bioconstructions

Examples of several different types of bioconstructions, which would fall into the categories of reef, bioherms, and biostromes, are included to show how the term bioconstruction is used in the European literature. In Spain, rugose corals and calcareous algae bioconstructions are also called biostromes (Rodríguez and Sanchez 1994). In Jurassic and Cretaceous strata in Germany, Rehfeld (1996) describes different forms of sponge bioconstructions which comprise bioherms, biostromes, and sponge meadows. The wave resistant calcisponge and algal reefs of the Capitan reef facies, partially wave resistant reef mounds and non-wave resistant skeletal mounds in the Guadalupe Mountains of New Mexico, are described as Permian bioconstructions (Noe 1996). Therefore, bioconstruction is a general term for limestone and dolomite deposits formed by colonial and sediment binding organisms which include reefs, bioherms, and biostromes (Table 1).

Conclusions

Bioconstruction is distinctly a European term for a limestone which has been built-up by colonial and sediment

Bioconstruction, Table 1 Beaufort wind scale

Beaufort No.	Name	Wind speed knot	m/s	Effect of wind at sea surface	Significant wave height (m)	Effect of wind on land
0	Calm	<1	0.0–0.2	Like a mirror	0	Still, smoke rises vertically
1	Light air	1–3	0.3–1.5	Ripples form with the appearance of scales, but without foam crests	0.1–0.2	Smoke drifts, vanes remain motionless
2	Light breeze	4–6	1.6–3.3	Small wavelets, crests appear glossy but no breaking	0.3–0.5	Leaves rustle, vanes move, wind can be felt on face
3	Gentle breeze	7–10	3.4–5.4	Larger wavelets begin to break, some scattered white horses	0.6–1.0	Constant movement of leaves and small twigs, flags begin to stream
4	Moderate breeze	11–16	5.5–7.9	Small waves predominant but fairly frequent white horses	1.5	Dust and loose paper are lifted, thin branches move
5	Fresh breeze	17–21	8.0–10.7	Moderate waves, distinctly elongated, many white horses, chance of spray	2.0	Small trees in leaf begin to sway
6	Strong breeze	22–27	10.8–13.8	Long waves with extensive white foam, breaking crests, spray likely	3.5	Large branches move, power lines whistle, stop lights sway, umbrellas difficult to control
7	Moderate gale	28–33	13.9–17.1	Sea heaps up and white foam from breaking waves begins to be blown in streaks, spindrift begins to be seen	5.0	Entire trees sway, some resistance to walkers, car feels force of wind
8	Fresh gale	34–40	17.2–20.7	Moderately high waves of greater lengths, edges of crests break into spindrift, foam is blown into well-marked streaks	7.5	Twigs break off trees, difficult walking against wind
9	Strong gale	41–47	20.8–24.4	High waves, rolling sea, dense streaks of foam, spray may affect visibility	9.5	Roof tiles lifted off, windows may be blown in, trees may topple
10	Whole gale	48–55	24.5–28.4	Very high waves with long overhanging crests, foam in great patches blown in dense white streaks downwind, heavy rolling sea causes ships to slam, visibility reduced by spray, sea surface takes on whitish appearance	12.0	Trees uprooted considerable structural damage to some buildings
11	Storm	56–66	28.5–32.7	Exceptionally high waves, sea covered with long white patches of foam blown downwind, wave crests blown into froth everywhere, visibility impeded by spray	15.0	Widespread damage, extensive flooding in low lying areas if wind is directed onshore
12	Hurricane	>66	>32.7	Air filled with foam and spray, sea completely white, visibility seriously impaired	>15.0	Severe structural damage to buildings, widespread devastation and flooding

binding organisms such as algae, corals, bryozoans, and stromatoporoids. It combines what North American geologists would refer to as reefs, bioherms, and biostromes.

Cross-References

- [Atolls](#)
- [Bioerosion](#)
- [Bioherms and Biostromes](#)
- [Coral Reefs](#)
- [Reefs, Non-coral](#)
- [Tidal Environments](#)

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Bioengineered Shore Protection

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In an effort to arrest shore erosion at many coastal locations and to provide protection to marinas and harbors, it may be necessary to construct structures in high wave energy zones. Current practice involves utilization of structures constructed using large armor stones, concrete and steel walls, and a variety of other “hard” engineering techniques. Quite often, these structures do not add to the aesthetic and recreational attributes of a site and may impact significantly on the local environment. Integration of bioengineered components into the design of breakwaters and shore protection systems can be utilized, in certain cases, to enhance the project by providing better biological habitat and ancillary water quality improvement. Thus the goal of a project changes to include not only the stabilization of the eroding area or the provision of “quiet” waters, but to increase the quantity and quality of habitat available to fish and waterfowl communities, while providing an effective and aesthetic control of natural environment.

Background

In both the engineered and natural environment, the flow of water often causes erosion. The causes must be understood before the problem can be addressed. In the coastal zone, the flow of water results from wave action, the associated runup and backwash, wave breaking, alongshore currents, and the natural flow of water along side and overtop of the high-tide shoreline. In addition to the interaction of the high-tide shoreline or lakeward structure with water, a considerable amount of animal and human activity create additional stresses on the high-tide shoreline. Bioengineering methods of shore

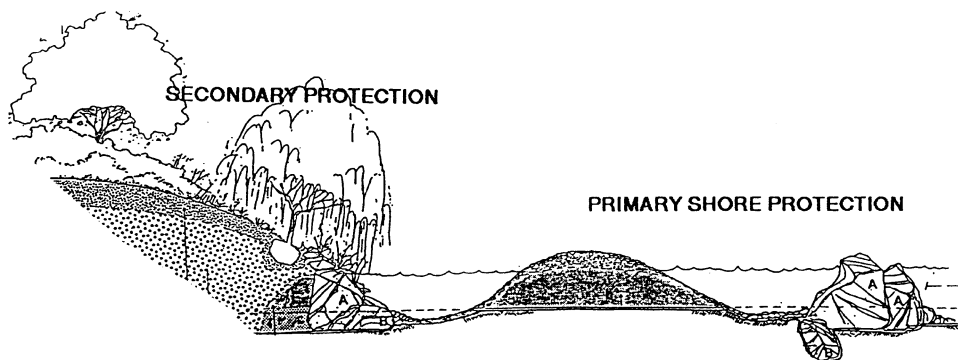
protection offer a practical solution that can also create an aesthetically pleasing and environmentally beneficial “buffer zone.” Bioengineering, in this context, is the utilization of vegetation, either by itself or in combination with other defense mechanisms, depending upon the local environment. The other defense mechanisms may include the use of rock lining, offshore islands, wave screens, and submerged shoals that limit the wave energy reaching a site. Quite often, these defense structures can be designed to provide significant enhancement to the environment, particularly in providing suitable fish habitat for spawning, feeding, and hiding from predators.

The value of vegetation for protecting the soil depends on the combined effects of roots, stems, and foliage. Roots and rhizomes reinforce the soil. Immersed foliage elements absorb and dissipate energy and may cause sufficient interference with the flow to prevent scour. In a sediment-laden environment, they may also promote deposition.

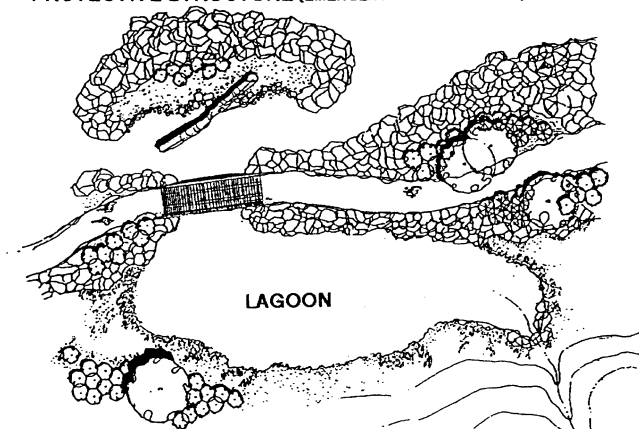
A coastline requiring protection can be considered as two separate areas and thus habitat enhancement can be geared toward two communities; the high energy nearshore environment and the onshore environment, which can be suitably modified to ensure low wave energy levels. Enhancement of the nearshore zone can include construction of rock revetments as reef habitat, inclusion of submerged offshore structures to reduce wave energy levels reaching the shore and primary wave defense structures which provide habitat enhancement potential by the nature of their design (Fig. 1). Selection of stone and design of its placement is developed in a manner to provide a reef like habitat beyond minimum stone placement required for the minimal shore stabilization. The effectiveness of both natural and artificial reef like habitats as fish community habitats has been well documented. Proper design and installation of rock will provide protective cover and feeding areas and will supply the needs for small aquatic and benthic organisms by providing protection from the high energy wave action and from larger predators.

Low wave energy areas can be created behind a primary defense such as an offshore rock structure or wave screen, or through the creation of lagoons behind stable control structures (Fig. 2). The development of constructed wetland pockets and areas for other shallow water plants can occur in these lagoons. This can be promoted by the establishment of “biological” riprap in the form of brush and woody plant debris. These materials will provide a setting that will foster the accumulation of shore plants from wind blown seed banks. As the brush decomposes, it provides a limited release of nutrients to the developing plant community and is eventually replaced by living plants. The establishment of new habitat will provide the opportunity for colonization by wetland plant and animal species that require quiescent waters. Transient use of the habitats by a variety of aquatic and migratory waterfowl is an additional potential for these environments.

Bioengineered Shore Protection, Fig. 1 Utilization of shoals



PROTECTIVE STRUCTURE (EMERGENT OR SUBMERGENT)



Bioengineered Shore Protection, Fig. 2 Development of a lagoon cell

Goals and Objectives

The designer is encouraged to consult specialists in the fields of coastal hydraulics, fisheries, geomorphology, biology, landscape architecture, or any field that could make the project a success. The design of bioengineered breakwaters and shore protection that functions environmentally requires a multidisciplinary approach. Usually, no individual has all the expertise required to ensure successful implementation.

The following geomorphologic, hydraulic, and biological changes may occur as a result of modification of the shore, which would occur from the creation of a marina or harbor, or from local erosion protection schemes:

- Loss or elimination of aquatic vegetation
- Loss or elimination of backshore vegetation
- Removal of specific nearshore bathymetric features
- Modified substrate conditions
- Modified hydrodynamic, flow, sediment, and water quality regimes
- Changes in nutrient conditions and reductions in food organisms
- Aesthetic degradation

- Reductions in habitat diversity and environmental stability
- Increased water temperatures.

The shore is a dynamic system where impacts are difficult to predict. Engineered structures, when properly designed and constructed, can provide both species and habitat diversity and thereby mitigate potential adverse changes. However, the goals and objectives of the shore protection design must be correctly identified early in the design process. The designer must be aware of the design goals and objectives to correctly identify, size, and locate the various functional elements within the system. Biodiversity within and adjacent to the shore is interrelated with the quality in updrift and downdrift areas. Changes to any one of these components may adversely impact on others.

Habitat Requirements

Aquatic life generally requires a habitat that contains the following:

1. Sufficient water depth and volume for each life stage.
2. Adequate water quality with preferred ranges of temperature, dissolved oxygen, PH, etc.
3. A variety of continuous hydrodynamic conditions varying from deep water to shallow water for breeding and cover. Also flow conditions that sort bed load materials to provide a good environment for bottom dwelling organisms are advantageous.
4. Adequate cover to provide shade, concealment, and orientation.
5. Adequate food to maintain metabolic processes, growth, and reproduction.

Shore improvements should be designed for the individual fish species. Specific requirements for reproduction, juvenile rearing, and adult rearing with regard to feeding location, concealment from predators and competitors, and sanctuary from flow extremes and ice formation varies between species. Loss of the natural bathymetric features, which are utilized by particular species as a result of implementation of shore

protection, could eliminate many of the requirements necessary to sustain significant biodiversity along the nearshore area. In addition, removal of existing shore vegetation, in either the emergent or submergent zones would significantly reduce or eliminate the potential to sustain a fish population.

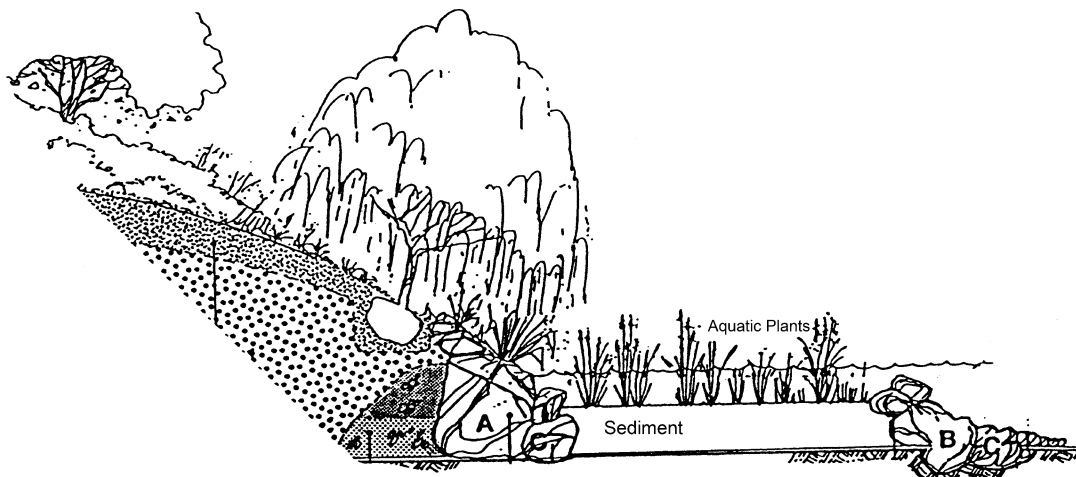
Utilizing Vegetation

In certain low wave energy environments, vegetation may be used by itself to provide suitable protection to an eroding shore. Reeds and other marginal plants can form an effective buffer zone by absorbing wave energy and restricting the along-shore flow velocity adjacent to the shore. They therefore have a protective value. Specific functions that they can perform include:

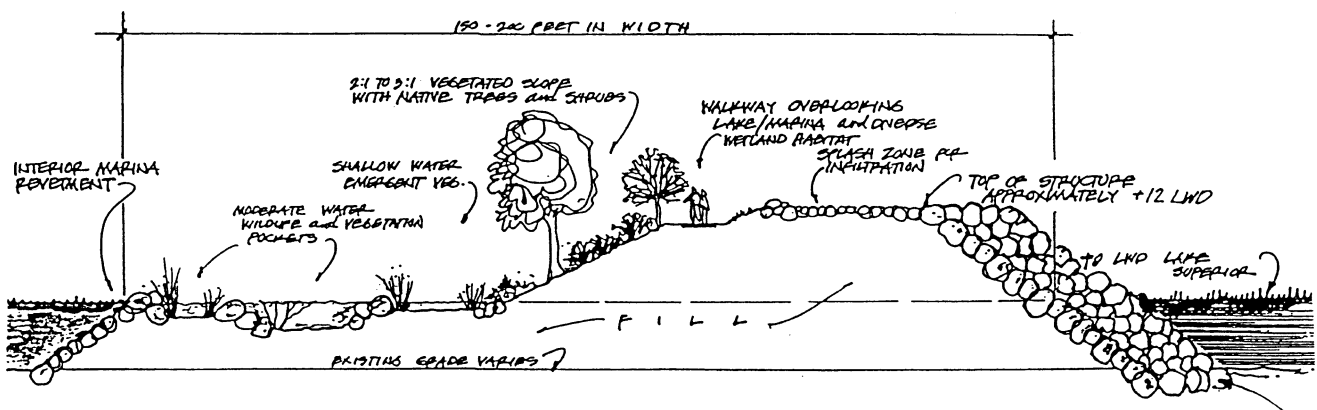
1. Absorbing and dissipating wave-wash energy.
2. Interference and protection of the shoreline bank from the flow.
3. Reinforcement of the surface soil through the root mat and prevention of scour of the bank material.
4. Sediment accumulation brought about by the dense plant stems.

Marginal plants require very wet ground and generally will not survive in water that is more than 0.5 m deep for long period of time. They flourish in conditions of low flow velocity and their integrity is weakened by wave action in excess of 0.5–0.75 m. Different species offer different levels of protection with regard to wave energy dissipation. For incident wave conditions under 0.5 m, reed beds having a width of 2–2.5 m may dissipate 60–80% of the incoming wave energy. In areas with higher levels of wave energy, riprap and geotextiles may be used in conjunction with vegetation to provide effective bank protection (Fig. 3). In areas of high incident wave energy, an area of low wave energy can be created behind a primary defense such as an offshore rock structure or wave screen, or through the creation of lagoons behind stable control structures (Fig. 4), as described above.

Natural methods of protection generally have low capital cost in comparison with conventional engineering methods. However, they may well have higher recurrent cost due to regular inspection, trimming and cutting, and repair. In areas where a combination of conventional and bioengineered



Bioengineered Shore Protection, Fig. 3 Emergent vegetation used in conjunction with stone



Bioengineered Shore Protection, Fig. 4 Control structure/lagoon

structures are required, recent experience at several sites on the Great Lakes has established that these techniques may cost 20–30% more than conventional techniques alone.

Possible disadvantages are that natural protection schemes take time to mature and to become fully effective. Depending on the type, natural protection may take several growing seasons to reach the desired standard of protection.

Bioengineering differs from other conventional forms of engineering in two key respects, which strongly influence the design approach:

1. Bioengineering involves considerable practical experience and judgment, as opposed to the application of quantitative design theory or rules.
2. Careful management is required not only in the establishment of vegetation, but also in its aftercare over the initial growing seasons.

Use of Vegetation Requires the Following Points to be Considered

The principal plant groups that can be used are aquatic plants, grasses, shrubs, and trees. Selection is based on consideration of the different roles to be performed by the vegetation, taking into account the physical and chemical properties of the soil, the climatic conditions, and the soil/water regime under which the plant must survive. Vegetation establishment may take several growing seasons and is a seasonal activity that must be managed and maintained. The engineer must prepare and agree to specific management objectives and a management program with the owner/client. This is in order to ensure that the vegetation is maintained in a fit condition to perform its intended roles.

Zones and Horizons of Natural Protection

With natural methods of protection, and particularly methods involving the use of live material, the effectiveness of different materials is strongly dependent on their location in relation both to the dominant external water level and to the subsoil soil/water regime. To achieve effective protection using natural materials, the designer will almost inevitably need to use different methods of protection in different zones and horizons of the shore (Coppin and Richards 1989).

Use of Reeds

The emergent and marginal types of aquatic plants, such as the common reed, bulrush, and great pond sedge, are frequently used for interference and protection purposes to form a protective margin along the shore at the waterline.

They also encourage siltation by absorbing current flow energy, and thus reducing the sediment-carrying capacity of the flow. Reeds can be easily weakened by erosion and loosening of the soil around the rhizomes due to wave energy. It is therefore necessary to protect the zone containing roots from high-velocity flow or significant wave attack. Provided this is done, the stems and leaves will protect the shore bank above.

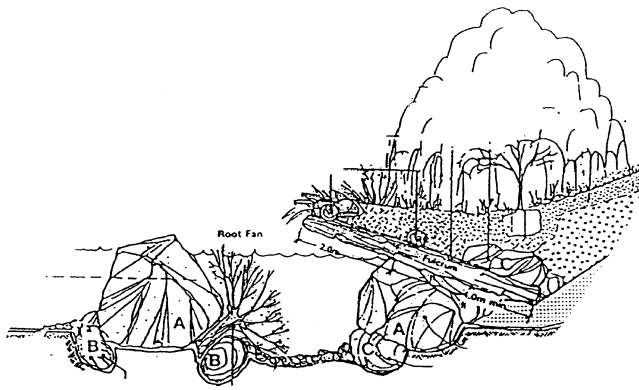
Uses of Shrubs and Trees

A limited range of trees are water-tolerant and can be used in bioengineering structures for bank protection in both the aquatic and damp zones. The willow, alder, and black poplar are the principal water-tolerant species. In particular, a dense root structure is able to provide some protection as well as substantial reinforcement effect to enhance the stability of the shore both above and below the mean water level. The willow and poplar are particularly useful for bioengineering because they can be propagated from cut limbs. The cut limbs can be placed such that secondary root growth develops and shoots sprout from dormant buds. Trees, which are not water-tolerant, do not have any major direct function in shore stabilization, although they may provide shade to control the growth of aquatic life as discussed earlier.

Use of Grasses

Grass is used very extensively in bank protection in the zones above the high water level. Grass roots cannot tolerate prolonged submergence periods. A wide variety of grass species and mixtures therefore are appropriate to satisfy the functional, environmental, and management requirements for a protection scheme. The principal functions which grass fulfills are those of interference, protection, root reinforcement, and soil restraint. The surface root structure forms a composite soil/root mat, which enhances the erosion resistance of the bare subsoil, and which is anchored into the subsoil by deeper roots. The engineering function of grass may be augmented by the use of geotextile or cellular concrete reinforcement to form composite protection. With both types of reinforcement, the visual effect of grass is retained. Erosion of grass cover by wave runup generally occurs by the scouring of soil from around the roots of a plant, thereby weakening its anchorage until the plant itself is removed by the drag of the flowing water. The effectiveness of grass protection can also be seriously reduced by any localized patches of bare soil or poor grass cover.

The rate of growth of different grasses varies considerably. Complete grass cover should normally be achieved by the middle of the first growing season while full protective strength of the sward is reached during the second season. Provision should be made for aftercare including mowing, fertilizing, and weed control.



Bioengineered Shore Protection, Fig. 5 Timber (CANTILEVER) to provide habitat and shade

Use of Timber and Woody Material

A variety of timber and other dead woody materials can be used in the shore protection scheme usually fulfilling reinforcement, protection, and sometimes drainage functions (see Fig. 5). Natural hardwoods will retain their integrity for 5–10 years if built into the bottom of a bank below the water level. Out of the water they can last longer but the worst environment for timber is the alternately wet and dry zone around mean water level.

Monitoring

As part of the project design for the shore stabilization enhancements, a monitoring program is required. The purpose of the monitoring program is to measure the success and applicability of the enhancement methods to other shore projects.

Baseline habitat conditions should be assessed by observation and characterization of existing conditions. A plant survey and macroinvertebrate sampling of the near-shore benthic environment and a terrestrial plant survey should be performed to document existing plant and animal populations. Incidental observations of birds should be made as part of fieldwork. Sampling of nearshore fish populations should be coordinated with local regulatory agencies. Post-construction monitoring of the establishment of biological communities should be completed to evaluate the success of a particular scheme.

Conclusions

Shore protection enhancements similar to those described in this entry have been successfully implemented at numerous sites on the Great Lakes, most notably in Canada at Red Rock Marina, Lake Superior; Thunder Bay

Harbor, Lake Superior; Kingston, Lake Ontario; and various reaches of the St. Lawrence Seaway, and at Bender Park, Lake Michigan; Silver Bay, Lake Superior; and in Louisiana (Gulf of Mexico) in the United States. The range of design wave conditions range from 0.75 to 4.5 m at these various sites. Many other projects are in the process of implementation.

Utilization of bioengineered shore protection, in concert with virtually transparent offshore protection (submerged breakwaters, wave screens, etc.) can provide for significant levels of protection while maintaining the natural beauty of an area, and, in most circumstances, providing significant opportunities for habitat enhancement and increased biodiversity.

Further suggested reading may be found below.

Cross-References

- [Beach Erosion](#)
- [Coastal Management Practices](#)
- [Geotextile Applications](#)
- [Monitoring Coastal Ecology](#)
- [Shore Protection Structures](#)
- [Vegetated Coasts](#)
- [Wetland Restoration](#)

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Bioerosion

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In his study of the erosion of steep cliffs around Huntington Sound in Bermuda by excavating sponges, Neumann (1966) defined the term bioerosion as the removal of consolidated material or lithic substrate by direct action of organisms. Soon after the term, bioerosion, was introduced, geologists and biologists described many different types of bioeroding

organisms including algae, bacteria, foraminifera, sponges, bryozoa, annelid worms, barnacles, gastropods, bivalves, echinoderms, fish, and mammals. The process of bioerosion was also reported from many different marine and non-marine environments ranging from mountain slopes to the tops of deep sea knolls, and from rocky intertidal zones and coral reefs to the flanks of continental shelves. Bioerosion has also been reported from climatic zones extending from tropical and subtropical to the subarctic and arctic. Several different types of experiments have been devised for studying the rates of bioerosion by different types of organisms and in different environments. Although bioerosion was first recorded for living sponges in the intertidal zone, evidence for bioerosion was also found in the ancient rocks extending back at least to the Silurian.

Bioeroding Organisms

Several different types of microbial borers have been described from modern environments and ancient rocks. Microbial borers including cyanobacteria and chlorophytes were found in modern reef environments at depths between 0 and 230 m, and boring heterotrophs are present between 100 and 300 m (Vogel et al. 1996). Evidence for boring algae (cyanophyta) has been preserved in Silurian bivalves and may be responsible for the silicification of their shells (Liljedahl 1986). Twenty species of foraminifera, ranging age from Jurassic (Callovian) to Recent, are known to make cavities in hard substrates (Venec 1996). The bioeroding foraminifera were found in turbulent, warm, shallow-water environments.

A wide variety of living and fossil invertebrates have been identified as bioeroders. Several species of boring sponges have been reported from reef areas in Bermuda (Neumann 1966) and Grand Cayman Island in the British West Indies (Acker and Risk 1985), and on a deep sea knoll at depths of 1,600–1,800 m (Boerboom 1996). The bioeroding mollusks include chitons, gastropods, and bivalves. *Chiton pelliserpentis* removed hardened mudrock during feeding at Mudrock Bay in Kaikoura, New Zealand (Horn 1984). The spawn of the gastropod, *Nerita*, settled on the sea bottom and eroded carbonate rocks at Cathedral Point in Costa Rica (Fischer 1980). The bivalve genus *Lithophaga* was an active chemical borer in reefs from the Carboniferous through the Eocene (Krumm 1992). Rock-boring echinoids excavated large cavities in reefs in the South Florida keys (Kues and Siemers 1974) and on Enewetak Atoll in the Marshall Islands (Russo 1980). The rock-boring barnacle, *Lithotyra*, eroded the rock face while grazing in the intertidal zone (Ahr and Stanton 1973). The polychaete annelid, *Eunice*, burrowed into the carbonate rocks along the shore of the Gulf of California.

Vertebrates including fish and mammals play an important role in bioerosion on reefs and mountain slopes. Parrotfish have been observed feeding on coral in reef environments and their rates of bioerosion were measured (Frydl and Stearn 1978). Recolonization experiments on coral reef communities near Aquaba on the Red Sea demonstrated that herbivorous fish were a major factor in structuring coral reef communities (van Treeck et al. 1996). In the Pyrenees of Spain, the indirect effect of digging by small mammals was considered more significant than the direct detachment of soil cover (Martines and Pardo 1990).

Rates of Bioerosion

Several different field experiments have been used to estimate the rates of bioerosion by different organisms and in different environments. On the carbonate coastline of Bermuda, experiments show that the sponge *Cliona lampa* is capable of removing 6–7 kg of material from 1 sq. m of carbonate substrata in 100 days, corresponding to an erosion rate of calcarenite of more than 1 cm per year (Neumann 1966). In Kaikoura, New Zealand, *Chiton pelliserpentis* removed mudrock from the surface at a rate of 47.3 g/sq. m on the high shore and 173 g/sq. m on the low shore (Horn 1984). This was equivalent to about 2% of total on the high shore and 5.5% on the low shore. On Moorea reef barrier flat in French Polynesia, bioerosion rates for echinoids was estimated at 4.5 kg/sq. m per year and for scarid fish at 1.7 kg/sq. m per year (Peyrot et al. 1996).

Environments of Bioerosion

Although most examples of bioerosion have been studied from tropical reefs and intertidal zones, bioerosion also has been reported from high-latitude environments, the outer continental shelf and deep-sea knolls. Algae borings were found in gastropod shells and echinoderm tests in the high-latitude, low-energy environments in the firths of Clyde and Lorne, Scotland (Akpan and Farrow 1985). Boring sponges were dredged up from Newfoundland from depths of approximately 1,600–1,800 m on top of Orphan Knoll, 550 km northeast of Saint John's (Boerboom 1996). Evidence for bioerosion was also found in the clastic sediments on the outer continental shelf around the Hudson Canyon off the eastern coast of the United States (Twichell et al. 1984). A workshop on bioerosion convened by Bromley (1999) has reviewed several different aspects of bioerosion ranging from the style of bioerosion in Late Jurassic reefs to the role of bioerosion in carbonate budgets in Indo-Pacific reefs.

Conclusions

Bioerosion by microorganisms, invertebrates, and vertebrates is widespread throughout many different carbonate environments from the early Paleozoic to the Recent. Bioeroding organisms have been reported from mountain slopes to deep-sea knolls, from the rocky intertidal zone to coral reefs and from the tropics to the arctic circle. The rates of bioerosion vary from a few grams per square meter to several kilograms per square meter depending on the organisms involved and the depositional environments.

Cross-References

- Atolls
- Bioconstruction
- Cliffs, Erosion Rates
- Coral Reefs
- Erosion Processes
- Karst Coasts
- Tidal Environments

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Biogeomorphology

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Biogeomorphology is a discipline that combines ecology and geomorphology. Geomorphology is the study of landforms and their formation. Ecology is the study of the relationships between biota and their environment. The environment is defined as factors that affect biota. These factors can be abiotic (physical, chemical), biotic (other organisms), or anthropogenic (humans). Abiotic geomorphological processes may affect biota and biota may in turn affect geomorphological processes. The interaction between both defines the discipline of biogeomorphology. *Biogeomorphology is the study of the interaction between geomorphological processes and biota.*

Essential Concepts

The term biogeomorphology was first used in the 1980s (Viles 1988), although earlier studies have been conducted that were focused on biogeomorphology without using this term. Biogeomorphology is studied in terrestrial as well as in aquatic systems. In coastal systems biogeomorphological interactions are clearly demonstrated in the shallow, productive waters, and in various sedimentary environments. Examples of biogeomorphological interrelationships include sand dune development, tidal flats, salt marshes, mangrove systems, and coral reefs.

Relevant geomorphological factors in coastal systems are bathymetry, bed composition (rock, gravel, sand, silt), and the transport of sediment. It also includes factors that drive morphological processes, such as water flow and wave energy. The biota involved in coastal biogeomorphology include plants and animals, ranging from very small (algae) to very large (whales).

The geomorphological influence on biota is in its most direct form the influence on habitats (living environments) of flora and fauna. The coastal morphology and geomorphological processes define the gradients between high and low, between wet and dry, and between sedimentation and erosion. These gradients and the processes that cause them are determinative for gradients in grain size of the sediment, nutrient levels, organic matter levels, and moisture. Plants and animals are tuned to specific conditions and will therefore be abundant in specific locations.

The biological influence on geomorphological processes is the influence of biota to create, maintain, or transform their own geomorphological surroundings. This is demonstrated by the influence of vegetation on the hydraulic resistance, erodability and sedimentation, or by the influence of fauna on sediment characteristics through bioturbation and biostabilization.

In some cases morphological processes are dominant over biological processes and therefore the biota have to adjust to their environment. In other cases biological processes are dominant. The most interesting are those cases where there is a mutual interaction that leads to feedback coupling of processes. When looking for these cases, it is important to examine the temporal and spatial scales of the mutually interacting processes. Biogeomorphological interrelationships can be found in several coastal environments, for both hard and soft substrates.

Biogeomorphology for Hard Substrates

On rocky shores and coral reefs a typical community of organisms thrives that affects the erosion rates of its substrate. Influenced by abiotic factors such as wave energy, splash water, inundation frequency and -period, depth, desiccation and substrate type, a clear zonation can be found of various cyanobacteria, (macro-)algae, fungi, lichens, molluscs, sponges, worms, sea urchins, fish, etc. Some of these organisms dwell on the surface of the substrate, while others live within the substrate. Their effect on erosion of the substrate is divided in “biological corrosion,” processes that modify the substrate but provides no erosion product, and “biological abrasion” (see “► [Bioerosion](#)”), processes that do generate an erosion product. Grazing, burrowing and boring on or in the substrate carries out biological abrasion, and is most significantly found in coral reef systems.

Biogeomorphology for Soft Substrates

In soft coastal systems, the interrelationships between geomorphological factors and biota can mainly be noticed for benthic fauna and flora. The presence of benthic species is affected by hydraulic and morphologic conditions, such as depth, current velocity, salinity, and grain size. The effect of soft substrate communities on geomorphology is divided into biostabilization and biodegradation. Biostabilization leads to an increase in soil resistance, preventing erosion, while biodegradation leads to an increased erodability.

Biostabilization by Plants

On tidal flats, small algae (diatoms) are capable of affecting the geomorphology. These diatoms can form extensive algal mats and excrete EPS mucus, which is a sticky substance made of polysaccharides that glues the sediment together and therefore protects the sediment against erosion. Sea grass is dependent on clear water, it needs sunlight to grow. A sea grass meadow slows down the current velocity near the bed and therefore sand and silt will not resuspend in the water, which otherwise would lead to turbid water. Furthermore, their root system binds the substrate. Ultimately, deposition of suspended sediment is encouraged in a sea grass meadow, which leads to the supply of organic material with nutrients, needed for growth.

Seaweeds are also capable of adjusting their physical environment by damping down wave energy; and salt marshes also play an important role in stabilizing sediments. Salt marsh vegetation makes fine sediment settle down resulting in a continuous heightening of the marsh. The higher the marsh gets, the more vegetation can grow and the better the marsh is protected against erosion. Other stabilizing effects result from cementation of beachrock by cyanobacteria and stromatolite formation by algae.

Biostabilization by Animals

Some macrozoobenthos can actively catch sediment particles from the water column and bring them to the bed. The presence of a mussel bank, for example, will alter the bed in different ways. Mussels slow down the water flow and they protect the bed against erosion. Mussels also actively catch small particles from the water column by filterfeeding and subsequently excrete these as pseudofeces. This results in a change in the soil composition to finer sediments.

Animal tube fields are also believed to stabilize the sediment, because there is a clear accumulation of fine particles and organic matter between the tubes. The tube itself may affect small-scale turbulence and therefore have a stabilizing effect, however, a great deal may be attributed to the community of microorganisms between the tubes that excrete mucus.

Other stabilizing effects result from large banks of dead shells and mucus binding by meio- and macrofauna.

Biodestabilization

Benthic fauna may destabilize the substrate by their digging and feeding activities (bioturbation). The constant mixing and recycling of sediment in the top centimeters of the bed results in a characteristic vertical particle-size profile. The selective uptake and excretion of preferred particle sizes results in sorting and pelletizing sediments. Together with the digging of burrows and the constant movement within the substrate, these activities lead to the generation of a surface micro-relief that has a higher hydraulic roughness and is more prone to erosion. Furthermore, bioturbation also affects the sediment water content, porosity, and sediment cohesion.

Scale Interactions in Biogeomorphology

Different physical and biological processes can have dynamic interactions when they operate on the same spatial and temporal scales. Processes that act on a very small scale may appear as noise in the interactions with processes on larger scales. Their effect can be accounted for by proper averaging procedures (e.g., for turbulence). Processes that act on a large-scale may be treated as slowly varying or even constant boundary conditions when studying their effects on processes on smaller scales (e.g., sea-level rise due to climate change). Techniques for scale interactions are reasonably well established in geomorphology (De Vriend 1991) and are based on scale linkage via sediment transport. In biology, however, population and community dynamics give rise to spatial and temporal structures that are not easily linked. In recent years, the importance of scale has been increasingly recognized (Legendre et al. 1997) as an essential aspect of understanding the biotic and abiotic processes that affect the biogeomorphology of coastal systems.

Cross-References

- [Algal Rims](#)
- [Beachrock](#)
- [Bioconstruction](#)
- [Bioengineered Shore Protection](#)
- [Bioerosion](#)
- [Bioherms and Biostromes](#)
- [Coral Reefs](#)
- [Reefs, Non-Coral](#)
- [Rock Coast Processes](#)
- [Shore Platforms](#)
- [Vegetated Coasts](#)

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Bioherms and Biostromes

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History

Originally coined by Cumings (1932), the word bioherm along with its brother term biostrome have been widely used in reef literature, but their proper stratigraphic definition is often misunderstood.

In the original meaning (Chevalier 1961) a bioherm was defined as a mound or lens-shaped organic build-up, edified by the skeletons of various organisms and lying unconformably inside a stratigraphic series of different lithology. Conversely, a biostrome was a flat, layered reef structure, wide or narrow in shape and causing no stratigraphic disturbance inside its sedimentary environment.

Discussion

Both words “bioherm” and “biostrome” were obviously coined for fossil build-ups, whose stratigraphic position in the sedimentary sequence can be studied; and they were also commonly used for the description of living or subfossil structures, whether the sedimentary environment of the latter is accessible or not to study.

Definitions vary according to authors: In the *Encyclopaedia Britannica* a bioherm is defined as an “ancient organic reef of moundlike form built by a variety of marine invertebrates . . . (and coralline algae). A structure built by similar organisms that is bedded but not moundlike is called a biostrome.”

Many geologists, however, extend these definitions to gravity deposited mounds or layers of skeletal remains, such as shells or broken coral, including reworked or transported material, as illustrated by Roger Suthren in his on-line

lectures in Sedimentology, a second year Geology module at Oxford Brookes University: “Bioherms: (are) mound or lens-shaped (biological build-ups). Some are in-place organic structures (reefs), others are banks of loose, transported carbonate sediment consisting largely of shells or skeletons. Biostromes: (are) laterally extensive beds, sheets or ribbons of carbonate material. Some have grown in-place (reefs); others consist of transported shells and skeletons.”

For Battistini et al. (1975) a bioherm is a: “lens shaped organic reef ... embedded *in situ* inside sedimentary layers of different lithological nature ... it may be surrounded by a peripheral talus of biodetrital sediments,” whereas a biostrome is a “layered, bank like organic reef of variable extension, creating no discontinuity inside the embedding sedimentary layers.”

There is, therefore, no general agreement upon a complete definition taking into account at one and the same time such different characters as: age, stratigraphic conformity or unconformity, along with the autochthonous or allochthonous nature of deposited organisms.

Furthermore, many authors (notably among biologists and geographers) tend to use “bioherm” as a general term not only for major biological build-ups such as extensive algal rims or coral reefs (e.g., see Adey and Burke 1976) but also for small-scale organic build-ups, for which the word “biostrome” would better fit. Bosence and Pedley, who had first used “bioherm” in a preliminary publication (1979) dealing with Miocene layers of calcareous algae in Malta, appropriately dropped it for “biostrome” in their final paper (1982).

It is, therefore, difficult for an actualist (whether geologist or not) to find criteria sufficiently precise and reliable to distinguish between the alternate notions of bioherm and biostrome. For example, an algal rim growing on the outer edge of a coral reef is indeed a bioherm, or a part of a bioherm since it takes an active part in the sedimentary processes of the latter, but the same kind of formation thinly coating a limestone or a volcanic shore, or on a vertical cliff, without altering sedimentation should be called a biostrome even if both formations are in continuity with one another.

Further difficulty lies in the fact that, for actualists, detrital accumulations of dead shells and broken skeletal material (generally mudsupported) are considered as something very different from a true build-up or reef, since the latter is fundamentally made of an *in situ* developed formation, resulting in boundstone or framestone lithologies *sensu* Bathurst (1971).

Conclusions

Unless bio-accumulated detrital mounds and layers are taken out of the definition of bioherms and biostromes (a revision that only geologists can decide), and the status of small-scale build-ups is settled, the use of the latter words should preferably be

restricted to the stratigraphic study of the fossil formations for which they were first coined (their associated detrital facies, and other types of detrital formations being included or not). Students of living reefs are conversely encouraged to prefer more general terms (such as “biological build-up,” “reef-like structure,” or “biogenic construction”) instead.

Cross-References

- ▶ [Algal Rims](#)
- ▶ [Coral Reefs](#)
- ▶ [Reefs, Non-Coral](#)
- ▶ [Sea-Level Indicators, Biologic](#)

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Black and Caspian Seas, Coastal Ecology and Geomorphology

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Coastal Zone of the Black Sea

The coasts of the Black Sea are rather uniform and slightly embayed. The Crimea is the only large peninsula

Pavel Kaplin: deceased.

protruding offshore. The wide opened bays facing the sea (Odesskii, Kalamitskii, Tendrovskii, Karkynitskii, Yarylgachskii, Burgasskii) as well as the above mentioned Crimean Peninsula are located in the northern part of the region. The southern, eastern, and western coasts are smooth and uniform with small bays. The total extent of the coastline exceeds 4,000 km (Fig. 1).

Zenkovich (1958, 1959) contributed much to the study of the Black Sea coasts. In the two-volume monograph, he described coasts of the former Soviet Union and analyzed dynamics and morphology of certain regions. Diverse coastal areas were described by investigators from different countries (Nevesskii 1967; Shuiskii 1974; Simeonova 1976; Kiknadze 1977; Zenkovich and Schwartz 1987; Shuiskii and Schwartz 1988; Kaplin et al. 1991, 1993). The American Society of Civil Engineers has recently published a collection of articles concerning the Black Sea coasts (Kos'yan 1993).

The environmental problems of the coasts have been discussed in many publications. The most complete summaries were given in the monographs of Sapozhnikov (1992) and Kuksa (1994).

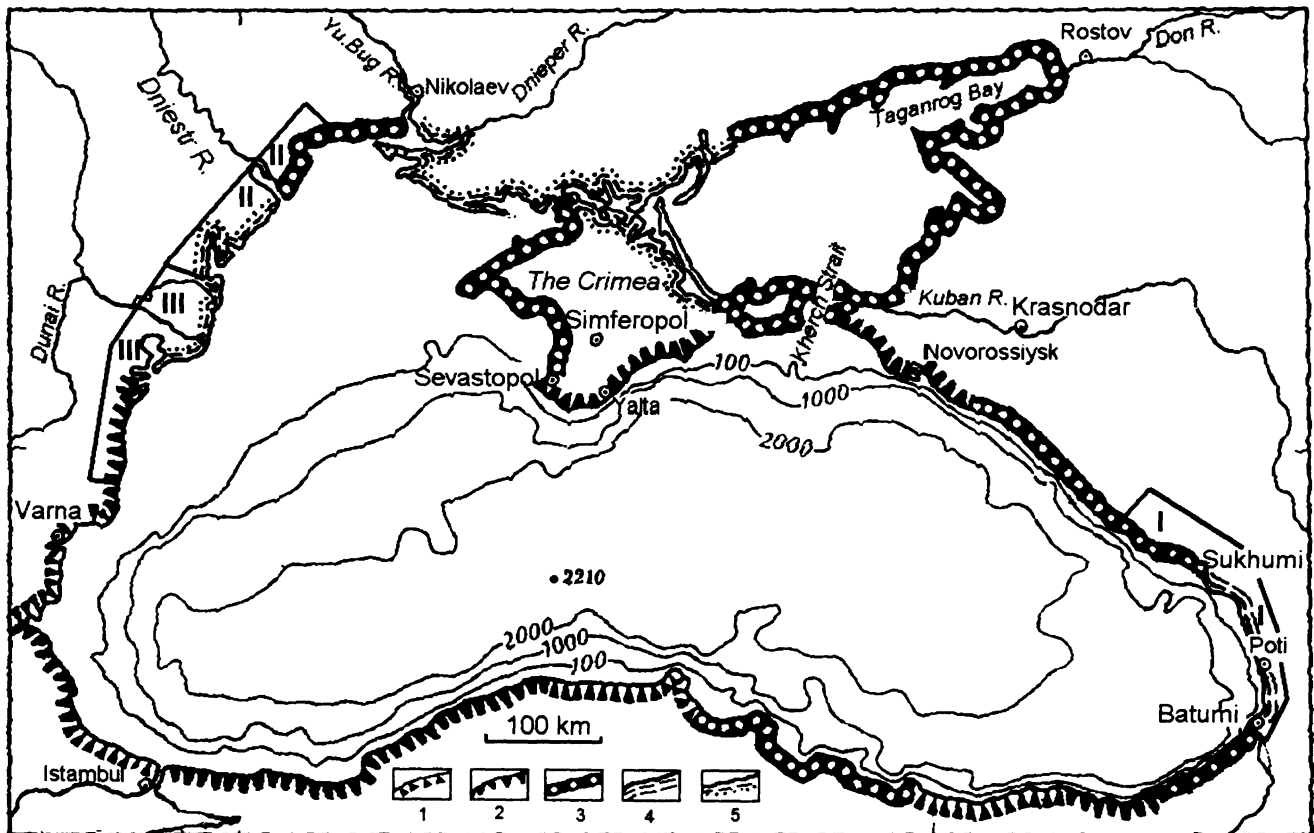
Large-scale investigations were carried out in the frame of the international INEP program "Black Sea Environmental

Program." Due to these activities about 2,000 analytical maps of the Black Sea natural environment were compiled, among them the map of the main sources of pollution in the nearshore zone with subsequent entry to the geoecological information system (Berlyant et al. 1999). The Geographic Information System (GIS) was processed at the Geographical Faculty of the Moscow State University. The users of this GIS may receive not only maps, but also the tables with the data on the amount of pollutants and other information concerning the sources of pollution and natural reserves of the Black Sea. A compact-disc "Black Sea GIS" was published by INOPS/ENVP in 1998.

Environmental Problems of the Coast

Two main problems could be outlined among the environmental problems of the coasts: (1) influence of the rising sea level upon coastal processes and intensification of erosion related to it; (2) increasing anthropogenic impact.

Anthropogenic impact is mainly manifested by water pollution. Water contamination by pesticides leading to degradation of bottom vegetation was revealed in shallow bays (Kuksa 1994). It is the result of disposal of freshwater from irrigation systems of Southern Ukraine. Water pollution



Black and Caspian Seas, Coastal Ecology and Geomorphology, Fig. 1 Types of coasts of the Black Sea and the Sea of Azov. 1, straight faulted; 2, erosional bight; 3, graded erosional and depositional; 4,

graded depositional; 5, liman and lagoon, erosional and depositional. Key study areas are shown by numbers (I–III)

caused a threefold decrease in the phytoplankton biomass in the nearshore zone and a 1.5-fold decrease in the zooplankton and zoobenthos biomass. Considerable pollution of the sea and especially its nearshore zone is determined by the influx of freshwater from the largest river of the region – the Danube. Its influence is noticed along the coasts of Ukraine, Romania, Bulgaria, and even Turkey. The Danube discharges enormous amount of oil-products, heavy metals, pesticides, and other pollutants. Pollutants are mainly accumulated in bottom sediments and biota. For instance, water plants of the Danube coast contain 0.007–0.020 mg/kg of mercury.

The concentration of pollutants discharged by the Danube decreases eastward (near Odessa and Sevastopol) and southward (in Romania and Bulgaria). Other rivers, the Dnieper, Inguri, Rioni, Chorokh and others, contribute much to the contamination of the nearshore waters.

Due to pollution of nearshore waters the role of biogenic sediments (mainly shells) in coastal dynamics decreases. At the end of the 1940s shelly sediments constituted 40–50% of coastal accumulative forms on the northwestern coast (Zenkovich 1982), while in the 1980s its contribution was less than 10% (Shuiskii 1974).

Another important ecological factor of anthropogenic origin is the influence of economic activity on the sediment budget in the coastal zone. Regulation of the rivers causes a sharp decrease in the solid river runoff and, hence, less sediments are supplied to beaches. Mass removal of sediments (sand, pebbles, gravel) directly from beaches, quarries, the nearshore zone, and river mouths considerably damaged the coastal zone. In the Caucasian coastal region this process started at the end of the last century when beach sediments were taken for construction of railroads. Mass sediment removal continued in the 1950s–1960s, when ports and other economic objects were built. During 1945–55, 100 million m³ of beach pebbles were removed from the Tuapse-Adler coast (Kiknadze 1977). As a result of this action, many beaches of the Caucasian coast became one-half smaller during two or three decades. This caused intensive coastal erosion. Of the 312-km-long Georgian coastline, 220 km were subjected to coastal erosion due to its retreat at a rate of 1–3 m/year. Active coastal erosion manifested by beach destruction was also recorded in the Crimea (Zenkovich 1982).

During the last few decades many countries have been taking efforts to protect their shores. However, many hydrotechnical constructions such as seawalls, groins, breakwaters, and others have intensified an adverse effect of the sea on the coast. Construction of artificial beaches appeared to be the most effective method. During 1981–86 in Georgia, about 8 million m³ of sediment was taken from subaerial quarries that facilitated creating artificial beaches with a total area of about 60 ha. As a result, a recreation zone was formed and the problem of shore protection in Georgia was practically solved (Kiknadze 1977; Zenkovich and Schwartz 1987). Creation of

artificial beaches or additional sediment supply to existing natural ones was undertaken in other regions as well (Odessa, Crimea, Bulgaria).

Coastal Geomorphology

In general, erosional coasts predominate along the Black Sea. Elevated mountainous coasts predominate in the eastern and southern parts of the Black Sea. This is a zone of young Alpine orogenesis. Graded and erosional accumulative coasts are typical of the western and northern parts of the sea. Geologically this zone is dominated by hard blocks protruding from the ancient Russian platform and remains of the Baikalian orogenesis. In the Eastern Black Sea erosional processes are especially active due to an extremely narrow continental shelf which sometimes nearly coincides with the coastline as in the Caucasus. Thus, the submarine slope has steep gradients allowing large storm waves to attack the coast.

Slopes of the Great Caucasian Ridge form the largest part of the Caucasian coast, since the axis of the ridge is subparallel to the coastline. This is the reason why cliffy coasts up to 200 m high prevail between Anapa and Sukhumi. The cliffs are cut in the steeply sloping flysch beds and its ridges are noticed in the submarine bench. In the southern part of the Caucasian coast, the Batumi region, foothills of the Little Caucasian Ridge reach the shore. The Colchis Lowland lies between the Great and Little Caucasian Ridges. It follows the large Alpine flexure. The lowland is swamped and its flanks are only slightly higher than the sea level. The lowland experiences a prolonged tectonic submergence. Many rivers flowing from the slopes of both ridges drain onto the Colchis Lowland. Despite this, sandy coasts do not migrate seaward. The heads of submarine canyons are located close to the mouths of the large rivers such as the Inguri, Rioni, Supsa, and others. The alluvial material is removed to the canyons instead of being accumulated on the beaches. Moreover, in many places the shores of the Colchis Lowland are eroded (up to 3 m/year).

The presence of large promontories near Adler, Pitsunda, Sukhumi, Burup-Talii are typical of the Caucasian coast. They are located near the river mouths and consist of the Holocene alluvium (Fig. 2). These promontories protrude far offshore and overlie a significant portion of the continental shelf. No other large accumulative landforms are present on the Caucasian coast. The beaches are associated either with numerous river mouths or places of longshore sediment drift discharge. They are mainly composed of gravel and pebbles.

As shown above, accumulative forms are subjected to active erosion. Its intensification is caused by natural reasons: sea-level rise at a rate of 1–2 mm/year and decrease in river discharge due to regulation of rivers and removal of beach sediments. Heads of submarine canyons contribute to coastal dynamics since part of the material transported by alongshore drift is accumulated there. For example, the Akula submarine

canyon near Pitsunda accumulates about 80 thousand m^3 of sediments per year (Kiknadze 1977; Kos'yan 1993).

Within the Georgian coastal zone alongshore drift is directed to the southeast (Fig. 2). Each sediment stream represents a dynamic system with its own source of sediment supply and areas of sediment loss (submarine canyons and steep slopes) or final discharge. The capacity of alongshore sediment streams ranges from 3–15 to 150–220 thousand m^3/year . A small alongshore sediment stream is directed to the north from the Chorokh river mouth to the Colchis Lowland.

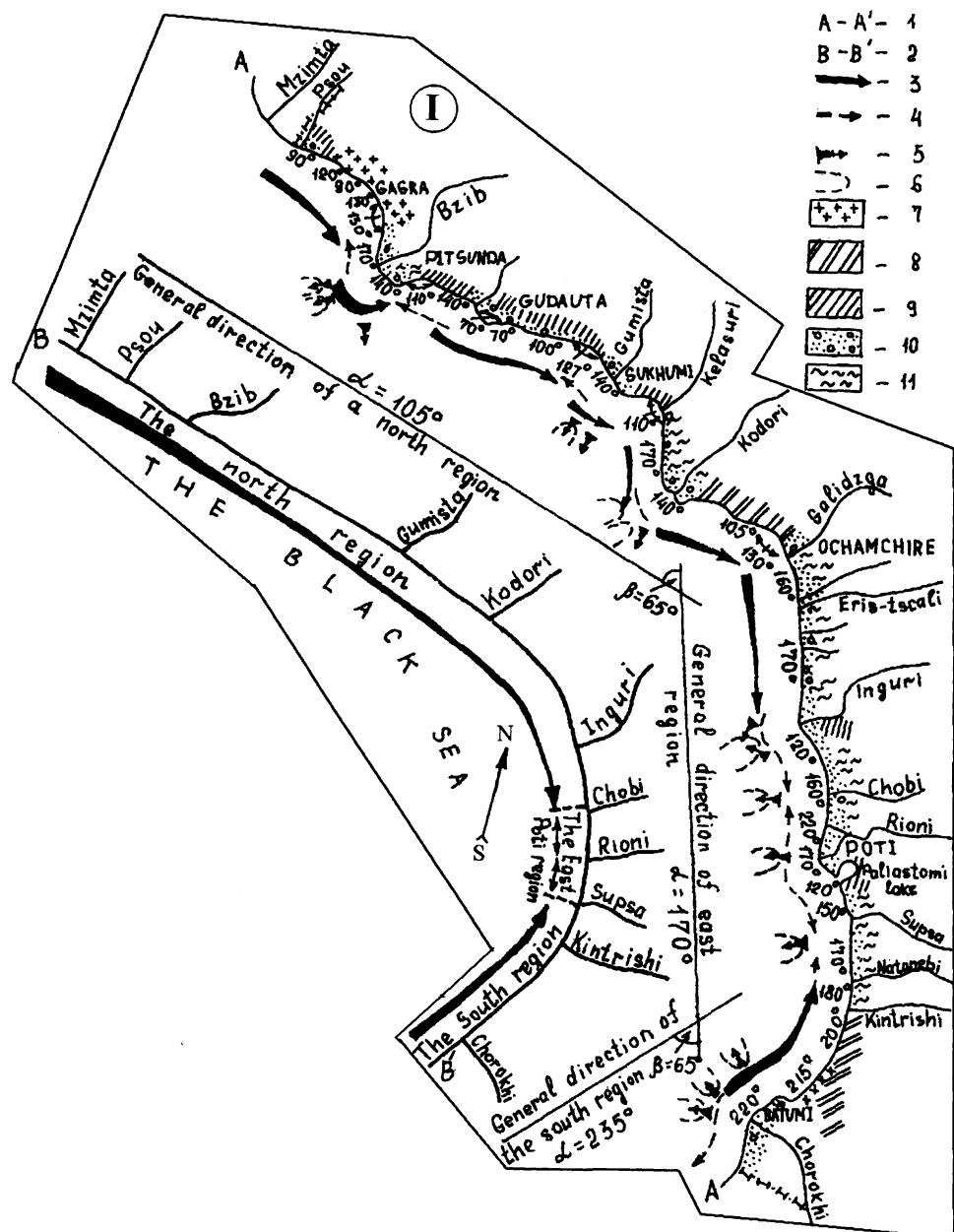
High erosional shores are typical of the mountainous coasts of the Crimea. They are subjected to active erosion since they are affected by severe winds (and waves) blowing

from the southwest and southeast. Shore destruction is accelerated by landslides occurring in clays. Sometimes the landslides have an area of hundreds of square meters. For instance, the town of Alupka is located on six large landslides and its stability is conditioned by several factors. Of these are influence of underground and surface waters, abrasion, load of buildings and other construction.

Many shores of the Southern Crimea are formed by the slopes of ancient volcanoes (Karadag region) and tectonic faults. Outcrops of volcanic rocks and limestones form capes separated by shores represented by soft shales, clays, and sandstones. Ria-coasts occur near Sevastopol and Balaklava.

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Fig. 2 Schematic map of morpho- and litho-dynamics of the Black Sea coast in Georgia. (After Kos'yan 1993). 1, modern coastline; 2, coastline during the period of the drop in sea-level rise 6–5 ka; 3, longshore sediment streams, their direction and relative actual capacity; 4, direction of migration and transport of finer sediment; 5, partial loss of sediments at considerable depths; 6, canyon heads and steep falls; 7, cliffed rocks with erosional relief; 8, semi-cliffed rocks (conglomerates, marl, schists, etc.) with erosional relief; 9, related rocks (marine and lagoonal clays) with plain relief; 10, loose deposits (pebbles, gravel, sands of terraces, dunes and beaches); 11, bog and lacustrine deposits



Beaches of the Southern Crimea are formed of pebbles, because finer sediments (more than 0.03) are transported down the steep submarine slopes. Removal of pebbles for building purposes caused the disappearance of beaches. However, some of them have been recently restored.

Many shores of the Southern Crimea are artificially protected. Dynamic interaction between different regions is weak due to the absence of large rivers supplying sufficient amounts of alluvial sediments to the coastal zone. Thus, local shore protection is successful and has no negative influence on adjacent coasts. Different, usually complex, engineering structures are used that protect coasts from both landslides and abrasion. Of these are embankments with seawalls, traverses, breakwaters, groins with artificial sediment filling between them, etc. (Zenkovich 1982; Kos'yan 1993).

Steep coasts of the Southern Black Sea are formed by densely forested northern slopes of the high Eastern and Western Pontus mountains stretching subparallel to the coastline. The mountains gradually lower westward and near the Bosphorus Strait their height does not exceed 300 m. Erosional and denudation coasts with steep rocky cliffs are widespread in Turkey. Only in separated small bays do the sandy-pebbly "pocket" beaches occur. Areas of sediment accumulation are associated with mouths of such large rivers as the Kizil-Irmak, Sakarja, and Eshil-Irmak. These rivers form rather large deltas prograding far offshore and nearly reaching the edge of the narrow continental shelf. Violent storms produced by severe northwesterly winds deflect the pathways of alluvium to the east thus forming flanked barriers (Kos'yan 1993).

The largest curves of the Turkish coast correspond to the lowland peninsulas of Bafri and Djiva, related to river deltas and the mountainous Injeburun Peninsula.

Western and northwestern coasts of the Black Sea are rather low with hilly plains of different origin (alluvial, marine, and alluvial-marine) facing the sea. The delta of the Danube, the largest river of Western Europe, is located here. It has a complicated structure. Besides common channel bars there are a series of cheniers (local name "grindu") marking the stages of delta progradation. The river mainly discharges through its northern Kiliiskii channel. Thus, the southern part of the delta is smaller and is being slightly eroded. Active utilization of the Danube water for irrigation by five countries reinforces erosion of the southern part of the delta. There are numerous water reservoirs in the delta: lakes-limans (northern part), complexes of lakes and lagoons (southern part), lakes (inner part). From the north the delta is bounded by the Budzhak plateau, and from the southwest by the lake-lagoon complex of Rozelm-Synop. Abundance of warm water and high fertility of soils favor plant and animal life.

Coasts to the northeast from the delta are represented by plains and low plateaus. The only exception is the anticline of the Tarkhankut Peninsula. Its steep slopes are mainly

composed of easily eroded loesses and clays. The rate of erosion ranges from 7 to 20 m/year (Shuiskii 1974).

These erosional coasts alternate with lagoons and limans. Limans represent the lower parts of river valleys that have been flooded during the Holocene transgression. Most of them are separated from the sea by sandy-shelly accumulative forms (spits or baymouth barriers). Specific environmental conditions exist in limans since their waters are warmer and less salty. As a result, productivity of waters is higher.

A considerable part of the coast is subjected to landslides. Both landslides and coastal abrasion destroy valuable territories of the Ukrainian steppes. At present, the accumulative forms in the mouths of limans are eroding. They are composed of sand layers overlying lagoonal clays, thus giving evidence for migration of the accumulative forms toward the lagoons (Shuiskii and Schwartz 1988).

Two opposite longshore drifts exist in the region stretching to the northeast from the Danube River to Odessa (Fig. 3).

Jagged coasts are characteristic of the region to the east from Odessa including the western Crimea. Adjacent lowlands experience relative submergence at a maximal rate of 30 cm/100 years as recorded in the inner part of the Karkinit Bay. Large accumulative forms are the most interesting elements of the coastal relief, that is, the Kinburn spit and the system of the Tendra-Dyarylgach spits related to it.

The Kinburn spit and Odessa shoal (to the west of it) originated in the place of the Dnieper and South Bug deltas junction. Under the Holocene sea-level rise deltaic sediments were reworked and a system of subaerial and submarine sand bars was formed. Dunes and salt lakes located on the Kinburn spit are parallel to the coastline.

The Tendra and Yarylgach spits represent a joined accumulative form that continues to grow. However, landward migration also takes place. As a result, the central part of the accumulative form became attached to the continental shore, and its distal end formed two separate spits.

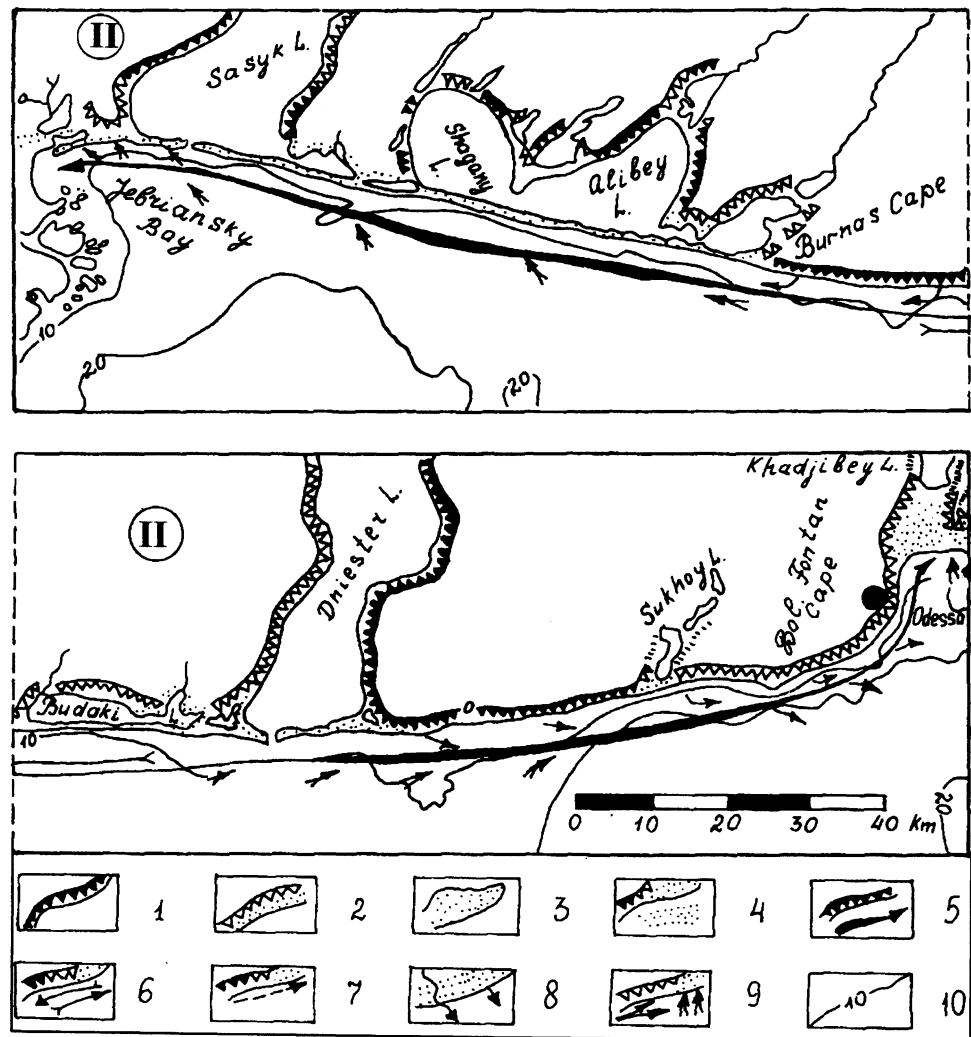
The eastern part of the Crimean Peninsula, together with the Taman Peninsula, form a single coastal region divided into two parts by the Kerch' Strait. The territory is covered by limans and lagoons associated with the ancient and modern delta of the Kuban' River. However, erosional shores are dominant. The coastline represents a series of arcs where clays of Maikopian age intercalate with solid rocks of Neogene age that form headlands.

Along the Kerch'-Taman' coast relics of the ancient accumulative forms and lagoonal silts were reported that allowed for reconstructing the Holocene history of the coastal area.

The western coast of the Black Sea lies in Romania and Bulgaria. The Romanian coast is subdivided into two parts. Its western part corresponds to the Danube delta that equals 78% of the delta surface. As mentioned above this part of the delta is now eroding at the rate of up to 7 m/year (Kos'yan 1993).

Black and Caspian Seas, Coastal Ecology and Geomorphology, Fig. 3

Geomorphological map of the northwestern Black Sea coast between the Danube River delta and Odessa. (After Zenkovich 1958). 1, active erosional scarps; 2, passive erosional scarps; 3, emerged coastal accretion bodies and coastal ridges; 4, emerged coastal accretion bodies and coastal ridges; 5, longshore sediment streams (thickness of arrows proportional to the capacity of a stream); 6, longshore sediment drift; 7, prevailing sediment drift; 8, offshore sediment drift; 9, onshore sediment drift; 10, depths in meters



Southward from the delta the coast is graded. It is composed of loesses, clays, and limestones of Neogene and Pleistocene age. A considerable part of the southern Romanian coast (51 km of the total 101 km) is abraded and consists of active cliffs 2–40 m high. The rate of abrasion averages 1–2 m/year, sometimes reaching 7 m/year. Maximal rates are characteristic for cliffs composed of loesses and clays. Capes are usually formed of Neogene limestones. Submarine benches are typical of the Romanian erosional coasts. Their width sometimes reach 1,100 m (Fig. 4).

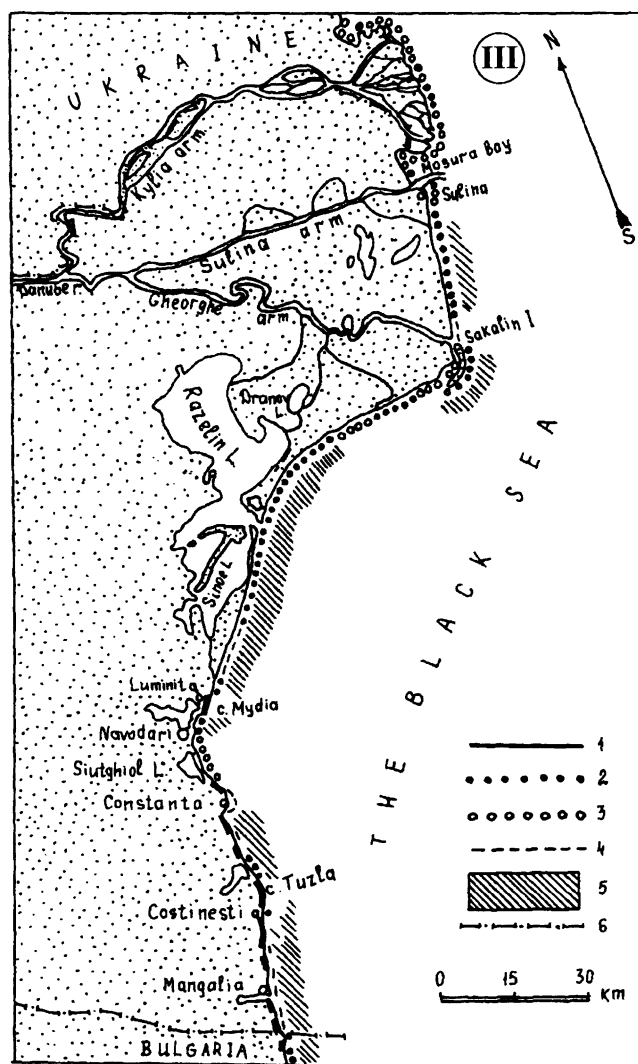
Coastal accumulative forms are represented by sandy beaches resting against cliffs and, near the river mouths, and by barrier forms separating lakes and lagoons. Sintghiol is the largest sandy barrier. Sedimentary material is supplied to the sea by the Danube delta and active cliffs.

The Romanian coast is actively used for recreation. To protect the coast from destruction certain efforts have been taken: construction of seawalls, cobble filling, etc.

The coasts of Bulgaria are mainly erosional. In southern Bulgaria erosional forms are restricted to the small bays of the

zone of Alpine orogenesis. Cliffs of eight different types are distinguished in this area: from 15 to 20 m high cliffs with even surfaces composed of uniform loess and clayey deposits to high cliffs (up to 60–90 m) with uneven surfaces and a series of landslide steps on the slopes. Such cliffs are widespread in the region between Kavarna and Balchik and Kranevo-Zlaty Pyaski. Earthquakes facilitate landslides thus considerably accelerating retreat of the coast. The average rates of erosion vary from 0.005 to 1 m/year. Maximal rates reach 30 m/year. The estimated amount of material released due to abrasion of cliffs is 1,344,100 m³/year.

Material produced by coastal abrasion and alluvium forms the beaches that occupy 28% of the Bulgarian coast (Simeonova 1976). Some of the beaches, like that at Varna, prograde at a rate of 0.75 m/year. Similar process operate near the mouth of the Kamchiya River and in the region of the popular resorts of Albena and Zlaty Pyaski. However, most of the accumulative coasts retreat. For instance, between Cherny Nos Peninsula and the Albena resort the rate of retreat is 0.12–0.63 m/year. Generally the rate of retreat grows in the



Black and Caspian Seas, Coastal Ecology and Geomorphology, Fig. 4 Morphology of the Romanian coast of the Black Sea. (After Kos'yan 1993). 1, active cliffs; 2, retreating coastline of accumulative coasts; 3, prograding coastline of accumulative coasts; 4, stable coastlines; 5, coastal sections with predominance of erosional processes in nearshore zone; 6, state frontier

northward direction. Erosion often results from the negative influence of human activities and underestimation of the role of coastal processes. To protect coasts from erosion Bulgarian engineers fill up tetrapods with stones and construct dams separating the bays from the sea and straighten the coastline.

The coasts are also protected by groins (often short and without any filling between them), seawalls and other less effective structures. The most effective means, such as creation of artificial beaches, are not implemented in Bulgaria.

Sea-level oscillations played an important role in the recent evolution of the Black Sea coasts. From the available data it follows that during the twentieth century sea level was steadily rising at the average rate of 2.1 ± 1.3 mm/year (Nikonov et al. 1997). This estimation is based on the data

collected at 70 points on the Russian, Ukrainian, Georgian, and Bulgarian coasts. The values exceed the average rate of the global sea-level rise, probably due to tectonic submergence of the Black Sea depression. Different estimations of the sea-level oscillations could be definitely attributed to different tectonic movements. The highest rates of sea-level rise were recorded in the Colchis Lowland, while the lowest ones were in the northeastern Black Sea.

Thus, it might be concluded that submergence of the coasts has been the main trend in their recent evolution. This process leads to abrasion of the cliffed coasts and erosion of the accumulative ones. However, under predicted conditions of more rapid sea-level rise, erosion will be intensified and many of the unique accumulative forms will be destroyed. First of all this affects the accumulative forms of the limans in the northwestern coastal area (Tendra, Binburn, Yagyrlach spits). Their destruction may have a severe impact on the ecology of limans that are the zones of extremely high bioproductivity. Sea-level rise will accelerate destruction of the Holocene accumulative forms on the Georgian and Turkish coasts. In this connection, all countries of the region must plan enhancing protective activities and carry out a longterm policy of coastal management.

Coastal Zone of the Caspian Sea

The coasts of the Caspian Sea display a great variety of natural environments being located in different landscape zones. Recently, the problems associated with the rapid rise of its sea level have generated particular interest, especially in the context of the expected accelerated rise of the global sea level (Dolotov and Kaplin 1996).

The history of investigations on the Caspian Sea coasts was discussed in detail in the monograph of Leont'ev and Khalilov (1965). Further generalization was given in the monograph written by Leont'ev with co-authors (1977). Present environmental problems of this area were outlined in the monographs of Kuksa (1994) and Zonn (1999).

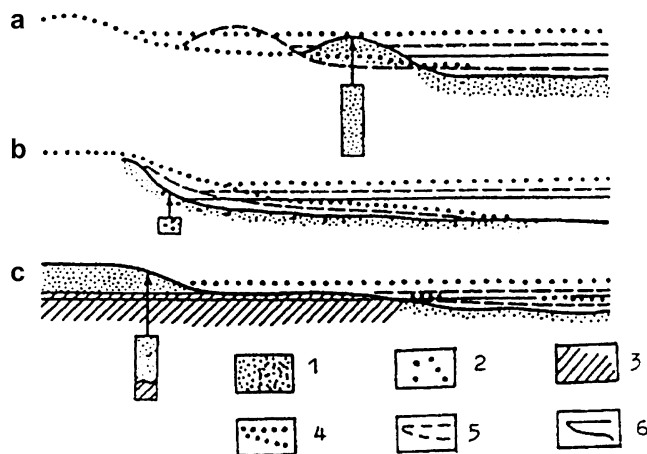
Environmental problems of the Caspian Sea coastal zone arise from active economic development of not only the Caspian Sea itself, but its drainage area and adjacent territories (Kaplin and Ignatov 1997). This region is distinguished by repeated sea-level changes, both seasonal and multi-annual. That is why the Caspian Sea is a natural laboratory for studying evolution of coasts under different sea-level oscillations.

In the modern historical period a rapid sea-level fall (of nearly 1.7 m) occurred from 1929 until the early 1940s. Dynamic changes in the coastal zone mainly depended upon the rate of sea-level fall (or decrease in depth in the nearshore zone) and the amount of sediments in the coastal area (Dolotov 1961).

On the shallow-water sand coasts that are typical of the Caspian Sea the relief-forming processes are controlled by sediment budget, gradient of the coast, and configuration of the coastline under sea-level fall going on at different rates. Three types of the coasts with different patterns of relief changes have been identified.

Continuous accumulation of sediments and progradation of coasts (Fig. 5a) takes place in case sufficient amounts of sediments are supplied to the coastal zone from adjacent land and shores (positive sediment budget over prolonged time period). If a positive sediment budget is replaced by a negative one progradation of the coast changes to landward retreat of the coastline (Fig. 5b) (irregular sediment supply). Under insufficient sediment supply erosion is replaced by preservation of sandy accumulative bodies that occurs when direct wave attacks over the former coastline have ceased (Fig. 5c).

Under sea-level fall the evolution of the Caspian Sea coasts went on in the following manner: continuous accumulation of sediments, emergence of the seafloor, and continuous seaward advance of the coastline. The area of coasts enlarged, and economic activity occupied new territories where settlements, roads, oil- and gas-pipelines, and resorts were constructed (Dolotov 1996). Taking into consideration the predicted future sea-level fall the Soviet Government decided to create a vast recreation zone in the coastal regions of Dagestan and Azerbaidzhan (Molchanova 1989). The Caspian coast offers several advantages over the Caucasian coast of the Black Sea. Sandy beaches are wider and longer here, solar radiation is more active and the number of sunny days in summer is higher (Veliev et al. 1987). Part of these constructions has already been built.



Black and Caspian Seas, Coastal Ecology and Geomorphology, Fig. 5 Processes of relief formation and sediment accumulation in shallow-water nearshore zone under sea-level fall. (After Dolotov 1996). (a) continuous sediment accumulation; (b) erosion of external edges of accumulative coasts; (c) preservation of sandy accumulative body; 1, sand; 2, pebble; 3, bedrock; 4, 5 and 6, successive positions of sea level

In 1978, an unexpected and sharp sea-level rise occurred. The average rate of sea-level rise was 14–15 cm/year, but in some years it was as high as 30 cm/year and even more. The direct influence of the sea-level rise was flooding of coastal lowlands and acceleration of coastal erosion. The indirect impact included the rise of groundwater, swamping of the coastal area, salinization of soils, and expansion of surge areas. Environmental conditions of both the nearshore shallow zone and adjacent land sharply changed. Sea-level rise favored erosional processes and general landward migration of the coastline.

At the same time, contrary to the general trend of sea-level rise, in some patches of the western coast accumulation of sediments went on and the coastline migrated seaward. This happens, when sediment input exceeds the amount of unconsolidated sediments flooded by the advancing sea. These coastal regions are of special interest since they give a unique opportunity for future economic development (primarily recreation) even under the ongoing sea-level rise.

Generally, the character of relief changes under sea-level rise depends upon the rate of this rise, relief-forming environmental processes (hydrodynamics), and sediment balance. Coastal morphology determines the character and rate of the natural catastrophic processes together with their impact on economy and population (Dolotov 1996).

Coastal Geomorphology

Based on differences in relief, the Caspian Sea coast is subdivided into four regions (Leont'ev et al. 1977).

The western coast receives about 50% of the total solid river runoff, while the material produced by abrasion is considerably less abundant. As a result of a positive sediment balance coastal erosion is suppressed. Under dominant north-westerly and southeasterly winds longshore drift streams are formed that smooth the coast and create accumulative forms.

River runoff is practically absent in the eastern regions. Active erosion in the recent past has not produced sufficient amounts of sedimentary material, since the cliffs are mainly composed of clays and limestones. Biogenic and chemogenic sedimentation went on slowly. As a whole, the coast is more embayed than in the west. In places where the coastline remains relatively straight over long distances, prevailing waves that are transversal to the coast form large accumulative barriers.

The northern coast is distinguished by extreme shallowness. Southeasterly and northerly winds predominate here. Abundant finegrained alluvial material supplied by rivers is transported in the form of suspension. Winds and on- and offshore currents form the coasts characterized by frequent and significant displacement of the coastline.

Southern coasts are represented (Voropaev et al. 1998) by coastal lowlands ranging in width from 1 (in the central part) to 60 km (near the large deltas of the Sefidrud and Gyurgyan

Rivers). More than 40 small rivers discharge into the Caspian Sea in this region. During the past 50 years, solid river runoff was considerably reduced due to construction of reservoirs on all the large rivers. Granulometric composition of the beaches changes from gravel–pebble (western Mazenderan) to sand (Gilidzhan, central Mazenderan), and silt (eastern Mazenderan).

Several types of accumulative coasts occur in the Caspian Sea: lagoonal shores, coasts with terraces and other accumulative forms, erosional and deltaic shores, mudflats, coastal lowlands formed by onshore winds and waves (Fig. 6).

Deltaic coasts occupy considerable parts of the coast – these are the deltas of the Volga and the Ural in the north, Terek and Sulak in the northwest, and Kura in the southwest. Even, quite recently, these rivers discharged considerable amounts of sediments into the sea thus supporting progradation of deltas. This process intensified during sea-level falls. In the northern Caspian Sea, where the Volga and Ural deltas are located, coasts have retreated seaward by dozens and hundreds of kilometers since 1929. In the late 1950s and 1960s large-scale hydroengineering projects were launched. This caused dramatic decrease in river discharge of such rivers as the Kura, Terek, and Sulak. The rise of the Caspian Sea level caused erosion of deltas. Coastal lowlands of the Northern Caspian region became inundated, and the previously accumulated coastal landforms were destroyed.

Large portions of the Caspian Sea coasts are flat. They were formed in course of regression that took place from the 1930s to the 1970s. Mudflats occur in the northern Caspian region, around Kirov Bay in southern Azerbaidzhan, in the region surrounding Krasnovodsk Bay and to the south from it. During regression aggradation of the northwestern coasts of the sea proceeded at a rate of 60–100 m/year, in the Kizlyar Bay it was faster and reached 150–200 m/year, and northward of it even 700–800 m/year. In the Kirov Bay wind-induced mudflats were up to 1.5 km wide (Kaplin 1997). The sea-level fall resulted in considerable advance of land in the eastern coastal regions. For instance, southward from the Cheleken Peninsula the average rate of land advance during the period from 1929 to 1957 was 34–36 m/year (Leont'ev et al. 1977). Naturally, shallow bays of the northern coast were dried up.

The change from regressive to transgressive regime has dramatically affected the drained shores with mudflats. Gentle gradients (close to 0.0001) of submarine slopes caused passive flooding of the coasts.

Sea-level rise resulted not only in submergence of land, but rise of the groundwater table and, hence, salinization of groundwater and swamping of the adjacent lowlands. The Kirov Bay was filled with water again. In the vicinity of the Kilyazinskaya spit (northern Azerbaidzhan), a flat coastal

terrace that formed in 1940 due to sea-level fall has been partly flooded and swamped. Considerable parts of coastal lowlands in the northern near Caspian region on both sides of the Volga River have been flooded. Accumulation of sediments went on in such bays as the Komsomolets, Kaidak, Mertvyi Kultuk, Bol'shoi Sor in the Bugaz Peninsula, and in depressions between Baer knolls that are widespread in the northern and northeastern Caspian lowlands. Wind-induced surges up to 1.5–2 m high caused passive flooding of this vast area. These phenomena resulted from a longshore drift and caused episodic abrasion of the coasts and erosion of the seafloor, producing furrows and ridge-and-runnel erosional forms. Wind-induced currents resuspended sands and silts previously accumulated on the seafloor. Suspended material was removed by currents and precipitates near the shoreline to form saltings (mudflats).

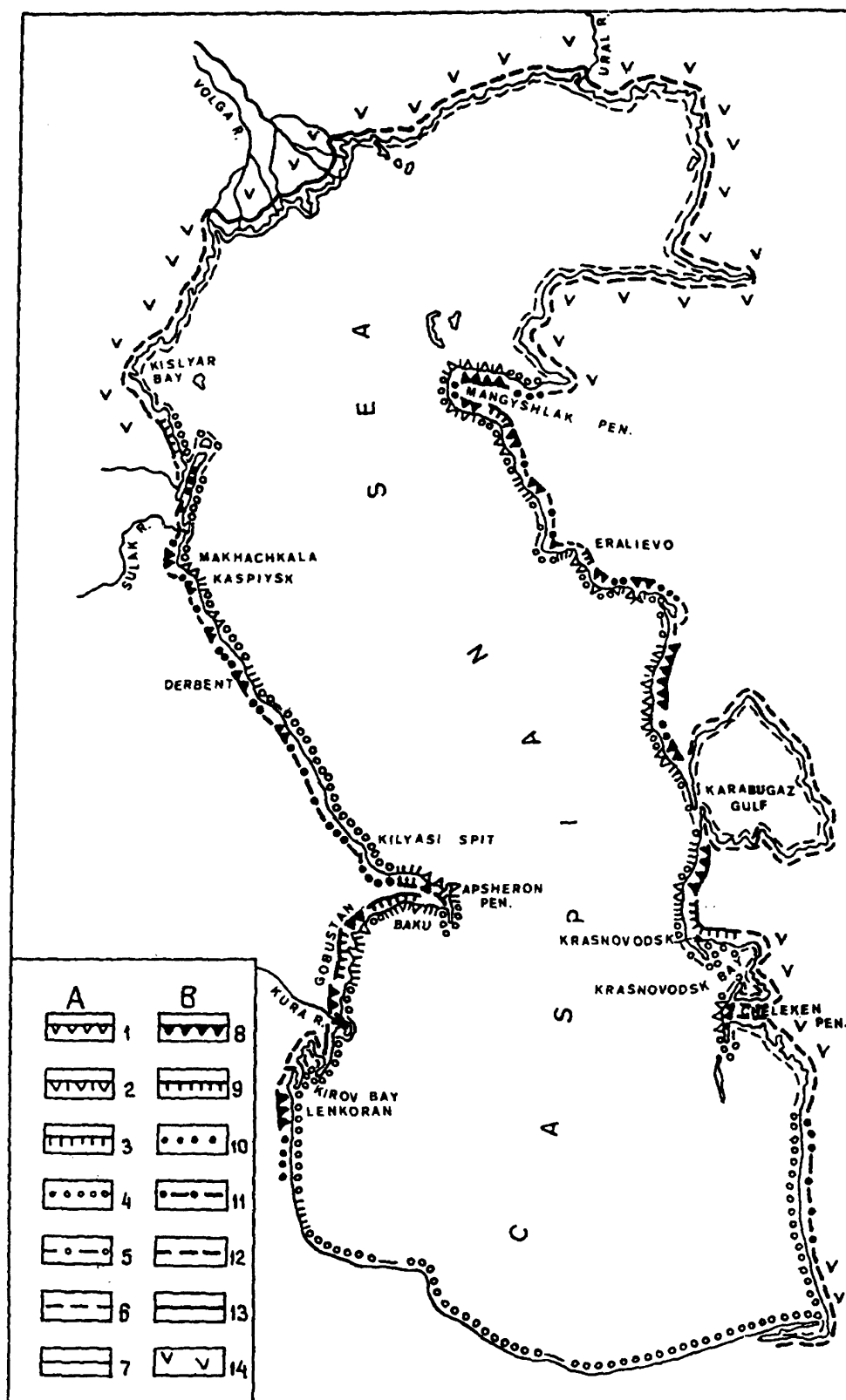
Sea-level rise changes accumulative coasts as well. As mentioned above, accumulation that took place during the regressive stage affected nearly the entire coastline, but the pattern was somewhat different. As sea level was falling, cliffs became passive, longshore drift ceased, and spits were eroded, especially at their base. We can exemplify the Kura and Agrakhan spits on the western shore, the accumulative form of Cape Rakushechnyi, Kenderli, and Krasnovodsk spits on the eastern one. With the drop in sea level, the middle and lower parts of submarine slope became eroded. The produced clastic material is gradually transported upward the slope and accumulated near the shore. Intensive landward transportation of sediments inhibited the longshore drift.

Due to the sea-level rise former cliffs became active again and abrasion intensified. Transgression changed the profile of submarine slopes, especially their upper parts. These changes are accompanied by erosion of the frontal part of accumulative forms or creation of bars near the water edge that later turn into beach barriers separating the sea from lagoons. Many lagoonal shores were formed in Dagestan and northern Azerbaidzhan, where gradients of the submarine slope equal 0.005. The present sea-level rise is responsible for accumulation of beach barriers separating lagoons. The beach barriers are overlapping lagoons thus giving an impression that the coastline retreats. However, lagoons keep expanding despite the landward migration of beach barriers since flooding of drained territories and the rise of the groundwater table favor further expansion and deepening of lagoons.

The coasts of Dagestan and Northern Azerbaidzhan, with steeper gradients (up to 0.01), have shore-attached bars formed of coquina. These bars have an asymmetrical profile giving evidence of their landward migration towards young terraces behind them. No lagoons are formed because the coasts are steep and lie above sea level.

Black and Caspian Seas, Coastal Ecology and Geomorphology, Fig. 6

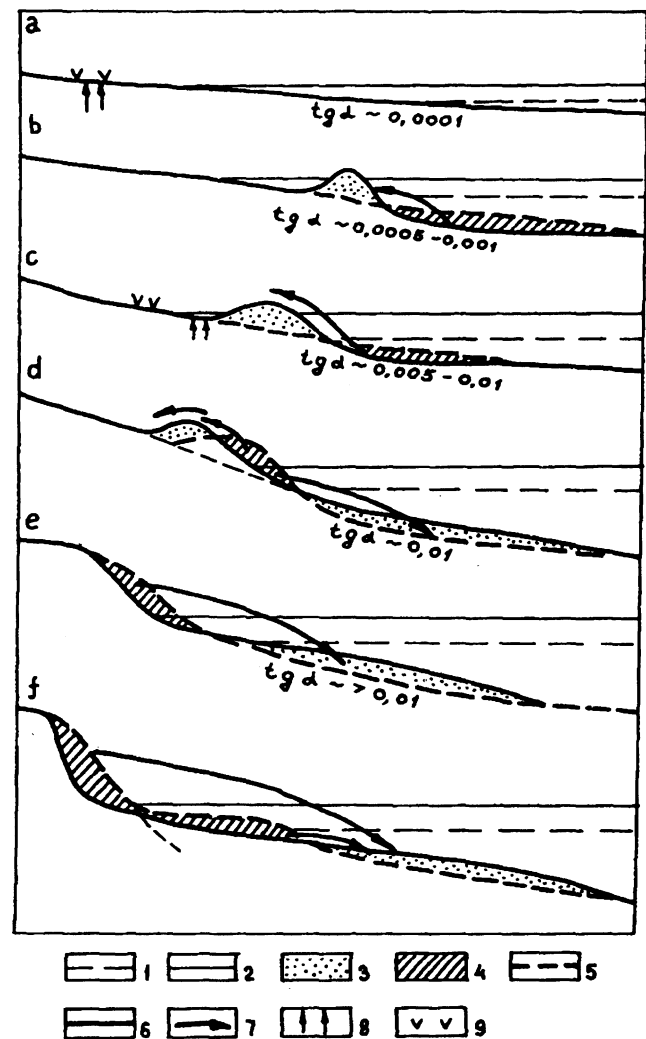
Types of the Caspian Sea coasts. (After Ignatov et al. 1993). A, regressive stage (prior to 1977). B, transgressive stage (after 1978). 1, erosional shore; 2, erosional shore with passive cliff; 3, erosional-accumulative shore; 4, prograding beach; 5, accumulative lagoonal shore; 6, mudflats formed by wind-induced surges; 7, deltaic coast. B, transgressive stage (after 1978). 8, erosional shore; 9, erosional-accumulative shore; 10, prograding beach; 11, accumulative lagoonal shore; 12, mud flats; 13, deltaic coast; 14, areas affected by transgressive flooding



Finally, slopes with gradients exceeding 0.01 are subjected to active erosion of both Holocene and recent accumulative forms. This trend leads to significant landward retreat of the coastline (Kaplin 1997). Evolution of such coasts follows the well known "Bruun's rule" (Bruun 1962).

The transgression has also affected erosional coasts. The latter are typical of the Eastern (Mangyshlak Peninsula, regions northward from the Kara-Bogaz-Gol Bay, Cheleken Peninsula) and, partially, Western (Dagestan, Lenkoran', Apsheron Peninsula) Caspian Sea. The share of erosional shores on the Dagestan coast increased from 10% to 40%, on the Azerbaidzhan coast – from 20% to 55%, Kazakhstan – from 8% to 13%, and Turkmenistan – from 7% to 22%. Until recently, some places on the Dagestan coast have been protected from erosion by offshore submarine ridges composed of limestone. Benches and ridges have been flooded by the transgression, during storms (especially surges) waves reach the cliffs. If the cliffs used to be protected by accumulative terraces, clastic material has been actively reworked; part of it being transported into the longshore drift and the rest removed down the submarine slope. In general, the above-water parts of the slopes were more actively abraded by sea waves, and the coastline rapidly migrated landward. In some parts of the Dagestan and Lenkoran', the rate of the process reached 20–25 m/year.

Therefore, the recent rise of the Caspian Sea level has significantly modified the dynamics of all identified coastal types. Evolution of the coasts subjected to relative submergence depends upon the gradient of the submarine slope (Kaplin 1997). Different ways in evolution of accumulative coasts under the sea-level rise (Fig. 7) have been discussed in several publications on the Caspian Sea (Kaplin 1989, 1990; Ignatov et al. 1992, 1993). The coastal dynamics of the Caspian Sea display a certain discrepancy between transgressive and regressive regimes on the one hand, and cycles of the coastal evolution on the other hand. A regressive regime usually corresponds to the cycle of accumulation. However, erosion of coastal accumulative forms started in the 1960s when sea level was still falling. Certainly, one of the reasons why erosion became active is economic activity, such as construction of barrages and irrigation networks that reduced the solid river runoff and led to the deficiency of sediments in the coastal zone. On the other hand, it is natural that an erosional cycle succeeds the accumulative one. The reason is that both a drop in the sea level and its rise under transgression are associated with reformation of the submarine slope and with its erosional cutback. Yet, during regression the zone of the shore slope erosion shifts seaward, not landward, involving parts of the outer slope where benthonic material is of a finer grain size. With time the deposits of fractions that can be transported up the slope and that can build the accretion forms are depleted. Although the rates of the Caspian coastal processes are much higher than of those in the world ocean,



Black and Caspian Seas, Coastal Ecology and Geomorphology, Fig. 7 Schematic representation of the Caspian Sea transgressive coasts in function of the offshore gradient. (After Kaplin 1989). 1, regressive sea level; 2, transgressive sea level; 3, sediment accretions; 4, erosion lens; 5, former profile of coastal zone; 6, present-day profile; 7, sediment drift; 8, groundwater rises; 9, bogs

these processes are essentially similar. Therefore, research into the Caspian Sea coastal dynamics has importance beyond its regional significance. It may be of great use for simulating the formative laws applying to the coastal dynamics of the world ocean, particularly as the ocean level is rising and is likely to keep rising in the future as a result of global warming (Kaplin 1997).

Environmental Problems of the Coasts

The present environmental conditions of the Caspian Sea are determined by the effect of rising sea level and increasing anthropogenic impact (Kaplin 1997).

Considerable sea-level rise has already adversely affected the economy of the coastal states of the Caspian region including the Russian Federation that occupies its

northwestern part – the Dagestan, Kalmykia, and Astrakhan' region. Many industrial and habitable buildings, recreational structures and other objects are now in the zone of flooding or in the zone of subsoil water penetration (Dolotov 1996).

The rising sea level poses threats as follows: flooding and underflooding of coastal areas earlier occupied by communication facilities, livestock farms, grazing lands, fish hatcheries, piers, fish spawning grounds, wildlife and nesting bird habitats. It also prevents cattle from accessing fertile pastures. Incursions of seawaters into areas occupied by human settlements or farms generally lacking treatment facilities resulted in capture and retransportation of technogenic products, municipal effluents and wastes toward the Caspian Sea increasing the supply of pollutants. Flooding causes malfunctioning of irrigation channels and transverse drains in the irrigated areas. Intensive shore erosion has caused losses of considerable land areas.

Special environmental problem is the underflooding of running and suspended wells in the oil and gas fields and subsequent propagation of oil products. Since oil production has or had been carried out over many decades it is evident that significant amounts of oil have been accumulating in the soil, finding their way into, and heavily polluting the sea. The oil film formed over large areas, dramatically reducing water evaporation and affecting the Caspian water balance (Kaplin 1997).

Starting up of the first turn of the Astrakhan' gas-condensate complex together with exploitation of oil and gas-deposits present a potential threat to existence of unique ecosystem of the river mouths in the Northern Caspian region (Kuksa 1994). Development of shelf oil and gas fields by the Caspian states strongly affects natural processes by pollution of water and bottom sediments, destruction of plant cover, and considerable reduction of fish resources (Kurbatova 1994).

In course of the Caspian sea-level rise some of the productive oil and gas fields on the coast will be flooded. The flooded area will also include prospective sites for deep exploratory well-drilling.

The major sources of pollution of the Caspian Sea water consist of: river (surface) runoff; untreated effluents discharged from enterprises, farms, or human settlements in coastal areas or in the river mouths; navigation accidents and technical processes in industries operating directly in the Caspian water area; surface washout during surges; and flooding of producing oil and gas fields, industrial sites, agricultural lands, and human settlements.

River runoff is the largest source of pollutants to the Caspian Sea accounting for 90% or even more of the total influx of pollutant. It is attributed to the fact that the Volga, Ural, Kura, and Terek rivers receive polluted effluents from various industrial facilities and farms along their entire courses. Concentration of various pollutants in river water

at river mouths exceeds the maximum permissible levels, frequently by a very wide margin (up to tenfold or more).

It should be noted that pollutants transport with river runoff is a rather constant process, varying only slightly in different years. The most important consequences are as follows: large amounts of pollutants carried to the sea by rivers under the influence of hydrometeorological factors (wind, currents, waves) penetrate the entire water column and bottom sediments, that later act as the sources of secondary pollution of seawater. Self-purification processes are unable to neutralize the water continuously impacted by chemical inputs. Environmental conditions in river deltas are seriously threatened, mainly in deltas and river mouth beaches. The latter represent the most valuable natural water complexes that act as the principal fish spawning and foraging grounds as well as waterfowl nesting places and rest areas.

The second important source of pollutants transported to the sea are effluents discharged from enterprises, farms, or human settlements situated directly on the coast. An extremely adverse effect on the marine environment is produced by sudden discharge of pollutants resulting from accidents at enterprises of treatment facilities as well as various failures of sewage systems (Kaplin 1997). All maritime cities, first of all, Astrakhan', Baku, Makhachkala, Turkmenbashi, are sources of pollution that drop to the sea sewage waters (Zonn 1999).

The third major source of seawater contamination are accidents (oil and oil products spills) occurring during navigation and exploitation of offshore oil- and gas-fields, as well as, owing to the sea-level rise, flooded coastal oil fields. Accidental spills of oil and oil products are a source of significant damage to marine ecosystems, because concentration of pollutants may be extremely high exceeding the permissible level by hundreds or by even thousands of times.

Biological resources and fish reserves are affected as early as at the stage of seismic exploration of oil and gas fields. Special damage was caused by blasting operations that were responsible for the large-scale mortality of sturgeons. During drilling operations on the continental shelf, a special threat is posed by sustained discharges of liquid and solid wastes that are associated with the drilling process. Environmental consequences in the areas of offshore oil- and gas-field development are observable 5–12 km away from the drilling site and are manifested as high levels of oil pollution of water, bottom sediments, aquatic and benthic fauna and flora, and reduced species diversity of benthic communities and degradation of their structure (Kaplin 1997).

The sea receives water wastes from many sources. From 2 to 5 tons of heavy metals, 60,000–200,000 tons of petroleum products, and more than 5 million tons of organic pollutants are dumped into deltas every year. The sediments are, therefore, contaminated by heavy metals, especially lead and cadmium, and their concentrations exceed many times

those of the natural background. The composition of organic matter within the limits of deltas and their flood-plains causes deterioration of the oxygen regime and leads to hydrogen sulfide pollution of the beaches.

The flooding and associated contamination of land initiates specific biogeochemical processes whereby anaerobic gas generation is stimulated. The waters then become contaminated by metal compounds, heavy hydrocarbons, carbon dioxide, and nitrogen compounds, as well as bituminous substances and aromatic (benzene) polycyclic hydrocarbons. Generation of hydrogen sulfide poses the greatest danger (Kaplin 1995).

Transformations of ecosystems in all rivers of the Caspian Sea basin occurred due to hydraulic engineering and hydro-energetic projects, exploration of oil- and gas-fields, oil-chemical production, irrigation of nearshore territories, and increase of industrial and domestic water supply (Zonn 1999).

Cross-References

- [Barrier](#)
- [Beach Processes](#)
- [Changing Sea Levels](#)
- [Deltas](#)
- [Marine Debris-Onshore, Offshore, and Seafloor Litter](#)
- [Oil Spills](#)
- [Sea-Level and Climate Change](#)
- [Shore Protection Structures](#)
- [Water Quality](#)

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Bogs

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Terminology

The term bog is used to describe certain forms of wet terrestrial vegetation. Unfortunately, in common with the words employed for many other categories of wetland, there are variations and inconsistencies in usage, regionally (particularly within Europe) as well as globally. Bog has been broadly defined so as to encompass all types of peat forming vegetation (see entry on “► [Peat](#)”) or narrowly defined to denote only plant communities which are dependent upon precipitation and dust for supplies of water and nutrients. The term peatland is more appropriate for the former. The latter “ombrotrophic” condition may be an absolute state, however, this is by no means always the case and it is perhaps not surprisingly, therefore, that such communities have floristic affinities with other wetland vegetation types. Consequently, recent authors (Wheeler and Proctor 2000) have preferred to use the term bog to describe a type of vegetation, that is, one which is usually dominated by either (or a combination of) sphagna (mosses), ericoids (dwarf shrubs), or Cyperaceae (sedges). Bogs are characteristically base-poor (with a pH < 5.0) and generally, though not exclusively, occur over a substratum of peat. Rather than peatland, an additional term, mire is preferred here to describe all forms of wet terrestrial vegetation. Bogs frequently grade into base-rich mires, which in Europe are referred to as fens. Fens may be herbaceous or wooded (fen carr in Europe), for which in the United States the terms marsh and swamp are, respectively, commonly used.

Classification and Distribution

Mire vegetation is strongly influenced by hydrology and topography and hydro-topographical relationships have been widely used in mire classification schemes. Fundamental is the division made between the ombrotrophic (atmospheric) and minerotrophic (via surface runoff or percolating groundwater) supply of water and nutrients. Water-logged peat can accumulate above the groundwater level (to a maximum depth of about 10 m) leading to the formation of raised mires. Such mires are entirely ombrotrophic and therefore base-poor, supporting only bog vegetation. Other mires are to some extent influenced by a mixture of ombrotrophic and minerotrophic sources, local scale variations in which are expressed through hydrochemical and floristic gradients. Such mires are by no means always base-rich and bog vegetation will occur either where the soils of the catchment are poor in soluble minerals (minerotrophic bogs) or where precipitation/evaporation ratios are particularly high. The latter situation is not uncommon along the Atlantic seaboard of North America and northwestern Europe. Here extensive, primarily ombrogenous, “blanket” bogs can cover the landscape spreading over relatively steep slopes and descending down to sea level.

On a regional and continental scale, the geographic distribution of mire types reflects the climatic regime. The limit of ombrogenous bog development in southeastern Labrador, for example, coincides with the 1100 mm precipitation isopleth (Foster and Glaser 1986). Globally mires are more widely distributed at high and low latitudes and at high altitude. Nevertheless, extensive lowland mires, such as the Everglades, occur even in the subtropics and tropics where in coastal districts they merge into brackish marshes and mangrove swamps. Regional and continental level reviews of bog and related vegetation types can be found in Gore (1983).

A variety of mire types are found in coastal situations (see entry on “► [Wetlands](#)”). Some North American and northwestern European sites show complete spatial gradations from salt marsh, through fen and minerotrophic bog to raised bog. Similar temporal gradations can be reconstructed from peats deposited within coastal sedimentary sequences. Species composition in coastal mires is likely to be additionally influenced by the input of sodium and chloride via salt-spray, brackish groundwater, or as a result of flooding episodes.

Origins and Development

Mire communities can develop from waterbodies (a process referred to as hydrosal succession) or over formerly dry surfaces. Both circumstances apply in coastal situations where vegetation changes in stratigraphic sequences are frequently used to infer waterlevel movements resulting from

fluctuations in relative sea level. However, it should be noted that any environmental process influencing waterlevel elevation or nutrient status is capable of producing vegetation change. Such change can also result from internal (termed autogenic) processes, most obviously through the accumulation of sediment. Therefore, while the growth of mire over a dry surface indicates rising waterlevels, mires can develop over marine/brackish sediments as a result of falling, stable, or even slowly rising waterlevels (if exceeded by the rate of sediment accumulation).

Vegetation sequences and the processes influencing the development of coastal mires are reviewed in Waller et al. (1999) with particular reference to stratigraphic information collected from the Romney Marsh depositional complex in southeastern England. *Alnus glutinosa* (alder) dominated fen carr (swamp) vegetation developed, above salt marsh clays, and prevailed at sites close to the upland edge and in neighboring river valleys, from ca. 6000 to 2400 year BP. This community appears to have been sustained both by inflowing base-rich water and rising relative sea-level (preventing vertical isolation from groundwater). At sites immediately behind a coastal barrier peat formation began later. Here salt marsh clays are followed by a sequence of herbaceous fen, minerogenic bog, and ombrotrophic bog. Bog development appears to have required both vertical and spatial isolation from base-rich water sources. The former was induced by a decline in the rate of relative sea-level rise and by climate change. An additional factor appears to be the mobility of the peat matrix. Peat formed from herbaceous vegetation is more mobile than woody peat and acidiphilous vegetation developing on such surfaces is therefore less likely to be flooded with base-rich water. Spatial isolation seems to have been achieved by the extensive landward accumulation of peat and the presence of the barrier. The importance of the latter is demonstrated by the widespread occurrence of bog in back-barrier environments in the Low Countries during the Holocene epoch. Having achieved independence from groundwater the ombrotrophic vegetation of Romney Marsh was able to continue growing for a further 1000 C14 years after other mire types within the depositional complex were subject to renewed marine inundation.

Paleoenvironmental Reconstructions Using Bog Sediments

Sediments derived from bog, in common with other organic deposits, comprise an important paleoecological archive. Preserved plant material (seeds, pollen, and vegetative remains) and faunal remains such as Rhizopods (testate amoebae) and Coleoptera (beetles) can be used to elucidate *in situ* environmental changes and in some cases changes occurring in adjacent habitats. Analysis of the pollen preserved in organic

sediments has proved a particularly powerful tool for understanding long-term vegetation trends. Mires may also contain archaeological artifacts (see entry on “► [Archaeology](#)”) and in northwestern Europe a number of exceptionally well-preserved human remains, referred to as Bog bodies, have been recovered. Ombrotrophic bogs, being dependent upon precipitation for their growth, have additionally been an important source of information on climate change during the Holocene epoch. In particular, stratigraphic changes from darker more decomposed peat to lighter fresh *Sphagnum* peat have been taken to indicate periods of faster peat growth and therefore wetter climatic conditions. Changes in bog stratigraphy at many locations across northwestern Europe (including coastal locations) indicate such a climate shift occurred around 2650 year BP (van Geel et al. 1996). Unfortunately, changes in bog stratigraphy are not always synchronous between, or even within, bogs. Growth rates vary not only geographically but also in response to local hydrological features.

Organic sediments derived from coastal situations are commonly employed in reconstructions of former sea level as they can be radiocarbon dated and certain plant communities can be related to a specific (“reference”) waterlevel range (see entries on “► [Peat](#)” and “Sea-Level Indicators, Biological in Depositional Sequences”). The latter is clearly not the case with sediments derived from ombrotrophic bog. Given the difficulties distinguishing between ombrotrophic and minerotrophic bogs on the basis of floristic composition, and the gradations possible between these conditions, organic sediments derived from acidiphilous vegetation should be avoided when collecting material for this purpose.

Human Exploitation

Mires are exploited as a source of peat for fuel and horticulture and, following drainage, for cultivation. Ombrotrophic sediments are best suited for the former purpose, minerogenic for the latter. The extensive exploitation of peat for fuel can be traced back to the medieval period in northwestern Europe. For example, large quantities were removed from a series of valleys on the edge of the coast of Norfolk in eastern England. Subsequent flooding created a series of shallow lakes referred to as Broad. Peat continues to be a major energy resource in a number of countries (Russia, Ireland). Reclamation of the lowland mire complexes in Europe occurred from the seventeenth century onwards through the construction of effective drainage channels subsequently aided by pumping (see entry on “► [Reclamation](#)”). Such activities result in the lowering of the land surface both as a result of sediment compaction (the loss of interstitial fluid) and erosion (as the surface organic sediments decompose). At Holme Fen, in Eastern England, where *Sphagnum* is an important peat constituent, the ground

surface fell by 3.87 m between 1848 and 1978 (Hutchinson 1980). The large-scale exploitation of bogs has increasingly led to calls for their conservation. Along with other forms of wetland they are included within the RAMSAR Convention (see entry on “► Wetlands”).

Cross-References

- Archaeology
- Coastal Climate
- Hydrology of the Coastal Zone
- Peat
- Reclamation
- Salt Marsh
- Sea-Level Indicators: Biological in Depositional Sequences
- Wetland Restoration
- Wetlands

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Boulder Barricades

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Definition, Distribution and Historical Development

Boulder barricades are elongate rows of boulders that flank the coastline, separated from the shore by an intertidal flat (Fig. 1). They are the result of ice transport and therefore are

found only in Arctic and sub-Arctic regions. They are formed by the grounding of boulder-laden ice rafts in nearshore zones during spring ice break-up.

In North America, boulder barricades have been reported in Labrador (Daly 1902; Rosen 1979; Rosen and McCann 1980); Hudson Strait, east Foxe Basin, Baffin Island (Bird 1964); and the St. Lawrence River (Brochu 1961; Dionne 1972). In other areas they have been reported in the Baltic Sea and Fennoscandia (Lyell 1854; Tanner 1939). Løfken (1962) utilized uplifted boulder barricades in northern Labrador as an accurate sea-level indicator to delineate the Holocene regression. While Tanner (1939) observed a decrease in barricade development corresponding to a reduction in tide range from 1.3 to 0 m in Labrador, the features do occur in other nontidal (i.e., Baltic Sea) areas.

Lyell (1854) first recognized boulder barricades as an ice-deposited landform. Daly (1902) introduced the term, but believed that they were an accumulation of boulders at the seaward limit of wave backwash, with ice playing a secondary role. Tanner (1939) concluded that the features were the result of boulder-laden ice cakes piling up against a fixed shore icefoot. Conversely, Brochu (1961) hypothesized that intertidal ice-cakes were moved seaward during ice breakup, pushing boulders to the low water line.

Boulder Barricade Formation

Monitoring of coastal ice in central Labrador during both winter and spring breakup are the basis for a model for the entrainment of boulders into ice and the transportation during spring breakup. In tidal regions, such as Labrador, intertidal ice freezes downward with increased freezing at each high tide. Boulders are frozen in the ice and become lifted from the intertidal bottom. Observations on a broad intertidal flat indicate that more boulders are lifted from the upper intertidal zone, so apparently the less-frequent lifting of the ice during spring tides was more effective at encasing boulders than the diurnal lifting from the lower intertidal zone. High melting rates occur from the ice surface in the late winter, so the continued freeze-down and surfacemelt result in the transportation of boulders up through the ice.

In the Baltic Sea off Tallinn, Estonia, which is nearly non-tidal, the infrequent lifting of ice for freeze-down and encasement, and floating for spring transportation may be due to meteorological tides that are a major cause of sea-level fluctuations in the region (Maurice Schwartz, personal communication).

In spring when the snow cover has melted, boulders have been observed sitting on intertidal ice pans (Fig. 2). The intertidal zone breaks up before offshore areas because of numerous tidal cracks and the decreased albedo of the mud-laden nearshore ice. Shore leads up to 1 km wide serve as



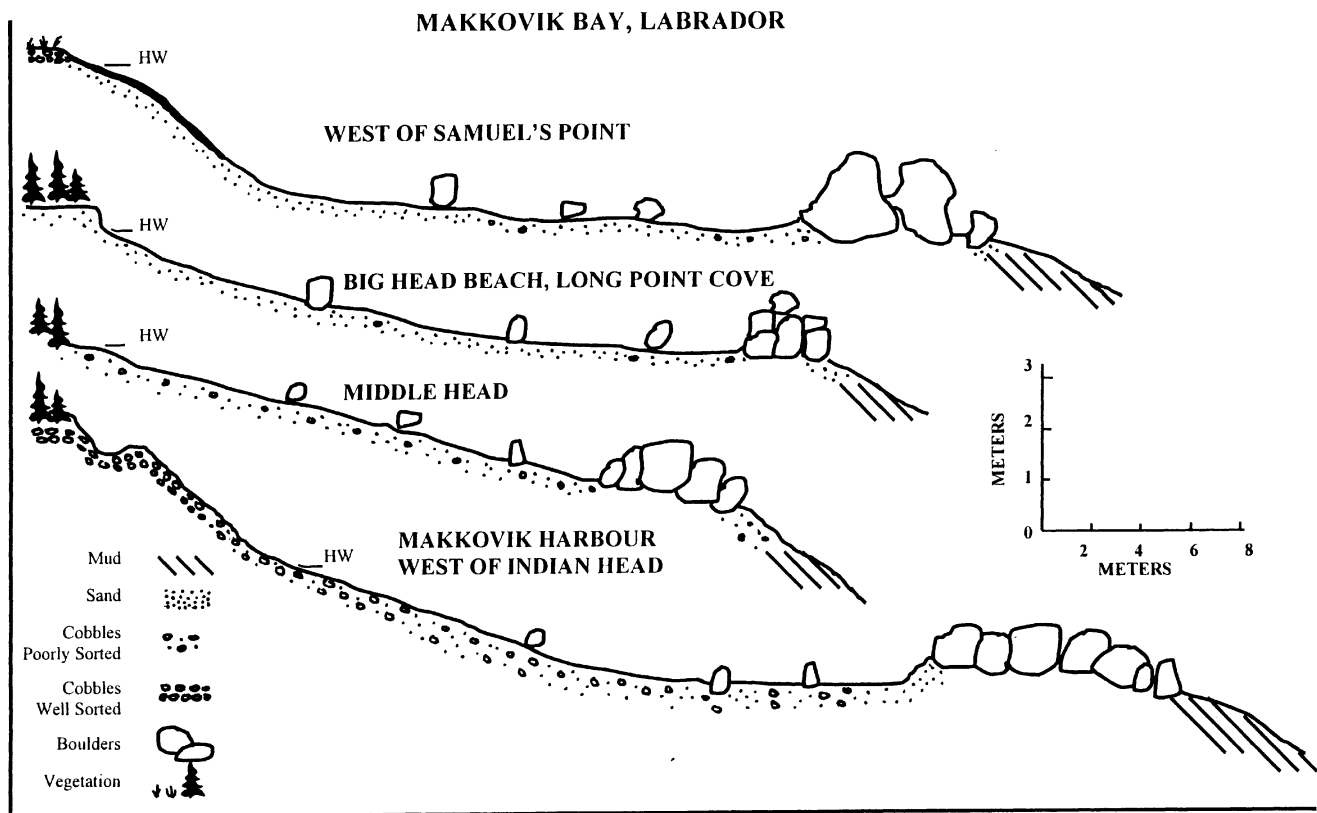
Boulder Barricades, Fig. 1 Boulder barricade in Makkovik Bay, Labrador



Boulder Barricades, Fig. 2 A boulder adrift on an ice cake, Makkovik Bay, Labrador

thoroughfares for these wind-transported ice rafts. The boulders may be randomly deposited in the nearshore as *boulder flats*, as commonly occurs on the deltaic flats at the heads of embayments. However, in central Labrador many of the

intertidal zones consist of uplifted marine clays with the top surface planed-off by contemporary wave and ice processes. This results in a slope-break near the low water line (Fig. 3). Since the ice thickness is comparable to the tide range, there is



Boulder Barricades, Fig. 3 Nearshore profiles at selected sites in Makkovik Bay, Labrador. Random boulders are common in the intertidal zone, while most accumulate as a boulder barricade landward of a steep

drop off to deeper water. (From Rosen 1982, with permission of Kluwer Academic Publishers)

a high probability for ice-rafts to ground at this position. Accumulation of boulders over successive seasons results in an intermittent barricade, which further serves to trap ice rafts during breakup. Landward of the slope break/boulder barricade position, random boulders, or boulder flats are also common.

At Tallinn, Estonia, the boulder barricades form in a similar setting. In this area, the nearshore is a rock-cut bench and the barricades accumulate at the seaward limit of this bench. (M. Schwartz, personal communication). Conversely, in the St. Lawrence Estuary, there is a range of nearshore boulder forms, including boulder flats, mounds, ridges, and pavements (Dionne 1972), which corresponds with no evidence of a nearshore slope break.

Summary

Boulder barricades are the result of grounding of boulder-laden ice rafts in the nearshore zone during spring ice breakup. Wind and tides are the major transport mechanisms. The requisite conditions for the formation of boulder barricades are: a rocky coastal setting, sufficient winter ice, and water-level fluctuations to entrain boulders in ice rafts; a

distinct slope break in the nearshore zone. Without the third condition, boulders will be desposited as boulder flats.

Boulder barricades are distinctly different from *boulder ramparts* and *ice-push ridges*, which are common on coastal and lake shorelines in that they are located seaward of the strandline, rather along the water line, due to the conditions discussed above.

Cross-References

- [Arctic, Coastal Geomorphology](#)
- [Ice-Bordered Coasts](#)
- [Paraglacial Coasts](#)

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Boulder Beaches

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Strictly speaking a boulder beach is one where the mean clast size meets the formal definition of “boulder” in the terms of the Wentworth grade scale, that is with a mean particle diameter of >256 mm (-8ϕ) (Wentworth 1922). However, the term is also sometimes used in a general sense to describe beaches where the sediment is a mixture of boulders and large cobbles.

Boulder beaches are found in high wave-energy environments where clasts of these large dimensions are released directly by erosion of bedrock, or where material is delivered to the shore zone by slope movements such as rockfall. In both cases sediment size is a function of joint spacing. Pre-formed boulders may also be supplied by erosion of Quaternary deposits such as glacial till, and by infrequent high-magnitude river floods.

While a considerable body of literature exists on gravel beaches, (usually concerned with pebbles and small cobbles), beaches of large cobbles and boulders have been neglected. This is partly because morphological response times are too long for consideration in the normal time-frame of academic field programs, and also because very large clasts are difficult to characterize. Some publications purporting to be general overviews of coarse sediments completely disregard boulders. Others set arbitrary upper limits to their area of interest in that, even where boulders form part of the beach sediment, sampling, experimental work, and subsequent analyses are confined to, at maximum, the cobble sizes. Most workers are reasonably precise about the lower size limit of the sediment studied, but the upper limit often remains vague: this may be partly due to the prevalent use of the ϕ scale, which becomes increasingly generalized in the coarser sizes. Terminology is

often imprecise, for example, the description “gravel” is commonplace, although it covers all clasts coarser than 2 mm diameter (-1ϕ), and so does not distinguish among pebbles, cobbles, and boulders.

Sedimentology

The few studies that have been carried out on boulder beaches have tended to look at details rather than broad patterns of sedimentation and morphology, for example, Bartrum (1947), Shelley (1968), and Hills (1970). The neglect of large clast beaches has led to attempts to apply sedimentation models derived from studies of pebble and cobble beaches to the boulder beach environment. Doubts concerning the validity of such extrapolations were fully confirmed by a comprehensive study of boulder beaches by H.L. Oak along the coast of New South Wales in Australia (Oak 1984). Oak proposed that boulder beaches demonstrate certain unique sedimentary characteristics that distinguish them as fundamentally different from pebble and cobble beaches. Hence, relationships established in the many studies of gravel beaches are often inapplicable.

The dominant characteristics of boulder beaches listed by Oak are:

1. A high wave-energy environment, competent to move large clasts.
2. Upbeach fining of sediment.
3. Abundant breakage of sediment.
4. Positively skewed size distributions.
5. Upbeach decrease in roundness.
6. No shape zonation.
7. No sphericity grading.
8. Low foreshore slopes, decreasing as particle size increases.

Of these characteristics numbers 2–4 and 6–8 contrast strongly with the known sedimentary characteristics of pebble and cobble beaches.

Sediment Size

Mean clast size is the most significant parameter determining the sedimentary character and behavior of a boulder beach. The pattern of general upbeach coarsening typical of smaller grade gravels is a function of two interlinked processes: (1) storm swash can carry most of the range of available clast sizes upslope, and (2) backwash is competent to carry a major part of this range at least some distance seawards. Neither of these is true of a boulder beach, and so clast size is the primary determinant of movement. Clast size decreases upbeach because the dominant boulders are so large that they can only be moved by traction as bedload, and even then perhaps only in very infrequent intense storms. Only the

sub-population derived from breakage of the larger clasts can be suspended. As wave uprush moves up the beach face, permeability reduces its volume, and gravity effects and turbulence reduce its velocity so quickly, that only increasingly finer material can be transported. Backwash effects are negligible, so the smaller clasts, including breakage products, remain where swash deposits them. Waves can move the larger boulders to the trim line at the base of the beach, but cannot move them any distance upslope.

As high-energy marine processes act on predominantly boulder-sized sediment winnowing of the fines produces a sediment assemblage dominated by a relatively small number of well-sorted large clasts, with a very minor subordinate population of smaller fragments, most of which have survived in the high-energy area only because of entrapment. The distinctive positively skewed size distribution of a boulder beach is attributed to the presence among the beach sediments of this tail of fines derived as breakage products of the dominant boulder population. Since large clasts resist continuous movement, spasmodic breakage during storms is the dominant size-reducing process (Bluck 1969; Matthews 1983). Abrasion is limited to the effects of passive sandblasting or the small movements of in situ abrasion, both of which are largely confined to a limited area at the base of the beach. In contrast, on pebble/cobble beaches breakage is minimal and most size reduction is achieved by attrition. The very fine products of this process will be removed in suspension, unlike pebble-sized breakage products, which may be retained on the beach.

Sediment Shape

Shape sorting (and the related characteristic of sphericity sorting) is poorly developed on a boulder beach. This forms a contrast with the characteristic shape zoning of a pebble beach, which results from selective clast transportation in which backwash plays an important role. On a boulder beach the sedimentation process is fundamentally different because the morphology is purely swash-formed. Selective shape sorting becomes increasingly ineffective where clasts are large and where wave conditions are turbulent. On a high-energy boulder beach shape sorting is insignificant because shape is only a dominant influence when entrainment forces are at critical thresholds for selective transport. When forces are not marginal, mass rather than shape, is the dominant control on net up- and down-beach transport potential.

Shape-controlled sorting processes are inoperative because, (1) even when storm waves are competent the prevailing bedload transport mechanism is basically insensitive to clast shape, and size will remain the dominant factor, (2) the large clast sizes and the high porosity of the beach means that backwash does not have the energy potential to create shape sorting, and (3) the rugosity of the beach surface militates

against the gravity-induced downslope movements of pebble grades, which are instead trapped in the voids between boulders. Only storm swash leaves its fingerprint and as a result the only primary structure imprinted on a boulder beach is swash-controlled upbeach fining. Size, therefore, exercises not only an initial control on upbeach sorting, but it is also the terminal control. As mean clast size decreases the size control typical of boulder sedimentation gradually gives way to the shape control associated with pebble and cobble sedimentation.

Sediment Roundness

Beach boulders are typically smoothed and rounded. On high-energy coasts clast transport, given free movement conditions, is very rapid. This casts considerable doubt on whether the angular, rough-textured blocks produced by wave quarrying and rockfall could possibly acquire such a degree of rounding and smoothing on a short (both spatially and temporally) unimpeded journey from source to boulder beach. Active and passive abrasion and rounding processes continue to take place in the beach environment, but their effects are largely confined to the base of the beach. Clasts at higher elevations on the beach face may have been emplaced by one high-magnitude storm. Marine influences will rarely reach these elevations, and only slow weathering processes can contribute to further rounding. It is doubtful whether the sum of these beach-face processes can entirely account for the evolution of clast shapes from an initially angular form controlled by geology, to the rounded, marine form characteristic of boulder beaches. It seems more likely that the majority of boulder beach clasts have spent some time in the intensely turbulent and abrasive hydrodynamic environment represented by traps such as potholes, gullies, and channels. Eventually a storm liberates the clasts to continue their journey to the beach.

Roundness is most pronounced toward the base of a boulder beach because the large clasts in this area experience marine action over longer time periods, and also because angular breakage products will be transported upbeach by wave action. Rounding cannot be taken as evidence of clast movements within the beach deposit, as rounded profiles can be attributed to pre-emplacement history (see above), and can also be created and maintained by in situ processes such as breakage, mutual attrition, and “water load abrasion”. The large well-sorted clasts found in the high-energy zone near the seaward margin of the beach characteristically demonstrate rounded, flattened ellipsoidal profiles. The remarkable stability of the lower part of a boulder beach in the face of high-energy wave action is probably due to a combination of large clast size, imbrication caused by strong unidirectional flows, and a variety of other fitting and interlocking processes acting on the beach fabric (Shelley 1968; Hills 1970; Bishop and Hughes 1989).

With distance upbeach angularity tends to increase, especially in the finer grades with more compact and platy shapes. This is a general comment, as the size- rather than shape-controlled swash transport mechanism will occasionally carry clasts exhibiting the full range of roundness well up the beach face. The major reason for the upbeach increase in angularity is the influence of breakage during infrequent storms. The products of breakage will remain as a component of the beach sediments because at higher positions on the beach face wave action that might winnow small-grade material is infrequent, backwash is ineffective, and the only agency acting to increase roundness is the relatively slow process of spheroidal weathering. On the boulders near the landward margin of the larger beaches, surface soundness deteriorates as weathering processes produce a rougher texture, and lichen colonization is common. In a general sense it is probably valid to consider the development of weathering rinds and lichen cover as indices of decreasing marine influence and movement. However, recent rockfall blocks on any part of the beach face carry a weathering/lichen signature from a subaerial environment, not a beach environment.

Sediment Orientation

On gravel beaches pebbles transported as bedload generally tend to be oriented with the long axis parallel to the shore, that is; transverse to the direction of swash movement. Such preferred orientation patterns are weakly developed on boulder beaches because a high velocity turbulent flow on a coarse bed leads to decreased regularity in orientation patterns. Clast collisions can change orientations, disturbing or perhaps even completely obscuring the pattern imprinted by the transport process. Clasts can also be oriented by the prevalent waves without undergoing net transport. Thus, while a bedload transport mechanism does generate preferred orientations, the generally weak development of this characteristic on boulder beaches is probably due to the interplay of high-energy wave action with the particularly rough surface of the beach deposit.

Beach Profile

On all types of beaches relationships involving slope are regarded as particularly significant because slope is usually considered the primary index of morphological response to wave action. There have been so few published studies on boulder beaches that discussion on their profiles must be tentative.

The most notable feature of the typical boulder beach profile is a lack of variation over time. Even during relatively severe storms changes involve only individual clasts, with the beach slope itself remaining unchanged. Some boulder beaches exhibit obvious concave upwards profiles. In detail,

many have one basically rectilinear main facet extending down to mean high-water mark or below, with overall concavity produced by narrow and rudimentary low angle facets in the intertidal area. The beach profile can be conceptualized as providing a “fingerprint” of the resultant of earlier swash/backwash interaction. On boulder beaches backwash is minimal so beach material is pushed shorewards to rest at an angle controlled by the balance of gravity and the dominant swash forces. For this reason, coarse beaches are more likely to be concave upwards than fine beaches.

On boulder beaches mean beach slope decreases as mean size increases. This characteristic appears to be in direct conflict with one of the basic tenets of coastal studies, that is, that coarser sediments produce steeper slopes, partly because the angle of rest is higher, but mainly because high percolation reduces backwash, which would tend to draw down material and lessen the slope. Shepard (1973) published a table in which the predicted average beach face slope for clasts in the 64–256 mm size range (-6 to -8ϕ) was 24° . However, the slopes of boulder beaches are considerably below this predicted value, an illustration of the dangers inherent in extrapolating from work on finer grade material. Most studies of boulder beaches record slopes in the range 6° – 14° with the mean lying around 12° .

Oak (1984) formulated an explanation for the finding that the slopes of boulder beaches are gentler than predicted. On all beaches of whatever mean sediment size, storms produce an equilibrium profile that is flatter than the pre-storm profile. The accepted principle that a beach must adjust to wave energy by flattening its profile holds true then, even for boulder-sized clasts. Beach angle does indeed increase with clast size, but only if waves can move all sediment. In practice only high-energy storm waves are competent to move large clasts, so the profile of a boulder beach is in fact a “lag” storm profile, adjusted to and formed by storm waves. On sand and pebble beaches lag times are short, and in the days after a storm infill will steepen the beach face. However, this does not happen on a boulder beach, as normal wave action cannot bring about a steep fairweather profile, so the beach typically exhibits a relatively low-angle storm profile. Therefore, the relatively gentle slope of a boulder beach, with its concave upwards profile, can be considered indicative of high swash velocities and minimum sediment storage, that is, a storm profile. The persistence of the profile simply reflects the fact that competent storms are infrequent.

Cross-References

- [Beach Sediment Characteristics](#)
- [Boulder Barricades](#)
- [Boulder Pavements](#)
- [Cliffed Coasts](#)

- [Cliffs, Lithology versus Erosion Rates](#)
- [Gravel Barriers](#)
- [Gravel Beaches](#)
- [Rock Coast Processes](#)
- [Shore Platforms](#)

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Boulder Pavements

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Striated boulder pavements can form either on intertidal surfaces in areas affected by floating ice (Martini 1981; Hansom 1983, 1986) or at the base of glaciers or on grounded ice sheets (A.G.I. 1974; Boulton 1978; Visser and Hall 1984). Pavements have also been described from fluvial environments (Mackay and Mackay 1977). Their distinctive nature also allows them to be used in the sedimentary record to assist in the reconstruction of past ice-affected environments (Eyles 1988). Pavements deposited subglacially are argued to be the result of accretion of boulders around an obstacle and to carry striations that are largely unidirectional. Although there are no detailed descriptions of such pavements forming in present glacial environments, they have also been described from the top surface of Quaternary deposits as well as buried within such deposits (Hansom 1983; Eyles 1988). Pavements formed on present cold-climate intertidal surfaces are thought to be the result of abrasion and bulldozing of boulder-lag surfaces by floating ice and small icebergs (Martini 1981;

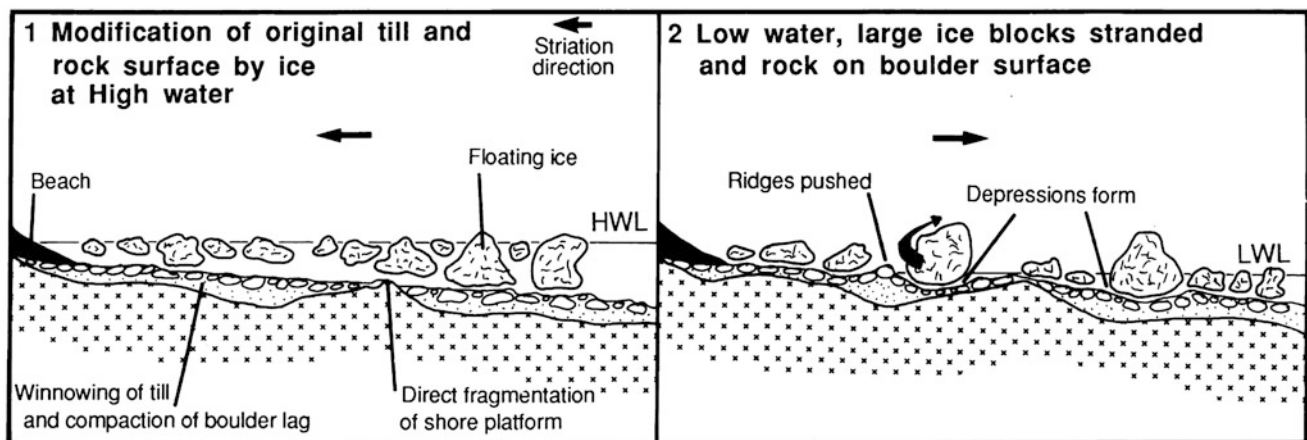
Hansom 1983, 1986; Gilbert et al. 1984; Forbes and Taylor 1994). The striations that the boulder surfaces carry are then controlled by the direction of movement of blocks of floating ice together with the rotational striations imparted when such blocks become stranded. Prerequisites for the development of intertidal boulder pavements are held by Hansom (1983) to be: (1) a boulder source; (2) frequent onshore movement of floating ice; and (3) a low-gradient intertidal zone. Given such conditions, the degree of development of the pavement seems to be controlled by the frequency of onshore ice movement, because the best formed pavements occur in areas subject to the highest frequencies of freely moving ice rather than areas that remain frozen for substantial parts of the year.

Marine boulder pavements are composed of smoothed boulders, often of up to 1 m in diameter, that are tightly packed together in the intertidal zone, the pavement surface appearing as a smooth, highly polished, and striated mosaic. The pavement surface is often interrupted by outcropping bedrock together with shore-normal furrows and polygonal depressions that can be up to 5 m across (Fig. 1). In the South Shetland Islands (see “► [Atlantic Ocean Islands, Coastal Ecology](#)”) they have been described as comprising a single layer of boulders underlain by a layer of clay containing locally derived lithologies, whereas in South Georgia, pavements are underlain by glacial till into which the boulders have been packed (Hansom 1983). The main processes involved in pavement development are summarized in Fig. 2. Floating ice blocks coming ashore onto a low gradient boulder-strewn shore bulldoze and pack loose boulders in the zone of grounding, initially in the upper intertidal but increasingly at the seaward edge. Some boulders may come from direct fragmentation of rock outcrops of any underlying shore platform that may exist and some may come from ice-raftered exotics. Polishing of the boulder surface is achieved by rock-shod floating ice abrading and striating the surface of the boulders (Hansom 1983). The orientations of the striations also inform the development processes of the surface polygonal depressions since the spread of striations on boulder ridges parallel to the shore can only be achieved by partially stranded ice blocks rotating on the pavement surface (Fig. 2).

In the Antarctic, boulder pavements are found in varying degrees of development across 10° of latitude from South Georgia to the Antarctic Peninsula and in Victoria Land, pavements of tightly packed and smoothed boulders locally veneer the shallow subtidal shore platforms of the Adare and Hallett Peninsulas (Hansom and Kirk 1989). The distribution and development of the Antarctic pavements show clear relationships between the frequency of floating ice grounding and wave processes. Where the frequency of ice grounding is high then the pavements are well developed. In the South Shetlands, the probability of floating ice and the percentage of ice concentration are both high. This limits the wave

Boulder Pavements,

Fig. 1 A well-developed boulder pavement in the South Shetland Islands, Antarctica. The polygonal depressions are caused by tidally grounded ice blocks which smooth and striate the boulders



Boulder Pavements, Fig. 2 A model showing the development of a boulder pavement at the extremes of the tidal cycle. Progressive stranding of ice blocks causes compaction, polishing, and striation of the boulders as well as forming depressions in the pavement surface

processes that destroy the pavement surface while ensuring frequent ice smoothing and packing. The result is a morphogenetic environment with a mix of ice/wave processes that is optimal for pavement development. Moving south away from this optimal ice/wave zone the incidence of grounding ice is reduced and so in the Antarctic Peninsula, wave processes are negligible, the incidence of fast ice is high and the frequency of ice grounding is low. Pavements here are poorly developed and embryonic. On the open coasts of South Georgia, well to the north of the optimal ice/wave zone, wave processes dominate, the frequency of grounding ice is low and so pavements are again poorly developed. However, within the sheltered inner fjords of South Georgia, wave processes are restricted, the frequency of floating ice is higher on account of glaciers calving into tidewater and, as a result,

the boulder pavements are better developed (Hansom and Kirk 1989). In the fjords of Vestfirðir in Iceland, similar forms also exist but are poorly developed as a result of the juxtaposition of a very limited ice-climate and a very low energy wave environment (Hansom 1986).

Martini (1981) suggests that the incidence of boulder pavements can be taken as a reliable indicator of intertidal ice action. Thus, the occurrence of emerged pavements at 5, 9, 12.5, and 17 m above sea level in the South Shetland Islands is convincing evidence of unchanged morphogenetic conditions in the area at least since the uppermost of the pavements was formed some 9000 years BP (Hansom 1983). Eyles (1988) uses boulder pavements in a similar way to reconstruct fluctuations in ice environments in the Gulf of Alaska during the early Pleistocene.

Cross-References

- [Antarctica, Coastal Ecology and Geomorphology](#)
- [Arctic, Coastal Geomorphology](#)
- [Boulder Barricades](#)
- [Boulder Beaches](#)
- [Glaciated Coasts](#)
- [Ice-Bordered Coasts](#)
- [Shore Platforms](#)

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Capping of Contaminated Coastal Areas

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The rapid and under-regulated development of industries in the past has left major contamination problems in several coastal and estuarine areas. Such contamination from historical chemical releases (industrial discharges, storm sewers, wastewater, landfill runoff, and leachate) severely impairs the ecological and recreational functions of the estuary/ocean. The discharged contaminants typically adhere to the fine sediments and settle in low-energy zones, which are conducive to deposition. Once settled, such sediments can exert significant oxygen demand, reduce benthic diversity, and result in poor water quality.

Contaminated sediment removal can be expensive because a high degree of efficiency and reliability is required in such operations. Capping is an attractive, nonintrusive and cost-effective method of remediating contaminated sediments in rivers and harbors, where draft restriction is not a major concern. The same physicochemical properties and hydraulic conditions that favored the initial adsorption to and deposition of the contaminated sediments typically favor successful containment by capping. Since capping is an *in situ* technique, it can often be accomplished at approximately 20–30% of the cost of a remedial dredging project.

Cap Definition and Considerations

Capping involves the controlled placement of clean sediment (usually sand) over the contaminated sediments in order to isolate them from the environment. The cap thickness is a function of several factors including desired breakthrough time, bioturbation, and contaminant flux. At high-energy

sites, the cap is often armored by a layer of stones. Where the size of the armor significantly exceeds that of the base material, a layer of intermediate size stones (filter layer) is provided to prevent hydraulic washout of the base material. Another special case is the use of a geotextile layer on top of the native sediment layer to provide added bearing capacity and load distribution, especially when the sediments are very soft and the cap is multi-layered.

To ensure sufficient protectiveness and isolation of the contaminants, the cap should withstand the worst design case physical and chemical events at the site. Major physical destabilizing forces include extreme water flows, storm waves, and tidal currents. Molecular diffusion, ground water-induced advection, hydrodynamic dispersion, and pore water flux due to consolidation are the major chemical destabilizing forces. A base sand isolation layer provides protection from chemical events and an armor/filter layer (usually riprap and gravel/cobbles) provides protection from physical forces. Bioturbation (sediment processing by benthic organisms) is another major destabilizing force causing mixing of the water/sediment interface, thereby releasing contaminants to the water column. Therefore, the design thickness of the cap should account for physical and chemical isolation, as well as bioturbation. Since the likelihood and type of organisms that may colonize the cap would vary from site to site, a site-specific survey of benthic organisms is often required. This could also be estimated by chemical and isotopic analysis of thinly sectioned sediment cores for sediment mixing depths.

The success of a capping operation depends on the proper selection of construction techniques for the cap and armor materials. Mathematical modeling of cap performance using design equations is frequently used since: (1) it can evaluate the effectiveness of the cap as a chemical barrier for various design parameters, and (2) it can quantify potential contaminant release rates to assess long-term effectiveness of the cap. A brief review of the cap design criteria and guidance for the selection of cap construction equipment are presented here.

Design Criteria

Special factors of interest to cap design/evaluation include cap behavior during placement, settlement of the cap, and native sediments following placement, potential contaminant release mechanisms, and stability of the cap in the long term (Mohan 2000). Design considerations for an underwater cap can be classified into the following six categories (Palermo 1991; Zeeman 1993; Mohan 2000):

Site investigations: A good field database is an essential starting point for an effective cap design. Investigations should include evaluation of water depth, currents, wave forces, flow velocities, extreme events (storms, floods), ice effects and forces, soil and foundation properties (strength, consistency, compressibility, and permeability), contaminant types and levels, and water column/biota contamination data.

Cap area: The cap should ideally cover the entire horizontal extent of the contaminated sediments. Mapping the horizontal distribution of the contaminants is essential to identifying this parameter. In special cases where sediment contamination is vast and widely distributed spatially within the waterbody, the extent of the cap should be determined based on human health exposure risks and isolation of the bulk of the contaminant mass to the food chain.

Cap movement during placement: Underwater caps may be placed using several techniques such as surface placement from barges, clamshell, or pipelines, placement using spreader boxes, tremie pipes, and underwater placement using clamshells. Several mechanisms are of interest as the cap is being placed. One of the main concerns during placement of the cap is the movement of the slug (velocity and properties) as it reaches the ocean or river bed. Theoretical considerations should include impact velocity of the sand, thickness and horizontal dispersion of the sand slug, and required number of lifts for obtaining the design thickness.

Geotechnical stability: Construction of a sand cap over soft native sediments may cause stability concerns due to the increased stresses in those layers. The primary stability concern during cap placement is the potential for the sediments to fail under the increased stresses, and to form “mudwaves” which

could be trapped in subsequent layers of sand cap material. This concern is particularly valid since contaminated sediments typically have very high water contents and low shear strengths, resulting from the lack of sediment consolidation. Cap stability analysis involves consideration of three major factors: (1) settlement of the cap and foundation material, (2) bearing capacity of the cap, and (3) slope stability of the cap.

Chemical stability: Chemical analysis is one of the major design elements since it defines the effectiveness of the cap in containing (and isolating) the contaminants from the water column. There are four major mechanisms that could potentially release contaminants through the cap system: (1) diffusion, in the absence of groundwater flow, (2) advection and dispersion, if groundwater flow through the cap is present, (3) release of pore water due to sediment compression, and (4) erosion of contaminated sediments due to pore water release during sediment compression.

Physical stability: An underwater cap must be designed for a specific level of protection. This implies that the cap/armor should be able to withstand the impact forces resulting from a certain return period event (typically the 100-year event). An armor layer is provided when necessary to protect the sand isolation layer from being eroded away due to hydraulic and environmental forces. In general, bottom velocities exceeding about 0.3 m/s (1 fps) can initiate sediment erosion. Considerations include protection from river flow, wave action, propeller wash forces, ice loads, and forces resulting from pore water release mechanisms.

Hydraulic filtering: The cap/armor should also be designed to withstand the impacts of the hydraulic filtering forces. There are two ways of providing this protection: (1) by installing a separate filter layer between the armor and cap layers, and (2) by installing a “graded” armor layer (i.e., with a wider grain size range) above the cap layer.

Construction Techniques

Table 1 lists the factors influencing the selection of cap construction equipment based on site- and material-specific properties.

Capping of Contaminated Coastal Areas, Table 1 Comparison of cap construction techniques (Mohan 1997, Proceedings Coastal Zone '97)

Equipment	Water depth	Material	Accuracy	Resuspension	Mixing	Cost
Barge discharge	Deep	Sand	Medium	Medium	Low	Low
Submerged discharge	Both ^a	All ^b	High	Low	Low	Medium
Skip box	Shallow	All ^b	High	Low	Low	Medium
Conveyor belts	Deep	Sand/rock	Medium	Medium	Medium	Medium
Dry/wet pumping	Both ^a	Sand/silt/clay	Medium	Low	Low	Low
Surface/underwater clamshell	Both ^a	Sand/rock	Medium	High	High	Medium
Modified dredges	Deep	Sand/rock	Medium	Medium	Medium	High
Thin-layer disposal	Deep	Sand	High	Medium	Low	Medium

^aDeep and shallow water application

^bMaterial types (sand, silt, clay, and rock)

Summary

Since the early 1980s, several pilot-scale and full-scale projects have been conducted to field-test the effectiveness of capping under a variety of site conditions (Strugis and Gunnison 1985; Palermo 1991; Wang et al. 1991; Fredette et al. 1992; Zeeman 1993; Averett and Francingues 1994; Nelson et al. 1994; Hull et al. 1998; Laboyrie and Flach 1998; Lillycrop and Clausner 1998; Shaw et al. 1998; and Mohan et al. 1999). These studies have established the validity of the environmental isolation provided by the capping process, thereby making capping a viable tool to be screened during coastal clean-up projects. Note that a final decision on the preferred remedy for a specific site should be based on technical, environmental, and economic analysis of the various alternatives (capping, dredging/capping, dredging/disposal, and no action) for the site.

Cross-References

- [Environmental Quality](#)
- [Geotextile Applications](#)
- [Human Impact on Coasts](#)
- [Marine Debris-Onshore, Offshore, and Seafloor Litter](#)
- [Water Quality](#)

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Carbonate Beaches

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Definition

Carbonate beaches have a significant proportion of the sediment fabric of biogenic origin and carbonate in composition.

Introduction

Carbonate beaches have a significant proportion of the sediment fabric of biogenic origin and carbonate in composition. Carbonate beaches are therefore wave deposited accumulations of sediment (sand to boulder in size) deposited on shores where a nearshore supply of biogenic debris is available. The carbonate detritus can originate in a range of environments and from a number of sources including coral and algal reef debris; mixed skeletal material including broken shells, foraminifera, bryozoa, gastropods, and ostracods; some terrestrial animals; and ooids.

In the tropics, they are associated with three separate environments. They are most commonly found adjacent to/or nearby coralline-algae reefs. They also occur in lee of tropical intertidal sand and mud flats where shells are winnowed to form shelly beach ridges-cheniers; and they may occur adjacent to carbonate banks in shallow tropical seas. In temperate regions, there are three separate sources of carbonate material. They are found adjacent to carbonate encrusted rocky shores and reefs. Temperate sea grass meadows along lower energy shores supply mollusks and other debris to sand flats and lower energy beaches, and are most extensive adjacent to some high-energy carbonate-producing, temperate continental shelves. Carbonate-dominated beaches also tend to exist in areas with

the above conditions, where there is insufficient material of terrigenous origin to significantly dilute the carbonate proportion, particularly in arid regions.

An addition, characteristic of many carbonate beaches is the postdepositional modification (diagenesis) of the sediment to form beach calcarenite (beachrock) in warm tropical waters, and aeolian calcarenite (dunerock) in more arid tropical to temperate environments. Once formed, the lithified sediments can have a dramatic impact on the nature and future evolution of the coastal system, particularly when the beach and/or dunerock is exposed to wave-wind processes. Even beaches with little carbonate may experience similar diagenetic processes particularly in warm tropical waters. Given this an additional definition of carbonate beaches are those whose processes and morphology are significantly impacted by the presence of carbonate grains and cements, particularly the formation of beach and dune calcarenite.

This entry will briefly examine the location of carbonate beaches, the major sources of carbonate sands, and the impact of the carbonate sands on beach morphodynamics.

Location

Carbonate beaches exist in tropical and temperate locations, including some in relatively high latitudes. The main prerequisite for a carbonate beach is a source of carbonate-producing detritus, and a mechanism to erode and/or transport it to the shore. Bird and Schwartz (1985) recorded 32 tropical and 14 temperate countries where carbonate beaches and beach and/or dune calcarenite have been recorded. Scholle et al. (1983) also provide a list of locations and references and where carbonate beach and dune systems have been recorded, Robbins and Magoon (2001) provide an overview of primarily tropical carbonate beach systems, and Short (2013) reports on Australia's 9,000 km long temperate carbonate coast and province.

Table 1 provides a summary of major beach carbonate sources and materials in tropical and temperate regions. The table and even the locations recorded in Bird and Schwartz is by no means exhaustive, as sediment characteristics are only available in the literature from a small minority of the world's beaches. This overview is based on the published data and locations at present. Future work will no doubt expand the range and perhaps types of carbonate beaches.

Tropical Locations and Sources

Tropical carbonate beaches generally occur between 25°N and S in association with three major carbonate-producing environments: coralline algal reefs, tidal flats, and carbonate banks.

Coral Reefs

Tropical carbonate beaches are predominantly located in association with coralline-algae reef systems, particularly fringing reefs, which lie in close proximity to the shore (Fig. 1), and all beaches associated with atolls and cays, where the eroded reef debris is the sole source of material. Tropical reef beaches include windward boulder ramparts formed in lee of exposed high-energy reefs, sand and coral cays formed on coral islands, and atolls and mainland beaches composed of sand through boulder debris (Scoffin and Stoddart 1983; Hopley et al. 2007).

Richmond (2001) provides an overview of tropical Pacific island carbonate beach systems including sediment sources, rates of supply, and physiographic settings.

Presence of an offshore reef is insufficient to guarantee carbonate beaches, as there must also be a mechanism to erode and transport the carbonate detritus from the reef to the shoreline. The world's longest reef system, the 2,000 km long Great Barrier Reef, lies between 20 and 200 km from the Australian coast. Most of the carbonate detritus is deposited tens of kilometers offshore close to the source reefs and does not reach the shore. The mainland coast is therefore

Carbonate Beaches, Table 1 Location, sources, and form of carbonate beach systems

	Source	Material	Beach morphology
Tropical			
Open coast	Coral-algae reefs	Coral-algae fragments	Reflective
Deltaic coasts	Tidal sand/mud flats	Mollusks	Cheniers, beach ridges
Shallow tropical seas	Carbonate banks	Ooids, skeletal, peloidal, pellets, aggregate sands	Beach ridges
Temperate			
Rocky coast	Encrusting organisms	Rocky shore biota	Variable
Low energy	Sea grass meadows, tidal sand flats	Mollusks, red algae foraminifera, bivalve fragments	Reflective/sand flats
High energy	Inner shelf	Mollusks, red algae, encrusting bryozoans, echinoids	Low to high energy + dune calcarenite



Carbonate Beaches, Fig. 1 Fringing coral reefs backed by carbonate-rich beaches at Cape Londonderry (left) and Point Blane (right), Northern Australia (A.D. Short)

dominated by terrigenous sediments, low in carbonate (<20%). Only scattered fringing reefs in the northern reef system actively supply carbonate sediments to the adjacent mainland beaches forming carbonate rich beaches (70–90%) (Short 2013).

Tidal Flats

Tidal sand and mud flats are extensive in tropical regions in association with river mouth deltas and protected embayments. These surfaces and adjacent subtidal sediments provide a habitat for a range of organisms, particularly mollusks. When exposed to episodic wave erosion and reworking the coarse carbonate detritus, together with any coarse terrigenous material, is eroded and reworked shoreward and deposited as shelly beach ridges, on sand flats, and cheniers, on mud flats (e.g., see Rhodes 1982).

Carbonate Banks

Carbonate banks occur in shallow tropical seas and are arranged perpendicular to shore as tidal-bar belts or shore parallel as marine sand belts. They may be composed of ooids and/or mixtures of skeletal, peloidal, pellets, and aggregate sands. Low tidal deltas and barrier islands surrounded by low-energy carbonate beaches can form in association with the banks. They have been recorded in South Florida and the Bahamas, and in Western Australia's Shark Bay (Scholle et al. 1983).

Temperate Locations and Sources

Temperate carbonate beaches occur outside the tropics in a range of temperate environments, from cool humid climates in Scotland and Ireland to Mediterranean climates in Greece and the Black Sea, but are most commonly associated with

arid to semiarid temperate coasts in North and South Africa and Southern and Western Australia. The three major sources of temperate carbonate sands are rocky shores, temperate sea grass meadows, and shelf biota.

Rocky Shores

Rocky shores play host to a range of encrusting carbonate organisms (Calhoun and Field 2001) whose death assemblages can be reworked onshore usually into carbonate enriched embayed or pocket beaches, with source headlands to either side and rock reefs offshore. Knuuti and Knuuti (2001) report on a Maine Beach (43°N) composed of 80% carbonate material consisting of barnacles, echinoids, mussels, and other material. Higher latitude carbonate beaches and dunes have also been reported in Ireland and Scotland (54 and 59°N) (Scholle et al. 1983).

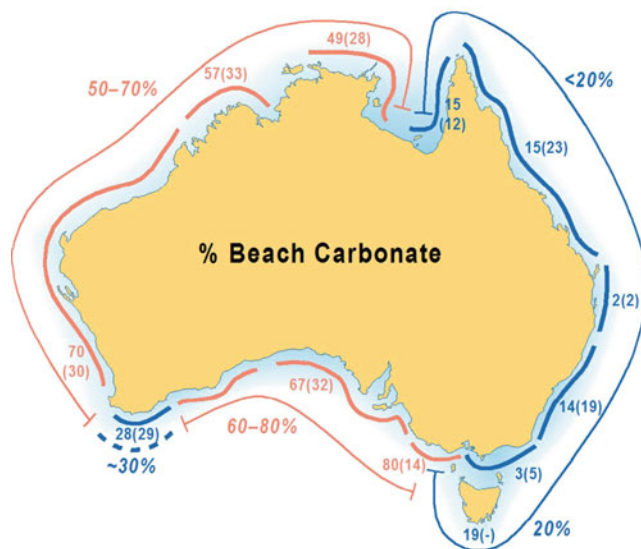
Sea Grass Meadows-Sand Flats

Sea grass meadows of *Posidonia* and *Zostera* provide a habitat for a range of mollusks, together with red algae, foraminifera, and bivalve fragments, as well as sea grass roots and detritus. This material is reworked shoreward to build intertidal sand flats backed by low-energy shelly beach ridges. Such systems occur commonly in the southern Australian gulfs and Great Australian Bight (Short 2013), and throughout Shark Bay on the central west coast.

Shelf Biota

Temperate continental shelves across southern and western Australia, around South Africa and along the western and north African, and southern and eastern Mediterranean coast contribute carbonate detritus to the most extensive temperate carbonate beach and dune systems. On the southern Australian shelf, James and Bone (2010) describe a shallow (30–70 m) inner shelf containing a range of

hard-bodied organisms particularly mollusks, red algae, encrusting bryozoans, echinoids and soft kelps, and sponges. The shelf is dominated by high-energy wave action, which erodes and transports carbonate detritus shoreward. Along the south and western Australian coast, it contributes to a 9,000 km coastal system containing 3,000 beaches occupying 70% of the coast, with regional beach systems composed of between 58% and 80% carbonate, and individual beaches reaching 100% carbonate (Short and Woodroffe 2009) (Fig. 2).



Carbonate Beaches, Fig. 2 Australian carbonate beach sands. Numbers indicate mean regional percentage of carbonate beach sand, with standard deviation in brackets. Note the dominance of carbonate sands across the southern and western coast derived from continental shelf biota, and in low-energy areas, from sea grass meadows, while in the northwest fringing reefs are a major source. (From Short and Woodroffe 2009)

Beach Systems

The beach systems associated with carbonate beaches behave similar to quartz beaches and are dependent on the sediment and wave-tide characteristics. They may be composed of boulders through fine sand, may range from wave to tide-dominated, and from reflective to dissipative. Their sediment characteristics and grain size is largely related to the sources and level of wave energy. Boulder coral ramparts and cays are associated with windward reef systems (Scoffin and Stoddart 1983). Most tropical carbonate beaches tend to be of low energy and composed of coarser sands and gravel including those in lee of reefs and sand-mud flats, the latter may also contain a high proportion of whole and shell fragments, a consequence of which is their description also as “shell ridges.”

Temperate beaches can also range from coarse to fine. Coarser sands and some gravel are usually associated with the lower energy beach ridges form in lee of sea grass meadows and their low-energy nearshore and/or intertidal sand flats. Shell Beach in Shark Bay, Western Australia, a very low-energy system formed in lee of shallow sand banks and sea grass meadows consists of 100% shells (Fig. 3). Local carbonate enrichment of energetic beaches by *Donax* species (commonly known as cockles, coquina, pippis) derived from adjacent foreshore habitats occurs through tropical and temperate locations.

The finest and best-sorted carbonate sediments are associated with shelf-derived carbonates on high-energy coasts. The erosion, transport, and energetic deposition process leads to a diminution of the source material into some of the finest (<0.1 mm), well to very well sorted, beach material. While there has been a limited amount of work in the sources of such material (e.g., Boreen et al. 1993; James and Bone 2010), there is no published work in the actual sources (and age) of the beach sands. This is



Carbonate Beaches, Fig. 3 Detail (left) and view (right) of Shell Beach, Shark Bay, Western Australia, which is composed of 100% mollusks derived from adjacent sea grass meadows (A.D. Short)

an area of considerable scope, particularly as it would provide information on the rates of contemporary carbonate sediment supply.

Diagenesis

A characteristic of carbonate beaches in all locations is the post-depositional modification of the carbonate fabric through geochemical processes, which commences soon after deposition. This is manifested as the commonly termed beachrock and dunerock. Beachrock, or more correctly beach calcarenite, refers to the precipitation from seawater of carbonate cements in the intertidal zone of tropical beaches, leading to the cementation of the intertidal beach zone. The beach and swash zone is an ideal environment for the precipitation of marine cements. Groundwater supersaturated with CaCO_3 is pumped by both wave and tidal activity through the usually coarse, highly porous, permeable sediments. The cement is derived from precipitation from the seaward matrix and not from solution of the in situ sand grains. Cements are typically aragonite or high Mg-calcite (10–14%). Cements precipitated from freshwater contain low Mg-calcite (5%). As a consequence, the rock normally has extremely well-preserved structures and is typically exposed as tabular, seaward sloping slabs of beachrock (Fig. 4). This process needs only a few years to decades to occur, though most would have occurred during the late Holocene.

Beachrock formation requires that sediments remain close to the surface and shoreline, the source of wave and tidal pumping. For this reason, they occur on stationary rather than prograding shores.

Once the beachrock is formed, it may remain buried and have little impact on contemporary beach processes. However, as the unconsolidated beach narrows the underlying beachrock will influence swash permeability and the groundwater table, leading to a super-elevation of the groundwater and exit point, both of which will enhance beach erosion.

When the rock is exposed at the shore, it behaves as an impermeable sloping seawall (Fig. 4), which will decrease permeability, enhance erosion, and retard sediment deposition and beach recovery. The rock itself may be eroded as tabular boulders, which are deposited to the lee as a boulder breccia (Fig. 5).

If the beachrock becomes stranded off the shore, as a shore-parallel or curvilinear reef, it will act as an attached or detached shore-parallel breakwater, with usually discontinuous form longshore. This affects wave breaking, attenuation, and refraction. Wave breaking on the reef may result in zero ocean wave energy in lee of the reef, either permanently and/or at low tide (Fig. 6). Wave attenuation over lower reef sections and/or at high tide will reduce waves in lee of the reef, while wave refraction over and around detached reefs will redirect the wave train. The impact of the above is first to decrease breaker wave energy at the shore and second to realign the backing sedimentary shore parallel to the refracted wave crests. As a consequence, shores in lee of beachrock reefs have lower energy beach systems and are more crenulate (Fig. 6).

In temperate and some more arid-tropical locations, longer-term pedogenic process in carbonate rich beach and dune deposits leads to the formation of calcrete (caliche). It consists of usually a loose thin top soil, underlain by laminar, breccia, and massive calcrete consisting of



Carbonate Beaches, Fig. 4 Beachrock detail showing well preserved, cemented beach laminations, (left); and shore parallel beachrock reef, both Cumurupin, Natal, Brazil (A.D. Short)



Carbonate Beaches, Fig. 5 Beachrock rich in coquina shells, backed by eroded rock forming a beachrock breccia, Marineland, South Carolina (A.D. Short)

recrystallized carbonate sands devoid of original structures, followed by mottled calcrete including lithocasts and lithoskels associated with vegetation roots and insect casts, and finally the loosely consolidated parent material, usually carbonate-rich dune sands (Scholle et al. 1983) (Fig. 7). Dune calcarenite occurs across North Africa to the eastern Mediterranean, in South Africa, and southern and Western Australia.

Dune calcarenite will have a similar impact to beachrock on beach morphodynamics. In addition, because calcrete will affect an entire beach-dune-barrier system, it is usually far more massive than beachrock and consequently have far greater impact on the adjacent or backing coastline (Short 2013).



Carbonate Beaches, Fig. 6 Three successive beachrock reefs at Shoal Point, Southern Australia (left); and linear beachrock reef at Port Gregory, Western Australia. Note the crenulate lee shoreline, and deep tidal, reef-controlled, channel (right) (A.D. Short)



Carbonate Beaches, Fig. 7 Detail (left) of laminar, breccia, and massive calcrete (left) and a 20 m high section of aeolian calcarenite containing two buried paleosols (right), Robe, South Australia (A.D. Short)

Summary

Carbonate beaches occur throughout the tropics in association with coral reefs, tidal flats, and carbonate banks, and in temperate latitudes adjacent to rocky shores and reefs, and in lee of sea grass meadows and carbonate-rich shelves (Short 2013). Each of these environments produces a range of carbonate detritus sourced from coral fragments, encrusting algae, bryozoa, foraminifera, mollusks, echinoids, and other animals. Depending on the erosion and transport processes, the detritus may be deposited at the beach as whole skeletons/shells, through to fine carbonate sands. Once deposited, it may contribute to backing coastal dune systems and be subject to cementation as beachrock in warmer tropical waters, and dunerock in arid to semi-arid carbonate rich beach-dune systems. Once formed, the calcarenite can be exposed or submerged as beachrock or calcarenite barrier reefs, causing wave breaking, attenuation, and refraction, which control the form of the lee coastline.

Cross-References

- Beach Sediment Characteristics
- Beachrock
- Cays
- Cheniers
- Coral Reef Coasts
- Coral Reefs
- Eolianite
- Sandy Coasts
- Tidal Flats

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Caribbean Islands, Coastal Ecology and Geomorphology

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Geography

Location and Topography

The main line of Caribbean Islands lie in an arc extending north from the coast of Venezuela and then trending westwards towards the southern tip of Florida; while other small group of islands lie off the northwestern coast of Venezuela and off the coasts of Central America. Most of the islands are located in the Tropics, and lie north of the 10°N line of Latitude and south of the 23°N Tropic of Cancer, although some of the islands in the Bahamas chain extend north of this line.

The Greater Antilles, consisting of four major islands – Cuba, Jamaica, Hispaniola and Puerto Rico – are, with the exception of Cuba, fairly mountainous with narrow coastal plains, while Cuba is dominated by extensive lowland plains. The Lesser Antilles consist of two chains, an outer chain from Barbados to Anguilla and continuing north to the Bahamas chain, consists of low-lying islands; while the inner chain from Grenada northwards are mountainous volcanic islands.

Geology

The Caribbean islands are quite young in geological terms, the Caribbean basin began forming less than 200 million

years ago. The West Indian Island arcs have been formed along a stress zone between the North and South American crustal plates. The inner chain of the Lesser Antilles developed along a line of weakness in the earth's crust. A series of submarine volcanoes formed, which rose above sea level to form the islands. Many of these islands still have signs of active volcanism such as, hot springs and sulfur emissions. Indeed there have been major volcanic eruptions in recent years, in particular Chances Peak in Montserrat, which started erupting in 1995 creating a volcanic emergency on the island and causing most of the population to flee to the northern third of the island or to other islands. This emergency continued until 1998, and now at the beginning of the twenty-first century, only the northern third of Montserrat is habitable and the dome on the volcano continues to grow at a significant rate. Other islands have also experienced eruptions in recent times, for example, Soufrière in St. Vincent and the Grenadines erupted in 1979.

The outer chain of the Lesser Antilles, and the Greater Antilles are older and have a more complicated history, consisting of volcanic episodes, alternating with periods of subsidence and sedimentation (Blume 1974). The outer chain of the Lesser Antilles consists of islands with volcanic bases capped by thick limestone. Islands such as, Puerto Rico have a central mountainous spine consisting of igneous rocks with coastal plains composed of sedimentary rocks on the north and south coast (Bush et al. 1995).

In more recent geological time there have been major changes in sea level, which have affected the geography of the islands. About 14,000 years ago, the sea level was 68 m below the present level and many associated islands, such as Puerto Rico and the Virgin Islands were joined by dry land. The sea reached near its present level about 5,500 years ago at that point the separation of most islands was completed. Trinidad became an island only in recent geological time, when erosion flooding of the Orinoco River separated it from the mainland across the Serpent's Mouth (Bacon 1978).

Climate

Lying in the Tropics, the islands have an equable climate with little change in temperature from month to month. Average temperatures in the Bahamas are 25.1 °C, and rise toward the equator, for example, 26.7 °C in Curaçao. Temperatures vary with altitude and aspect. Precipitation amounts vary from island to island and within islands depending on topography and aspect. Most islands have a wet and a dry season, with the main wet season being from May/June to October/November, some islands may have two rainfall maxima. The dry season is generally from December to April.

The Northeast Trade Winds determine the weather in the Caribbean throughout the year. These blow with great

constancy from directions between northeast and southeast. The Trade Winds are strongest between June and July and from December through March. During the first half of the year they usually blow from directions between northeast and east, veering more toward the southeast during the mid and latter parts of the year. The trailing ends of frontal systems moving from west to east over the southern part of the United States sometimes affect the Greater Antilles, particularly between October and April.

The Caribbean Islands lie in the hurricane zone. Many hurricanes originate as tropical waves off the west coast of Africa and travel across the Atlantic Ocean gaining energy from the warm ocean waters. As tropical waves strengthen, they pass through several stages, including tropical depression and tropical storm (see Table 1) before reaching hurricane strength.

A hurricane is an intense low-pressure tropical weather system with maximum surface wind speeds that exceed 118 km/h (74 m/h). Hurricanes are graded according to their sustained wind speed (wind gusts may be significantly higher) using the Saffir–Simpson scale of 1–5, see Table 1. The hurricane season lasts from June 1 to November 30, although rarely hurricanes can occur outside of these limits. September is the month when most hurricanes are experienced. Hurricanes generally move in an east to west/northwest direction across the Caribbean region, although exceptions have been recorded, particularly later on in the hurricane season, for example, Hurricane Lenny, a category 4 hurricane, formed just south of Jamaica in November 1999 and moved in an easterly direction and eventually hit Anguilla and St. Maarten.

Oceanography

Water temperatures are generally above 25 °C. The region is affected by the North Equatorial Current, which flows generally from east to west across the Caribbean Sea. The surface current then turns northwards through the Yucatan Channel, where a part flows into the Gulf of Mexico and the other part

Caribbean Islands, Coastal Ecology and Geomorphology, Table 1 Categories of tropical low-pressure systems

Category	Sustained wind speed	
	km/h	m/h
Tropical wave	Up to 36	Up to 22
Tropical depression	37–60	23–37
Tropical storm	61–117	38–73
Hurricane category 1	118–152	74–95
Hurricane category 2	153–176	96–110
Hurricane category 3	177–208	111–130
Hurricane category 4	209–248	131–155
Hurricane category 5	249+	156+

forms the Florida Current, which eventually becomes part of the Gulf Stream.

The average tidal range in the Caribbean is small, 0.3 m (1 ft), and 0.5 m (1.5 ft) at Spring Tides. Most islands experience a semi-diurnal tidal pattern. There are two very deep trenches in the Caribbean Sea. The Puerto Rico Trench follows a west–east trend and extends from north of the Dominican Republic to the northern islands of the Lesser Antilles, the maximum depth is 9,200 m. The Cayman Trench lies between Jamaica and the Cayman Islands.

Waves generally approach the region from the east, varying from northeast to southeast, depending on the Trade Wind regime. As a result, in the Lesser Antilles, where the islands generally trend north to south, there is a very clear distinction between the windward, exposed, east coasts with high wave energy, and the leeward, sheltered, west coasts. This distinction is not always so clear in the Greater Antilles, although here it is often the north and east coasts which are the most exposed to high wave energy.

During the Northern Hemisphere winter months (October to April), swell waves from North Atlantic low-pressure systems are often experienced in the Caribbean Islands. These swell waves travel southwards and impact north, east, and west facing coasts in particular. Also during the winter months, waves between the northwest and northeast may be experienced as frontal systems move across the southern United States and sometimes also affect the northern shores of the Greater Antilles.

Very high-energy waves from “unusual” directions may also be experienced during tropical storms and hurricanes. During such events, hurricane waves may directly impact the normally sheltered west coasts of the Lesser Antilles.

Biogeography

Small islands can be expected to have less diversity of species and habitats than continents. Newly formed islands may be isolated from land areas populated by plants and animals by large expanses of sea. And these islands can only be colonized by those species capable of making the sea crossing, as a result they will have fewer species than neighboring mainland areas (Bacon 1978). The degree of isolation within an archipelago is another factor. Trinidad was separated from the South American mainland in relatively recent times, thus its flora and fauna are very similar to the neighboring mainland. However, the “oceanic” islands of the rest of the Caribbean have no such obvious relationship with the mainland. Distance from the mainland is another factor, for instance the number of species of amphibians and reptiles decreases northwards throughout the Lesser Antilles.

Coastal Characteristics

It has been argued that islands do not have coastal zones, instead the whole island is a coastal system, extending from the highest peak to the offshore reefs. Everything that takes place inland, be it vegetation clearing or pesticide application, also affects the coast. The concept is obviously very relevant in some of the very small islands, such as the low-lying cays of the Virgin Islands and the Bahamas, and also some of the small volcanic islands where there is just one central peak or mountain range, such as Saba and Nevis. And while it also has merit for the mid-to large-sized islands, such as Dominica and Jamaica, for example, it is more difficult to apply here because of the overall complexity of those islands. It has further been suggested (Nichols and Corbin 1997) that an Island System Management framework represents a way to achieve the objectives of Integrated Coastal Management.

Beaches

Differences in the structure and character of an island’s rocks exert a strong influence on the coastline, whether it consists of cliffs, low rocky platforms, or some other form such as beaches or wetlands. With tourism playing such a major economic role in the Caribbean region, beaches are economically the most important part of the coastal zone. With the region’s dependence on “sun, sea, sand” tourism, the very economic fabric of some islands is determined by its beaches and their characteristics. This is not to say that beaches are unimportant for island residents, indeed the opposite is the case, since islanders value their beaches highly as places for recreation, sports, and simple enjoyment. Furthermore, island beaches provide natural protection for very valuable real estate located along the coastline.

Morphology. Beaches in the Caribbean Islands range from long straight beaches extending for several kilometers, such as on the east coast of Barbados from what is known as the East Coast Road, just north of Bathsheba, to Long Pond and Greenland; to tiny pocket, bayhead beaches, extending just a few hundred meters, and enclosed by cliffed or rocky headlands, such as Lime Kiln Bay on the west coast of Montserrat. There is endless variety, not only in shape and geomorphology, but also in sediment size and type, and the presence or absence of rivers. Incoming wave energy and the presence or absence of coral reefs, are other important factors controlling coastal morphology.

Beach material. The word “beach” in the Caribbean Islands is usually taken to imply sand beach. However, geologically, beaches may consist of other sizes of particles (clay, silt, sand, gravel, cobbles, or boulders). In this entry, however, and given the regional importance of sand beaches and

tourism, the word “beach” implies sand beach, unless otherwise qualified. The color of the beach sand usually reflects its source. In Anguilla, for instance, the beaches are for the most part composed of white calcareous sand, with patches of shells mixed in, this sand is derived from the offshore reefs and the breakdown of the limestone of which the island is composed. In Dominica, many of the beaches are composed of black sand with grains of olivine and magnetite, brought down to the coast by rivers, and derived from the erosion of volcanic rocks. The yellow silica sand along Trinidad’s north coast is derived from the erosion of the Northern Range’s sedimentary rocks.

Not all beaches consist of sand, larger size ranges from gravel to boulders, make up many beaches in the volcanic islands, for instance at Argyle on the east coast of St. Vincent the beach is composed of large stones and boulders. Some of the more exposed beaches on the south coast of Tortola in the British Virgin Islands are composed of large coral fragments and boulders.

The geology and the wave energy determine the type of material on the beach. Strong wave action washes out the sand particles leaving a steep beach profile as is often found on the Atlantic-facing beaches. While calmer conditions often result in fine sand being deposited and a gently sloping beach. Some beaches have different types of material according to the season, for instance some leeward beaches in the Lesser Antilles have stone beaches during the months with higher wave energy, October to April, while during the calmer summer months, sand is moved onshore to cover the stones.

A rock formation found on many Caribbean beaches, either on the beach or seaward of the beach is called beachrock. This consists of sand grains and other beach material, including stones, cemented together by calcium carbonate. Beachrock forms within the body of the beach, beneath the sand surface and near the water table. Once the covering sand has been eroded, the beachrock formation hardens into rock. Exposures of beachrock are indicative of erosion (Cambers 1998).

Beach flora and fauna. Conditions above high water mark in the spray zone are very unfavorable for plants because of the dry sand and lack of water. Plants that grow here must store rainwater in succulent tissues, and have small-sized leaves to reduce water loss. Two plants often found above the high water mark are the Beach morning glory (*Ipomoea pescaprae*) and the grass Seashore dropseed (*Sporobolus virginicus*). They stabilize the sand surface and make conditions more favorable for other species such as Beach bean (*Canavalia maritima*), and Sea purslane (*Sesuvium portulacastrum*). However, high waves during storms frequently destroy this temporary beach vegetation, which then has to begin again establishing a foothold on the sand.

Inland from the spray zone there is a belt of woodland trees. These must withstand salt spray, strong winds, and saline soils. Common species include Sea grape (*Coccoloba uvifera*), Seaside mahoe (*Thespesia populnea*) and the West Indian almond (*Terminalia catappa*). These all have deep roots, which play an important role in stabilizing the sand. They also act as a windbreak. Another important tree is the Manchineel (*Hippomane mancinella*), this deep rooting tree plays an important role in slowing down beach erosion, however, the milky juice, which exudes from the branches when cut, may blister the skin, and the green fruits or small “apples” are poisonous. Another tree, the Casuarina (*Casuarina equisetifolia*), is an introduced species, which grows very rapidly in the coastal environment. However, it casts dense shade and produces a “carpet” of thick needles, which reduce the growth of ground cover. The coconut palm (*Cocos nucifera*), while not a native tree (it was introduced from the Indo-Pacific region) is very common in the islands and is often planted by tourism developers. However, it is very shallow rooting and is often the first tree to fall during high winds associated with tropical storms and hurricanes.

The beaches provide a habitat for a variety of crustaceans, mollusks, and other animals. Common species include the ghost crab (*Ocypode*), the Chip-chip (*Donax*), the Sand dollar (*Mellita quinquesperforata*). Jellyfish sometimes washed up on the beach include the Portuguese man-of-war (*Physalia*), and the Tennis ball jellyfish (*Stomolophus*). Birds seen at Caribbean beaches include the Brown pelican (*Pelicanus occidentalis*), Brown booby (*Sula leucogaster*), and Frigate bird (*Fregata magnificens*).

Sea turtles nest on beaches in the Caribbean Islands, the most common species are the Leatherback (*Dermochelys coriacea*), Green (*Chelonia mydas*), Hawksbill (*Eretmochelys imbricata*), and Loggerhead (*Caretta caretta*). Most of these species are declining as a result of over-harvesting, loss of habitat, etc. Many islands have laws defining closed seasons or total bans on the catching of sea turtles or poaching of their eggs.

Beach changes. Coastlines are areas of continuous change where the natural forces of wind and water interact with the land. These changes have been taking place for millennia. History provides the example of Cockburn Town in the Turks and Caicos Islands, where Back Street had to be renamed Front Street at the beginning of the twentieth century as erosion took its toll.

Beach changes have been measured in many of the Caribbean Islands since the 1980s as part of a regional program sponsored by the United Nations Educational, Scientific and Cultural Organization (UNESCO) and the University of Puerto Rico Sea Grant College Program. (This monitoring program is known by the acronym COSALC, Coast and Beach Stability in the Caribbean Islands Project, or as it has

been recently renamed “Managing Beach Resources and Planning for Coastline Change, Caribbean Islands”).

Beaches change seasonally in the Caribbean Islands, in many islands a winter/summer pattern can be distinguished, with beaches eroding during the winter months, October to April, as a result of higher Trade Wind waves and winter swells from the North Atlantic storms, and the same beaches accreting in the summer months when wave energy is lower. However, the pattern often varies between different coasts, so it is unwise to generalize.

An initial analysis of the beach monitoring results (Cambers 1997a) showed that 70% of the beaches had eroded and 30% had accreted. Mean annual erosion rates varied between 0.27 and 1.06 m/year. Natural and anthropogenic factors are involved. The natural factors include winter swells, tropical storms and hurricanes, and sea-level rise. Among the most important anthropogenic factors are beach and dune mining, building too close to the beach zone, badly placed sea defenses, and deterioration of nearshore coral reefs.

Hurricanes appeared to be the major factor controlling the erosion. For instance, in Dominica (Cambers and James 1994) there was significant erosion after Hurricane Hugo in 1989, followed by accretion in the years 1990–92. However, the beaches never returned to their prehurricane size, before further hurricanes impacted Dominica in 1995.

During 1995, three tropical storms/hurricanes hit the eastern Caribbean Islands within a 3-week period: Tropical Storm Iris, Hurricane Luis, and Hurricane Marilyn. There was extensive damage, both to the islands’ infrastructure and environment especially erosion of the beaches. The COSALC database was used to determine the extent of erosion of the land behind the beach (dunes or coastal lands) as a result of these events. Dunes on the north coast of Anguilla retreated as much as 30 m as a result of Hurricane Luis. Erosion rates were much lower further away from the hurricane center.

During the last part of the 1990s, the islands in the Lesser Antilles, from Dominica to Anguilla, experienced several tropical storms and hurricanes, leading to serious erosion. These numerous high-energy events appeared to introduce a certain vulnerability to the coastal systems making recovery slower and less sustained.

Beach protection. The increasing coastal development in the Caribbean Islands, particularly over the last two decades, coupled with the ongoing coastal erosion, has resulted in many property owners and sometimes government agencies building structures to slow down or stop the erosion. These structures include revetments, seawalls, groins, and offshore breakwaters. They have had varying degrees of success. Some islands, for example, St. Maarten and Anguilla, have attempted to restore their beaches by nourishing them with sand from offshore. Again, these measures have had varying degrees of success, at Maunday’s Bay in Anguilla, the beach was nourished three times during the period 1995–99, only to

have the sand washed away by a high-energy or hurricane event in 1997, 1998, and 1999.

There are no easy answers to this dilemma, especially since every beach is individual and must be treated as such. The use of adequate coastal development setbacks, which ensure that new coastal development is placed a “safe” distance away from the active beach zone has been proposed. A methodology has been developed, based on using existing data on erosion, hurricanes, and sea-level rise to determine a specific setback distance for each beach (Cambers 1997b). Thus, beaches experiencing more erosion have higher setback distances. This methodology is at present being applied in Anguilla and Nevis.

Undoubtedly erosion is going to continue in the Caribbean Islands, and the problems of how to protect beachfront property and to maintain beaches, will likely intensify.

Cliffs and Rocky Shores

Limestone and volcanic cliffs abound in the Caribbean Islands. Caves and “blowholes” exist in limestone cliffs in many islands, for example, Barbados, Bahamas and Turks and Caicos Islands. Steep rock slopes and cliffs are the most predominant coastline type in some of the volcanic islands, for example, Montserrat and Dominica.

Rocky shores are characteristically areas of rapid erosion, and they provide a great variety of habitats for marine plants and animals (Bacon 1978). Animals include barnacles such as *Tetraclita squamosa* and *Chthamalus fissus*, snails of the genus *Littorina*, mussels such as *Brachidontes exustus*, and crabs such as the Rock crab (*Grapsus grapsus*). Sea eggs, for example, *Lytechinus*, may be found on limestone and other hard rocks. Green algae are abundant and include *Chaetomorpha* and *Cladophora*, and some species of red and brown species may also occur. Boulders may be seen encrusted with calcareous algae.

Sand Dunes

Sand dunes are mounds of sand that often lie behind the active part of the beach. In the Caribbean Islands, they range from very low formations, 0.3–0.6 m in height to large hills of sand up to 6 m high. There may be several parallel rows of dunes. Dunes form when sand is carried by the wind from the dry beach area to the land behind the beach. When the wind encounters an obstacle such as a clump of vegetation it slows and the sand is deposited. Significant sand movement will take place when the wind speed, measured at a height 1 m above the ground, exceeds 6 m/s (Bagnold 1954). In the Caribbean Islands, average wind speeds equal or exceed this value in the months from June to July and from December through March. During storms/hurricanes, dunes are often eroded and the sand may be deposited offshore, after the storm/hurricane, the sand is returned to the beach and the slow

process of dune building may recommence. As a result of Hurricane Luis, dunes in Anguilla showed an average retreat of 9 m, recovery is expected to take decades, if it occurs at all (Cambers 1996).

Dune vegetation promotes the trapping of sand, however, as with beach vegetation, dune plants have to adapt to harsh conditions including high temperatures, dryness, salt spray, and the deposition of sand (Craig 1984). Dune vegetation in the Caribbean Islands includes the same plants as those described colonizing beaches. In addition, Sea lavender (*Tournefortia gnaphalodes*) has been observed to be a very effective shrub in stabilizing dunes.

Unfortunately sand dunes are often seen as prime sites for sand mining activities in the Caribbean Islands. Many dune areas have completely disappeared as a result of mining for construction material, for example, at Josiah's Bay in Tortola in the British Virgin Islands, and at Diamond Bay on the east coast of St. Vincent, where extensive black sand dunes, more than 6 m high have now gone. At Isabela on the north coast of Puerto Rico, the mining of an extensive system of dunes has taken place in the last two decades, leaving just a narrow strip of residual dunes, which are sometimes breached during high wave events.

Efforts have been made in some island to restore dunes using sand fences, for example, at Arecibo in Puerto Rico, and Shoal Bay North in Anguilla. Research has shown that a sand fence with a 50% porosity (ratio of open space to total area) can result in a 1.2 m high dune forming in 12–24 months (Clark 1995). Work with sand fences in the Caribbean islands and elsewhere has shown that one of the critical factors in dune growth is the width of dry sand beach in front of the fence.

Mangroves

In estuaries, lagoons and coastal mudflats, characterized by mud deposits and sheltered wave environments, mangrove ecosystems have developed characterized by trees of the red mangrove (*Rhizophora mangle*). Red mangroves have prop roots and can grow on newly deposited sediments and on sand bars and coral reef crests (Bacon 1978). Large mangrove swamps can be seen on the Gulf Coasts of Trinidad, and the south coast of Jamaica in the Portland Bight area. As sediment gradually accumulates around the roots of the red mangrove, conditions become suitable for the establishment of the black mangrove (*Avicennia germinans*). This species has a horizontal root system from which arise "breathing tubes" pneumatophores. A third mangrove species grows among both red and black mangroves, the white mangrove (*Laguncularia racemosa*). A fourth mangrove species, Buttonwood (*Conocarpus erectus*) is considered a transitional species between true mangroves and terrestrial vegetation.

Coastal mangroves are very important to fisheries and marine productivity. They provide a habitat, which is a breeding ground and nursery area for many species. Juvenile fishes live in the mangrove swamps until they are mature enough to migrate to the open sea, these include young butterfly fish, angel fish, tarpon, snappers, and barracudas, in addition there are juvenile spiny lobsters and shrimps. Other animals include worms, small crustaceans, and sea cucumbers. Clusters of oysters cling to the red mangrove roots, there are two common species, the mangrove oyster (*Crassostrea rhizophorae*) and the flat tree oyster (*Isognomon alatus*) (Jones and Sefton 1978). There are also sponges and sea anemones. Numerous insects, lizards, tree snails, and birds also live in the mangrove swamps.

There is a complex food web within the mangroves, starting with the mangrove leaves which fall and decay and are then eaten by shellfish, shrimps, crabs and fishes, which in turn become the prey of other juvenile fishes.

Because of their intricate root systems, mangroves also provide good protection to the coastline during storms and hurricanes. As a result, mangrove lagoons, such as Paraquita Bay in Tortola in the British Virgin Islands, are used as a safe mooring area during storms and hurricanes (Clarke and Lettsome 1989).

Coral Reefs

Among the most ecologically diverse systems in nature, coral reefs play an important role in the protection and formation of many Caribbean beaches. A coral reef is made up of many millions of coral polyps, tiny animals that secrete limestone to form themselves a stony skeleton. A healthy coral reef is home to many different plants and animals. Stony or hard corals are the main reef builders, while soft corals for example, the sea fan (*Gorgonia flabellum*) are flexible.

The main types of reefs found in the Caribbean are fringing reefs, which border a coastline, for example, there are several fringing reefs along the west coast of Barbados; patch reefs, which are isolated clumps of coral sometimes measuring only a few meters in diameter; and barrier reefs, which are separated from the coastline by a deep lagoon or channel, for example, the barrier reef off the east coast of Andros Island in the Bahamas.

Coral reefs are especially important to beaches because they protect the shore from high waves. Fringing reefs and barrier reefs often grow very close to the sea surface. Incoming waves break and expend their energy on the reef, and while secondary waves may reform in the lagoon and ultimately affect the beach, they are much lower in energy. Thus in many ways reefs are also natural breakwaters.

Coral reefs also act as a sand source for many beaches. Many fishes, for example, the Parrot fish (*Scarus*), Butterfly fish (*Chaetodon*), and Trigger fish (*Balistes*), actively feed on

the coral, biting off chunks of coral and excreting coral sand, which is then moved towards the beach by wave action.

Coral reefs are vulnerable to natural forces, such as hurricanes, increased seawater temperatures, human activities such as pollution, offshore dredging, and ship anchoring. Hurricane Luis, a category 4 hurricane, passed over Anguilla in 1995. A quantitative offshore benthic survey had been conducted the previous year (1994), and comparing this with a post-hurricane survey in 1996 showed that 61% of the intact live reefs, either hard coral or soft corals, were degraded to rubble or, bare rock. Previously intact, though dead, elkhorn coral reefs (*Acropora palmata*) were largely reduced to rubble. Damage was most severe at the inshore reefs (Bythell and Buchan 1996).

It may take some years before the damage to a coral reef is manifested at the beach itself, such as through a reduced breakwater effect. And then, it may be difficult to separate out this cause from other causes of beach erosion.

Sea Grass Beds

Sea grass beds are important habitats found in shallow near-shore waters. They are often found between the beach and coral reefs. There are three main types of sea grasses: Turtle grass (*Thalassia testudinum*), Manatee grass (*Cymodocea filiforme*), and Shoal grass (*Diplanthera wightii*). Sea grasses are flowering plants, not algae, and they spread mainly through a system of underground creeping stems. They are an important marine habitat for many marine animals and their larval stages. They also provide foraging grounds for commercially important species such as grunts and conch. Sea grass beds maintain clear coastal waters by filtering sediments from the water and holding them down with their complex root systems. Calcareous algae living among the sea grass beds produce carbonate sediment.

Sea grass beds are very vulnerable to hurricanes, and often after a hurricane, beaches may be covered with thick mats of dead sea grass, more than 1 m thick. Hurricane Luis in 1995, reduced the sea grass bed cover by an average 45% in Anguilla (Bythell and Buchan 1996).

Problems Facing Caribbean Islands' Coasts

Tourism development. Tourism growth is taking place in most of the Caribbean Islands, indeed the tourism business has been termed “the engine of growth” (Patullo 1996). The Caribbean’s share of world tourist arrivals is less than 2%, but this is triple that of South Asia’s share. The insular Caribbean ranked ninth in the world on the basis of its tourism receipts in 1990, and in 1994 the Caribbean Tourism Organization estimated that its 34 member states grossed US\$ 12 billion from tourism. Tourism is a big business in the Caribbean Islands.

However, besides bringing economic growth and employment for many, tourism also brings environmental problems,

partly because in its early days, in the 1970s and 1980s, tourism growth was largely unregulated. This created a precedent and by the time environmental controls were being brought into place, around the early 1980s, much of the damage had already been done. Islands such as Barbados were already facing serious pollution and degradation of their nearshore fringing reefs, Jamaica was dealing with beach exclusivity such that local residents could no longer gain access to certain prime beach sites, and even small islands, such as Montserrat, were seeing a construction boom and increased mining of their beaches.

Efforts to try and reverse some of these trends dominated the 1990s and are continuing today. However, it is difficult to find many success stories, since environmental concerns usually come second to economic growth priorities. And with so much of the Caribbean’s tourism industry, from the hotel rooms to the airlines to the tour operators, being controlled by foreign interests (Patullo 1996) it is sometimes difficult for the islands to take control of this vitally important industry. Yet, in order for the islands to truly benefit from tourism in the long term, this is what must happen.

Climate variation. Predictions about global climate change as a result of global warming still have a considerable degree of uncertainty. However, climate variation can be traced back over the last century at least, and can be used to predict climate trends in the near future. For instance, hurricanes in the North Atlantic Basin have been shown to have a cyclical pattern (Sheets and Williams 2001). There are periods of two to three decades when there are many hurricanes – active periods, followed by periods of similar duration when there are fewer hurricanes – quiet periods. Records show that the 1900s to the 1920s was a quiet period, followed by an active period from the 1930s to the 1960s. Recently, there has been a quiet period, which extended from 1971 to 1994. However, since 1995, the Atlantic Basin has entered an active period, with the last 5 years of the 1990s showing an increased number of hurricanes and in particular an increased number of very intense hurricanes (category 3+). For instance, the years 1995–99 showed an average of four category 3+ hurricanes per year, compared with the previous 5 years, 1990–94, when the average was one category 3+ hurricane per year.

While these are average figures referring to the entire Atlantic Basin, this trend can be seen in many of the Caribbean Islands, particularly the northern Lesser Antilles, from Dominica to Anguilla, where many islands have experienced at least one hurricane per year since 1995. This climatic variation, while not yet fully understood, appears to be related to several climatic factors such as the rainfall in West Africa, upper level winds and ocean currents. With the massive tourism-related coastal development, which has

taken place in coastal areas in the Caribbean Islands since the 1970s, this trend is of considerable concern to the islands.

While not directly related to climate variation, another infrequent phenomenon, about which the Caribbean Islands must become increasingly concerned, is the threat of a tsunami. Given its tectonic instability, the region is long overdue for a tsunami event. The last major tsunami in the Caribbean was on October 11, 1918, when a tsunami generated by an underwater earthquake, hit northwestern Puerto Rico and caused considerable damage as well as more than 100 deaths. The increased coastal development that has taken place in recent decades makes the Caribbean Islands very vulnerable to a tsunami threat. Added to this, is the fact, that as yet, there is no tsunami warning system in place in the Caribbean region.

The challenges for integrated coastal management in the Caribbean Islands. In seeking sustainable development, the Caribbean Islands face many problems. These include their small size, relative isolation, vulnerability to natural disasters, limited economic diversification and access to external capital (Commonwealth Secretariat 2000). They will have to develop their own solutions to these problems and limitations.

Integrated coastal management is one approach that can be adapted to their needs. This will involve adding an intersectoral, horizontal dimension, which includes civil society, to what is essentially a top-down sectoral system of management in most islands. This will not be easy given the dichotomy of timescales between the political agenda (4–5 years) and the environmental agenda (decades). Lying at the center of all the ongoing efforts to achieve sustainable development and integrated coastal management is the need for effective and efficient communication.

Cross-References

- [Coastal Climate](#)
- [Coral Reefs](#)
- [Mangroves, Ecology](#)
- [Mangroves, Geomorphology](#)
- [Meteorological Effects on Coasts](#)
- [Mining of Coastal Materials](#)
- [Small Islands](#)
- [Tourism and Coastal Development](#)
- [Tourism, Criteria for Coastal Sites](#)

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Carrying Capacity in Coastal Areas

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The concept of capacity has received considerable attention as a result of increasing anthropogenic pressure in certain natural environments. Much consideration has recently been given to

increases in coastal populations, with the implication that the carrying capacity of the world's coast is finite and such considerations form part of several coastal management initiatives (UNEP 1996).

Johnson and Thomas (1996) argue that present interest in tourism capacity is due to growth in tourism combined with increasing awareness of environmental issues. The concept is particularly important in the coastal zone which is undergoing rapid change as a result of demographic changes and industrialization (see Kay and Alder 1999, p. 21) in the context of global climate and sea-level change. In its broadest sense, carrying capacity refers to the ability of a system to support an activity or feature at a given level. In the coastal zone, these systems can vary greatly in both scale and type, and range from small salt marshes through large beach resorts to entire continental coasts. The activities or features that they support are also varied and include, for example, beach recreation or species abundance. The term "carrying capacity" does not therefore have a single precise definition. Rather, it is a broad term that covers a range of different concepts. These concepts are related by the idea that systems such as beaches have certain limits or thresholds. For example, a maximum number of animals can be grazed on any given dune system. Attempting to determine the actual limits is often problematic and raises some fundamental questions. In the case of dune grazing, various criteria could be used to define the carrying capacity. This could involve assessing the effects of grazing on, for example, the physical integrity of the site, its ecological status, or its recreational value. In practice, these features may be interrelated.

The situation is further complicated by the subjective nature of certain limits. For example, the point at which the aesthetic impact of grazing becomes unacceptable is difficult to define and may vary from one location or cultural setting to another. In recognition of the diverse nature of carrying capacity as a concept, a variety of types of carrying capacity have been identified. Most of these fall into the following categories: physical, ecological, social, and economic.

Physical carrying capacity: This is a measure of the spatial limitations of an area and is often expressed as the number of units that an area can physically accommodate, for example, the number of berths in a marina. Determining the physical capacity for certain activities can, however, become problematic when subjective elements are introduced. For example, the maximum number of people that can safely swim in a bay depends on human perceptions and tolerance of risk.

Ecological carrying capacity: At its simplest, this is a measure of the population that an ecosystem can sustain, defined by the population density beyond which the mortality rate for the species becomes greater than the birth rate. The approach is widely adopted in fisheries science (e.g., Busby

et al. 1996). In practice, species interactions are complex and the birth and mortality rates can balance over a range of population densities. In a recreational context, ecological carrying capacity can also be defined as the stress that an ecosystem can withstand, in terms of changing visitor numbers or activities, before its ecological value is unacceptably affected. This approach raises the difficult question of defining ecological value and what constitutes an unacceptable change in it.

Social carrying capacity: This is essentially a measure of crowding tolerance. It has been defined as "... the maximum visitor density at which recreationists still feel comfortable and uncrowded" (De Ruyck et al. 1997, p. 822). In the absence of additional changes, beyond this density visitor numbers start to decline. The social carrying capacity can, however, be influenced by factors such as the recreational infrastructure, visitor attitudes, and sociocultural norms.

Economic carrying capacity: This seeks to define the extent to which an area can be altered before the economic activities that occur in the area are affected adversely. It therefore attempts to measure changes in economic terms (Rees 1992). Examples from the coastal zone might include examining the effect of increased numbers of trailer parks on agricultural activity in dune systems.

In addition to these single discipline assessments, there are a number of composite measures such as recreational and tourist carrying capacity. These attempt to define the threshold of an area for tourism or recreation by combining a range of indicators (Sowman 1987). The actual carrying capacity of a coastal area assessed according to any of the above approaches depends largely on the nature of the area. Carter (1989, p. 357) noted that "Coastal environments vary considerably in their ability to absorb anthropogenic pressure. The carrying capacity of dune grassland is many orders of magnitude below that of rock cliffs." While this may be true, at least in some views of carrying capacity, it should be borne in mind that capacities are not necessarily fixed in time. They can often be altered by management practices for example, the provision of recreational facilities can increase the social carrying capacity of an area. They can also alter in response to wider environmental changes, whereby a change in mean sea temperature could affect the ecological carrying capacity of an area for a range of species, or a shift in social attitudes could alter what was considered acceptable degradation. As Arrow et al. (1995, p. 520) have noted: "Carrying capacities in nature are not fixed, static or simple relations. They are contingent on technology, preferences, and the structure of production and consumption. They are also contingent on the ever-changing state of interactions between the physical and biotic environment."

Cross-References

- Coastal Management Practices
- Economics of Beaches
- Environmental Quality
- Human Impact on Coasts
- Tourism and Coastal Development
- Tourism, Criteria for Coastal Sites

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Cave and Karst in Coastal Settings

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Definition

Coastal cliffs and terraces comprise a predominant portion of the world's coastlines (Emery and Kuhn 1982). Caves and related geologic features are common components of these landscapes. Karst and pseudokarst processes have repeatedly shaped these complex landscapes over time, leaving physical clues of their mechanisms, such as caves, within a wide range of coastal rock types.

Introduction

The geomorphology of coastal landscapes can provide a wealth of information to coastal engineers, archaeologists, biologists, geoscientists, and resource managers. Modeling the evolution of coastal karst landforms requires correctly identifying these physical clues in order to dissect and reconstruct the geologic record illustrating such processes. Reconstructing complex coastal geomorphologies can further support multidisciplinary studies of associated biodiversity, climate change, and past, present, and future human uses of coastal landscapes.

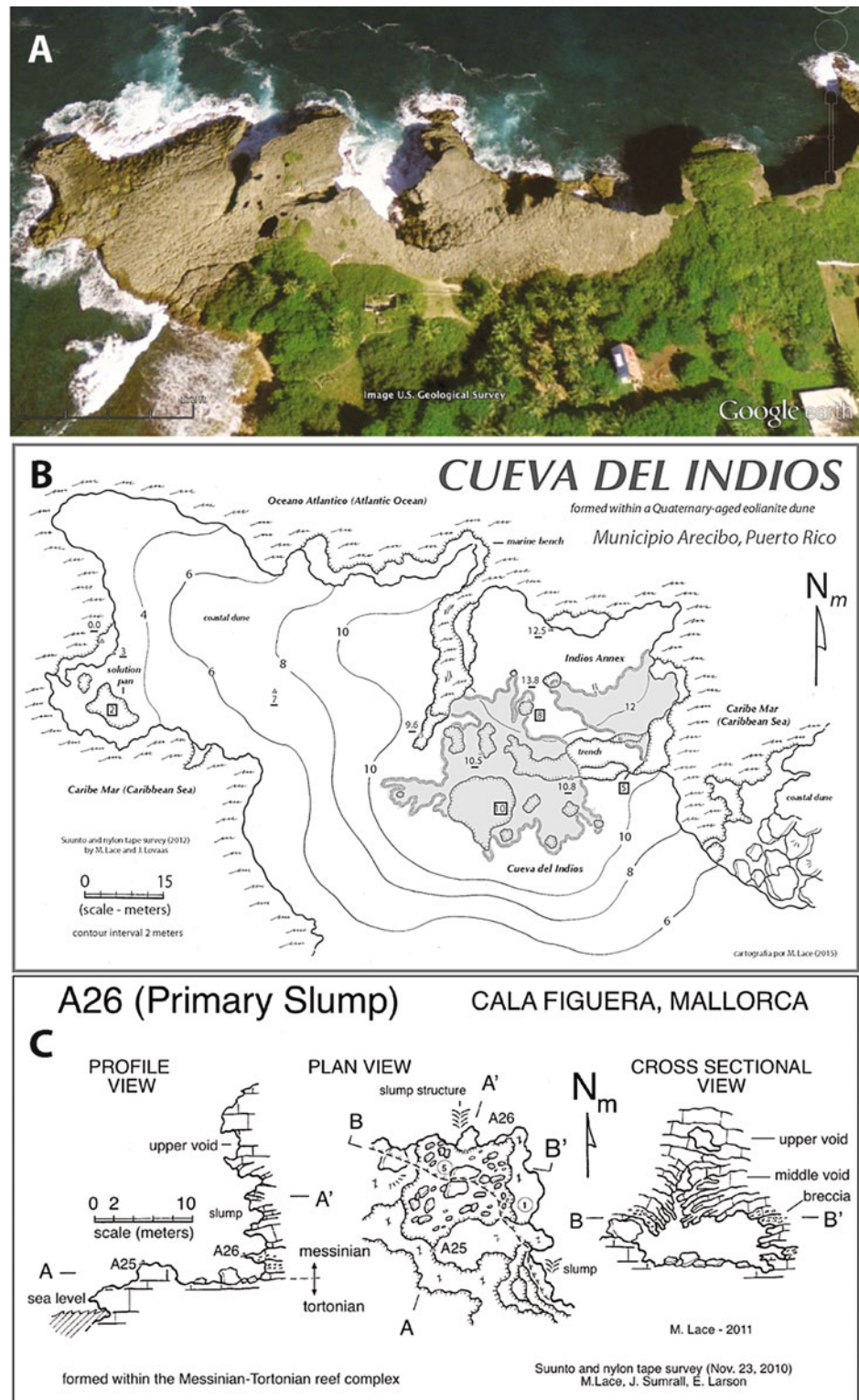
A wide range of tools are available to define cave and karst topographies, including aerial and terrestrial LiDAR, satellite imagery, remote sensing techniques, and the application of basic geological mapping techniques to define coastal structures. Detailed cave maps are critical baseline components in such an analysis, effectively integrating multiple data sets into a single representation defining site geomorphology. Figure 1a illustrates a segment of fossilized dune ridge composed of eolian calcarenite that has been heavily denuded by surface erosion and lateral littoral breach of its seaward flank. Detailed site surveys revealed littoral modification of the ridge and evidence of erosion intersecting a preexisting void (Fig. 1b). Such examples are commonly encountered in carbonate coasts shaped by multiple speleogenetic processes and cave maps can serve as powerful interpretive tools. Integrating surficial and subsurface landscape features as a function of associated lithologies and their relationship to past and present sea levels can form a more comprehensive model of coastal karst landform development (Fig. 1c).

Karst Versus Pseudokarst Development

Rock surfaces within coastal margins can exhibit an astonishing array of structural features generally known as “coastal karren” which are shaped by freshwater and seawater chemistry, bioerosional and bioconstructional processes. The action of multiple karst or pseudokarst mechanisms on carbonate coasts can generate distinctive surficial morphologies, particularly in tropical environments (Taboroši and Kázmér 2013). Subsurface morphologies, including caves, formed within the same environments are equally complex.

Correctly identifying karst features also requires a clear understanding of pseudokarst mechanisms (i.e., features formed by non-dissolutional processes). While karstic processes create dissolutional voids with distinctive morphologies, littoral erosion processes, often acting on the same environment, similarly form their own distinctive structures. Sea caves, for example, are an abundant

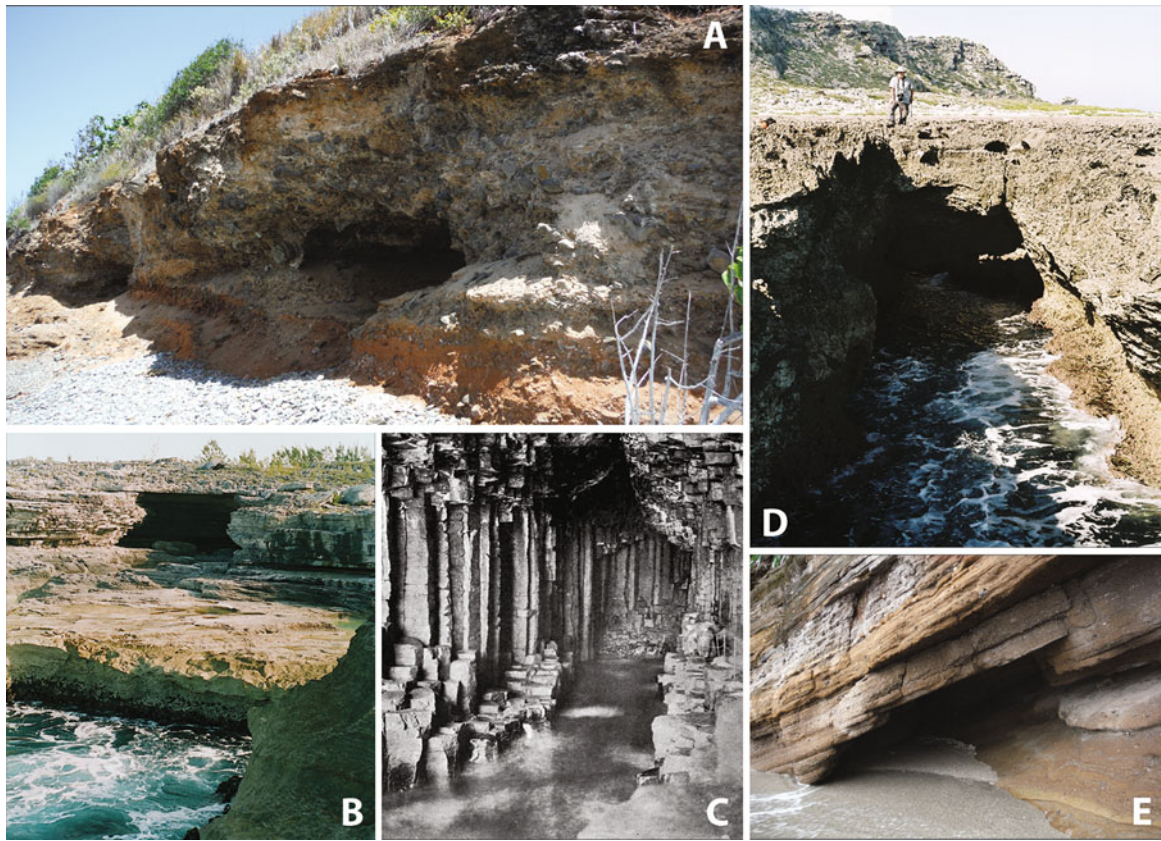
Cave and Karst in Coastal Settings, Fig. 1 Mapping coastal karst. (a) Satellite imagery of coastal calcarenous eolianite (Puerto Rico). (b) The extent of cave development within the same lithified coastal dune illustrated by aerial footprint via applied cave surveying. (c) Cartographic example of recording episodic cave development as a function of past sea levels, shoreline modification, and associated lithologies



cave type found in coastal margins worldwide, but the range of cave types occurring in modern or relict shorelines is diverse.

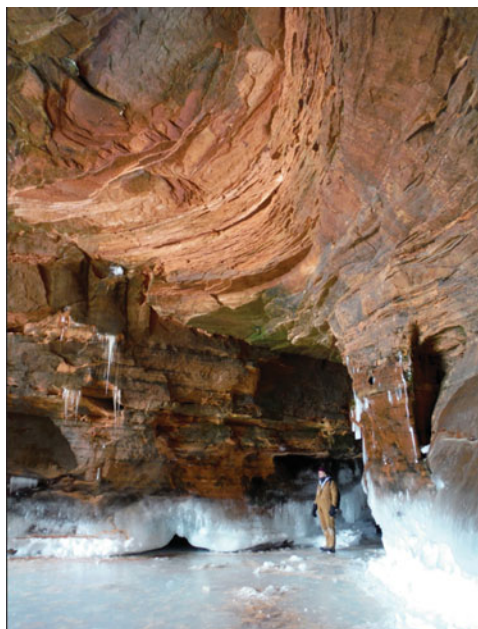
Modern coastal geomorphology includes detailed examination of both karst and pseudokarst caves and features with

equal consideration in terms of their role in coastal landscape instabilities, biodiversity profiles as well, and modern cultural significance. Such an approach avoids theoretical or exploration biases that could compromise an effective landscape evolution model.

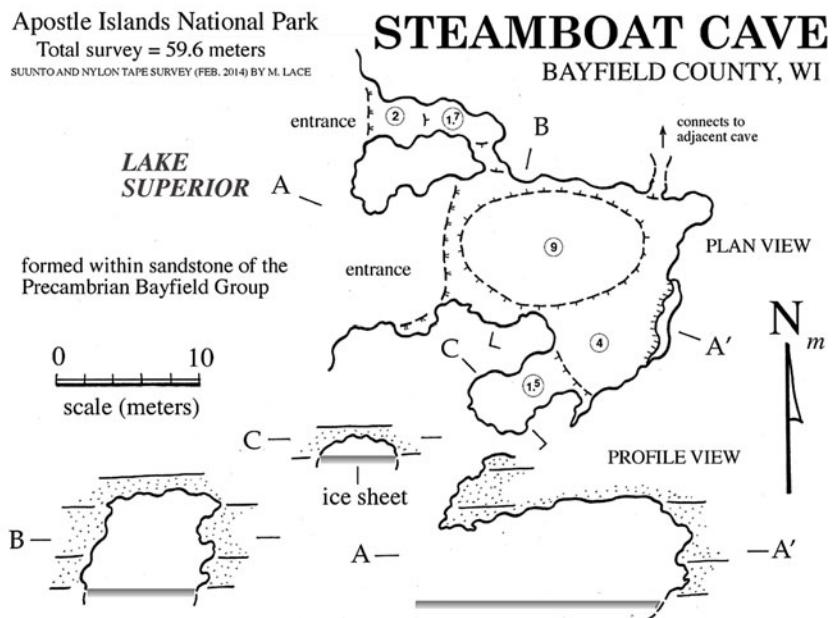


Cave and Karst in Coastal Settings, Fig. 3 Sea cave morphologies. (a) Sea cave formed within a complex sequence of volcanic tuff and conglomerate (St. Croix, USVI). (b) Sea cave formed within eogenetic eolian carbonate (Eleuthera, Bahamas). (c) Sea cave formed within

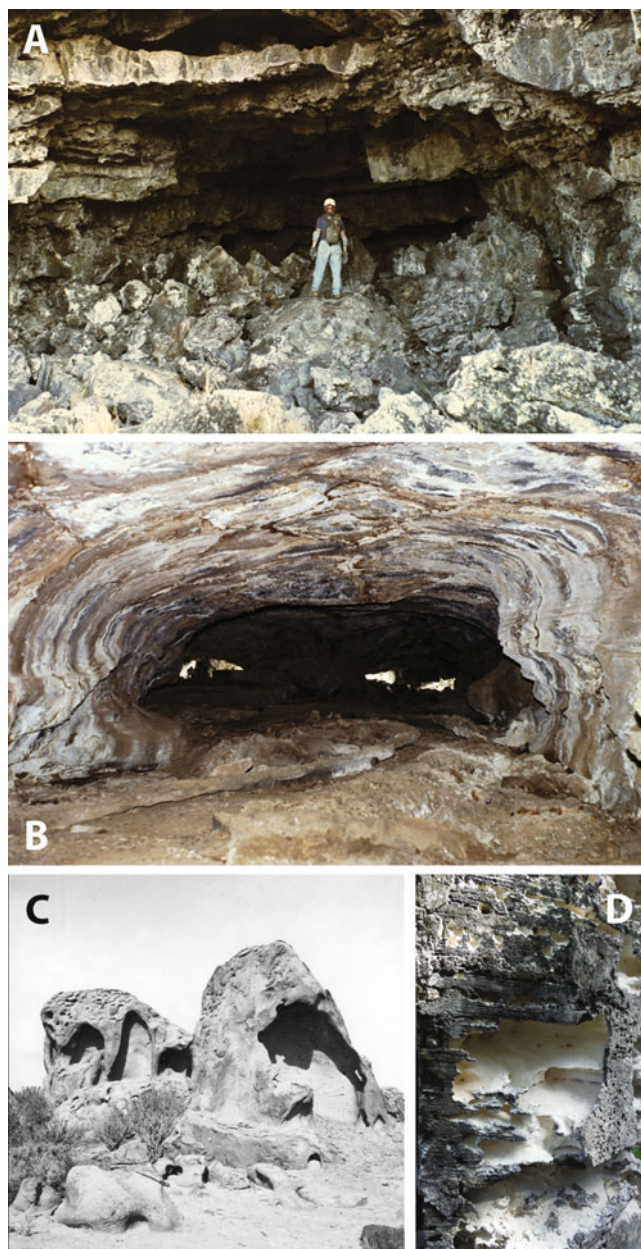
columnar basalt (Fingal's Cave, Staffa Island, Scotland) (Johnson 1919). (d) Sea cave formed along a fracture within a Pleistocene-aged marine terrace (Haiti). (e) Sea cave formed by wave energy acting on a discontinuity within reworked volcanic sediments (Grenada)



Apostle Islands National Park
Total survey = 59.6 meters
SUUNTO AND NYLON TAPE SURVEY (FEB. 2014) BY M. LACE



Cave and Karst in Coastal Settings, Fig. 4 (a) Littoral cave development in a freshwater setting. (b) Map of littoral cave formed within Precambrian sandstone (Lake Superior, Wisconsin, USA)



Cave and Karst in Coastal Settings, Fig. 5 (a) Lava tube development. (b) Characteristic lava tube passage morphology (Hawaii'i, USA). (c) Tafoni formed in granitic blocks (Caldera, Chile), United States Geological Survey archives. (d) Tafoni development in an exterior limestone wall of a colonial period building (Middle Caicos)

Karst caves include fluvial caves (i.e., stream caves), pit caves, and flank margin caves, with the latter representing one of the most abundant cave types observed in coastal carbonate settings worldwide (Mylroie 2013). The term “flank margin cave” refers to voids formed along the *flank* of a landmass by dissolution activity associated with the

distal *margins* of a freshwater lens. The unique structure, laminar flow, and chemistry associated with freshwater lenses can generate complex maze-like structures (Fig. 2a) displaying characteristic cusped (or ramiform) morphologies (Fig. 2b) that are readily distinguishable from caves formed by other mechanisms. Coastal processes can progressively degrade such cave structures to the point where only vague remnants remain. Many natural bridges found on carbonate coasts are actually relict flank margin caves that have been reduced to simple cross section of a once larger complex subsurface void (Fig. 2c).

Pseudokarst caves include sea caves, lava tubes, constructional caves, mechanical fractures, talus caves, and tafoni (Mylroie and Mylroie 2013).

Sea caves (or littoral caves) are common in nearly all carbonate or non-carbonate coastal rock types and have long served as iconic representations of coastal cave development (Fig. 3). The action of wave energy on structural defects, discontinuities, or bedding planes within a cliff face can produce an array of cave morphologies (Fig. 3b–e). Such erosion can reveal complex coastal stratigraphies, including various volcanic tuffs and conglomerates (Fig. 3a). Coastal basalts can also support sea cave development such as the iconic Fingal's Cave (Island of Staffa), which is formed within columnar basalt (Fig. 3c). Similar cave structures have also been noted in continental freshwater settings, such as the active shoreline perimeter and islands of Lake Superior (Fig. 4).

Constructional caves are formed by the accumulation of host rock material (e.g., tufa or coral) instead of the removal of material to create a void. Tufa caves occur in both continental and coastal settings worldwide, with mineral deposition rates shaped by the hydrogeology unique to each setting. Coralloid caves are cavities formed by the progressive accumulation of reef materials. Such voids are shaped by complex reefal ecosystems and in turn can serve as unique habitats for marine cryptofauna. Talus caves are voids formed as a result of host rock displacement, for example, when a segment of coastal cliff breaks away and forms a cavity beneath it when it comes to rest.

Lava tubes are also considered pseudokarst caves and are found within many coastlines and interior regions of volcanic origin. The extensive lava tube systems of the Hawaiian Islands display unique passage morphologies shaped by the associated flow dynamics (Fig. 5a and b.). Tafoni are cavities formed predominantly by salt air weathering (Owen 2013). They occur in numerous carbonate and non-carbonate rock types (Fig. 5c). A tafone (singular) or tafoni (plural) can form at a comparatively rapid rate, generating voids on millimeter to multimeter scales. They are not uncommon in modern roadcuts and



Cave and Karst in Coastal Settings, Fig. 6 Marine pedestals. (a) Pleistocene reef remnant within an active intertidal zone (Puerto Rico). (b) Pedestal on an uplifted Pleistocene terrace (Barbados). (c) Karstic

remnant formed on eolian calcarenite coast in the absence of uplift (Eleuthera Island, Bahamas)

on historical building exteriors composed of soft carbonates, potentially contributing to structural destabilization of such structures and complicating long-term preservation strategies (Fig. 5d).

To further complicate the task of defining coastal landscapes, caves often exhibit signs of multiple processes having influenced their eventual morphology. Caves initially formed by one process can be completely overprinted by secondary processes. Thus, identifying the true geologic origin of an individual cave or a larger landscape structure can prove problematic.

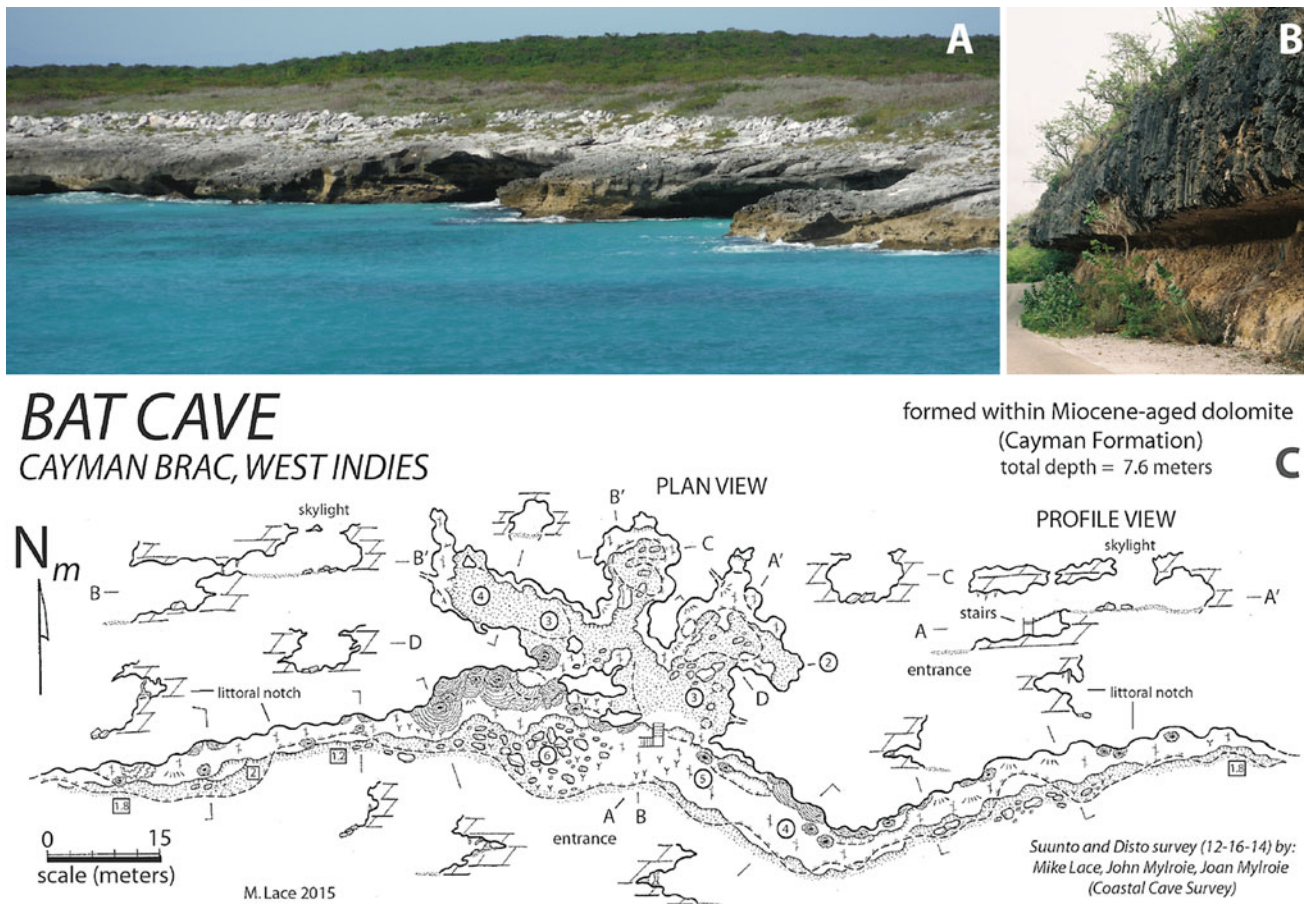
Sea Levels and Cave Development

Coastal features can effectively be used to identify past and present sea levels, such as mangrove development, marine terraces and pedestals, beachrock deposits, or littoral notches (Myroie and Myroie 2013). Marine pedestals, for example, occur in a wide range of carbonate shorelines, including modern intertidal zones (Fig. 6a), paleoshorelines subjected

to tectonic uplift (Fig. 6b) or in comparatively tectonically stable settings (Fig. 6c). Such features can be used to identify shoreline structures and in some cases be used to approximate associated surface lowering rates based on the pedestal height.

Caves also serve as comparatively durable physical records of past sea levels compared to other littoral markers as their remnant structures are often visible even when surface erosion or cliff retreat has profoundly altered associated coastlines influenced by tectonic uplift and/or glacioeustatic sea level changes.

Sea caves are an obvious example as they are formed by erosion of a vulnerable segment of a coastal escarpment, as are storm surge deposits (Fig. 7a). Littoral notches can serve a similar function as they are similarly formed by bioerosion and/or the result of wave energy focused on a confined lateral segment of intertidal cliff exposures (Fig. 7b). Figure 7c illustrates an extensive littoral notch formed along the base of an uplifted dolomite paleoshoreline escarpment. As the notch progressively advanced laterally into the cliff face, it breached a series of preexisting voids (flank margin caves).



Cave and Karst in Coastal Settings, Fig. 7 (a) Coastal eolianite shoreline exhibiting cave development in the modern intertidal zone overlain with surge deposited rubble (Great Abaco, Bahamas). (b) Littoral notch within uplifted Pleistocene terrace (Bonaire). (c) Map

illustrating extensive littoral notch formed along a dolomitic cliffline that has intersected a preexisting cave void (Cayman Brac, British West Indies)

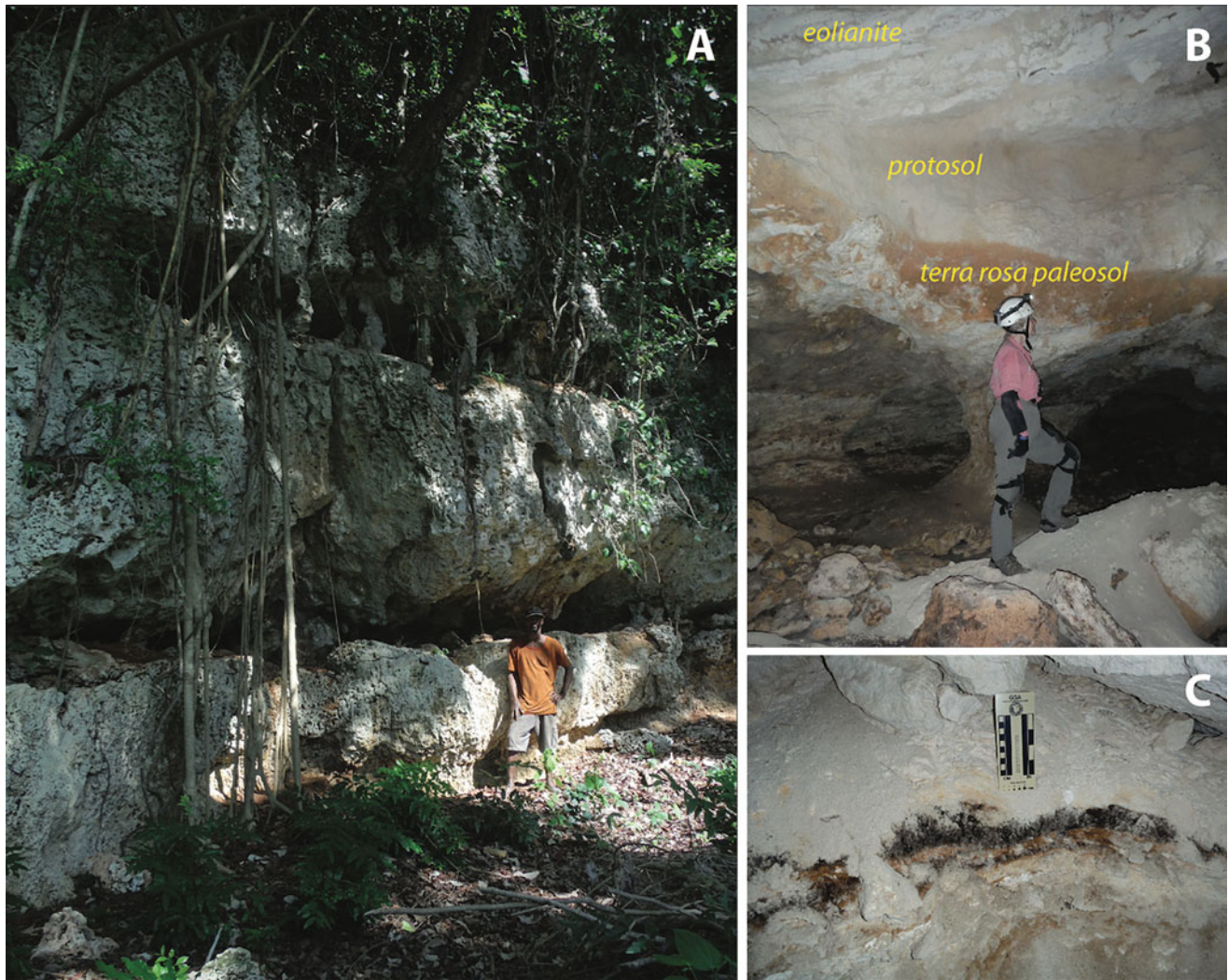
Flank margin caves also serve as useful sea level markers as they form by dissolution process dependent on a freshwater lens supported by stable sea levels at different periods. Similar to the scenario in Fig. 7c, Fig. 8a shows a Pleistocene-aged coastal escarpment displaying a record of multiple sea-level driven events, including the breach and denudation of pre-existing voids (Fig. 8a). However, some cave types may mimic types associated with past sea levels, resulting in an incorrect interpretation. Tafoni, for example, are cavities formed by salt air weathering and not by direct contact with sea in an active intertidal zone (Owen 2013). They may appear to be sea caves if one is not familiar with their subtle yet distinctive morphologies.

The very structure of caves can reveal a wealth of information not readily apparent from examining the associated surface geomorphology. Coastal karst systems, for example, can harbor clear records of past sea level conditions through careful examination of cave stratigraphy (Fig. 8b).

Similarly, sediments preserved in the cave environment can offer complex sequential records of past processes and climates as well as providing a preservational matrix harboring associated paleobiodiversity and archaeological materials (Fig. 8c) (Sasowsky and Mylroie 2007; van Hengstum et al. 2011).

Hydrology and Biodiversity of Coastal Karst

Caves and karst features serve as integral components of complex hydrological systems both in interior and coastal landscapes (Ford and Williams 2013). Many continental areas of the world rely on potable water extracted from complex karst aquifers. Similarly, populations in coastal zones often depend on freshwater lenses of varying depths and stabilities (Fratesi 2013). Many coastal and interior karst settings are hydrologically linked, forming unique habitats



Cave and Karst in Coastal Settings, Fig. 8 (a) Multiple past sea-level indicators preserved in a karstic, Pleistocene-aged cliff face (St. Helene, Bay Islands, Honduras). (b) Cave stratigraphy (Stepwell Cave,

Abaco Island Bahamas). Note exfoliated protosol at subject's feet. (c) Complex cave sediment stratigraphy (Isla de Mona, Puerto Rico)

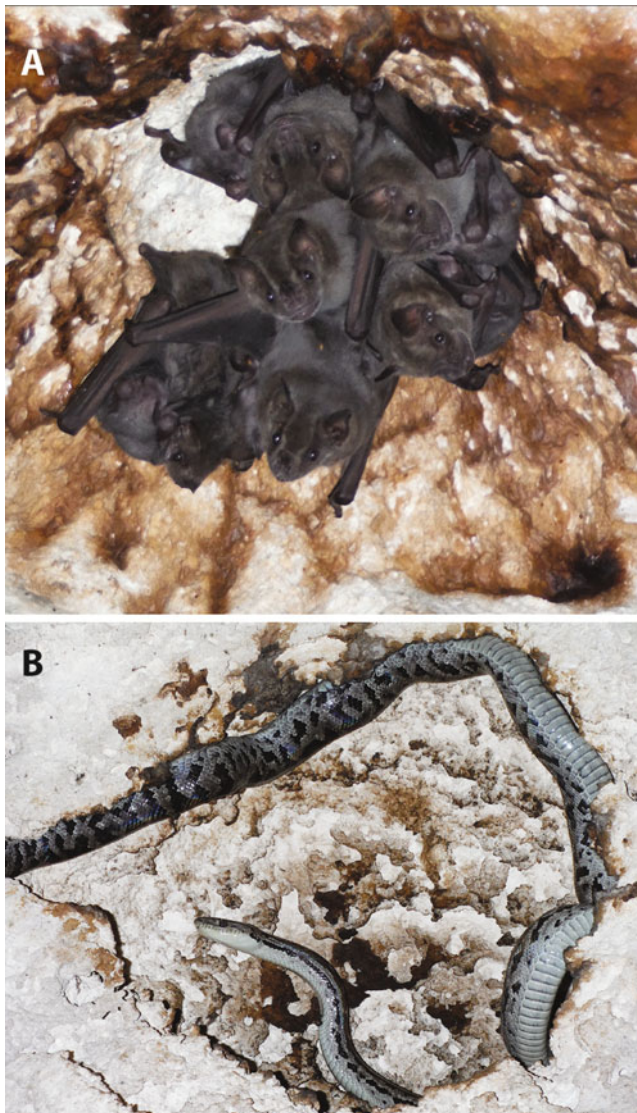
with equally unique vulnerabilities, such as karst in estuarine environments (Myroie et al. 2015). Though the hydrology of these settings can be influenced by varied mechanisms, they commonly share vulnerabilities to contamination due to land use patterns. Understanding the hydrology associated with littoral zones, particularly in karstic regions, can support more effective water resource management and water quality.

Cave biosystems are known to be complex and diverse as they reflect the inherent biodiversity specific to littoral landscapes on multiple scales (Fig. 9). Coastal areas in particular harbor cave environments shaped by the transition from interdependent terrestrial to marine habitats, as reflected in modern versus fossil records within cave sediments within extensive blue hole systems of the Bahamas (Lace and

Myroie 2013a; Steadman et al. 2007). Anchialine environments, for example, are littoral cave systems flooded by the sea and often harbor unique species (Culver and Pipan 2009). Cave fauna generally encompasses a wide range of species ranging from those adapted to the underground environment (i.e., troglobitic) to those making opportunistic use of caves (i.e., troglolytic).

Coastal Karst Landforms as Cultural Landscapes

Caves can preserve complex past records of fauna and flora but can serve as effective repositories of traces of human activities as defined by coastal archaeology. Caves have long been used as complex ritual spaces where cultural



Cave and Karst in Coastal Settings, Fig. 9 (a) Roosting bats in a Puerto Rico cave. (b) Bahamian boa perched in a cave ceiling pocket while hunting bats (Great Exuma, Bahamas)

traditions have been expressed for millennia, influencing our collective understanding of human migration and settlement patterns over time (Lace and Mylroie 2013a). Archaeological records preserved in caves include human remains from mortuary uses, petroglyphs and pictographs from ceremonial uses of such underground spaces, and complex cultural deposits that can be composed of ceramics, lithics, textiles, and/or faunal remains. Yet, there remain significant gaps in our understanding of the complex relationships between coastal landscapes and human activities over time. Current research efforts in many insular and continental margins are focused on filling in these blanks, building more comprehensive comparative models of cultural progression as a function of changing coastal geomorphologies.

Cave and Karst Resource Management in Littoral Settings

Coastline changes are the result of a dynamic interplay of natural and anthropogenic influences that are intimately linked to geological processes and climate change in coastal settings (Harff et al. 2007). Consequently, many carbonate coastal zones can benefit from detailed karst resource inventories and stability assessments in the face of land use pressures, such as expanding residential and commercial development on limited coastal margins (Fig. 10a) or potential threats to coastal water quality (Fig. 10b).

Coastlines understandably draw a range of human activities, including settlement, commerce, and tourism. Natural bridges, for example, often result from progressive cliff retreat and wave-induced erosion of preexisting voids or faults within carbonate or non-carbonate coastal escarpments. Some of these structures have been successfully developed as coastal tourism destinations, yet they can be structurally ephemeral components of dynamic coastal landscapes, as shown by the collapse of the Azur Window on the Maltese coast (2017) and the natural bridge of Aruba (2007).

Sustainable resource management and visitor safety protocols in such case can benefit from a clearer understanding of the geomorphologic origins of subsurface voids, respective lithologies, and structural integrities specific to karst and pseudokarst landscapes (Lace and Mylroie 2013b; Gutiérrez et al. 2014). For example, sea cave development imparts a predictable destabilization of a coastal escarpment as a function of visible progressive littoral erosion and associated cliff retreat. Alternatively, flank margin are voids that form within carbonate coasts without entrances. As such, a coastal area may harbor multiple destabilizing voids – some visible once they are breached but others potentially unknown if they remain intact within the host rock.

Summary

The expanding field research into coastal karst and the resulting databases support emerging models of cave development, cave biodiversity, and archaeological, historical, and modern human use patterns specific to karst landscapes. These studies rely on a critical common denominator – a clearer understanding of the comparative coastal geomorphologies unique to these settings and their role in the management and preservation of karst on site-specific, regional, and global scales.

Cave and Karst in Coastal Settings, Fig. 10 Coastal karst management. (a) Dense residential development on coastal karst (Providenciales, Turks and Caicos) and (b) Kitchen Sink Pit (Acklins Island, Bahamas)



Cross-References

- [Archaeology](#)
- [Changing Sea Levels](#)
- [Developed Coasts](#)
- [Marine Terraces](#)

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Cays

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Definition

A coral cay is a reef island formed from biogenic sediments derived from the adjacent reef. Cays form where sediments produced on the reef are transported by tidal currents and refracted waves to a focal point on the reef top where they accumulate. Although cays can be large features, many square kilometers in area, most are relatively small and low-lying, rarely rising more than 3.0–4.0 m above sea level. Most cays are dominated by sand and/or rubble or shingle-sized sediments. Cays begin as subtidal or intertidal banks, but they may later accrete above sea level. Cays are initially unvegetated and unconsolidated, but in time they may develop soils, acquire a vegetation cover, and become partially lithified (not necessarily in that order).

Because cays are formed of largely unconsolidated sediments, they may remain mobile and unstable, particularly where the focal point of refracted waves and sediment deposition is sensitive to small changes in wave direction, either due to variability in the wind/wave climate or because the position of the focal point shifts significantly with small changes in wave direction due to the shape of the reef. Cay shape is strongly controlled by the shape of the host reef platform and its influence on wave refraction; oval-shaped cays commonly form on oval-shaped reefs, and linear cays typically develop on more elongate reef platforms. Reef size appears to be less influential; small reefs of less than a hectare support cays that almost fully occupy the reef top whereas many larger reefs either lack cays or support only very ones. The critical requirements for cay formation appear to be the availability of transportable sediments, a reef morphology and wave regime that combine to define a reasonably stable depocenter on the reef top, and the availability of adequate hydrodynamic energy to transport sediments to a position where they can accumulate.

Cay Sediments

The size, shape, and specific gravity of biogenic sediments available for cay construction are strongly influenced by the skeletal structure of dominant calcium carbonate producing organisms on the surrounding reef (Perry et al. 2011). Coarser sediments (gravels and above) tend to be dominated by coral fragments typically derived from branching corals such as *Acropora* sp., although locally constituents such as mollusk shells may predominate. The transport of coral rubble across the reef top requires considerable energy such that coral rubble banks parallel and relatively close to (usually within a 100 m) the windward reef edge are common features on many reef tops. These banks of coarse coral clasts are often associated with tropical storms, with higher-energy events stripping material from windward reef slope and delivering it to the reef top (Ryan et al. 2016). Because these coarse deposits remain relatively stable between storm events, they may become cemented and can eventually support vegetation, albeit generally only the most resilient of species.

Because biogenic sands can be transported further across the reef platform, sandy cays tend to be located towards the leeward margins of reef platforms. Sands between 0 and 1.5 ϕ tend to dominate the size fractions, but size alone is not a reliable indicator of the hydraulic behavior and transportability of biogenic sediments produced by different organisms as variations in shape and density are also important. Organisms that produce biogenic sands important to cay construction include corals, coralline algae, *Halimeda*, bivalve and gastropod mollusks, and a range of foraminiferans. For example, the foraminiferan *Baculogypsina sphaerulata* contributes around half the sand on the Raine Island on the northern Great Barrier Reef (Dawson et al. 2014), and together with two other species (*Amphistegina lessonii* and *Calcarina hispida*) comprises more than 30% of the sands at Green Island near Cairns (Yamano et al. 2000). Many cays in the Maldives are dominated by *Halimeda* leaflets (Kench et al. 2005). Coral is often a significant component, but on many sand cays, the dominant biogenic sediments are derived from organisms that produce skeletal material at sand size (such as many foraminiferans), or from skeletal material which reduces to sand size relatively quickly (such as *Halimeda*). In contrast, it may take centuries to millennia to produce sands from larger coral skeletons (Perry et al. 2011). The durability of the sediment is also critical, especially where sediments must be transported considerable distances across reef flats from zones of production to the cay shoreline, or where the cay shoreline is dynamic and movements of the shoreline can abrade beach sands leading to high attrition.

Differences in sediment composition between cays, and differences in sediment composition on a cay through time, are strongly influenced by ecological conditions and the resultant composition of carbonate producing communities

living on a reef flat. For example, cays dominated by foraminiferans generally occur on reefs with a shallow outer algal rim covered with algal turf rim that becomes exposed at low tides. Where cays were initially dominated by coral sands or other biogenic clasts later shift to foraminiferans the change is generally interpreted as resulting from the development of the shallow algal rim which is the preferred habitat for key sand-producing foraminiferans, either as a consequence of the reef accreting to sea level or in some cases as a result of late Holocene sea level fall (Yamano et al. 2000).

Hydro-Geology and Lithification

Cay hydrology was for a long time explained using the simple Ghyben-Herzberg model in which a freshwater lens derived from rainwater “floats” over denser seawater, with little mixing and the assumption that the cay is formed of homogeneous materials. This model requires that a cay is large enough to capture adequate rainfall recharge relative to losses and shows that a freshwater lens may rise to a meter or more above and extend several meters below sea level on larger islands. The simplest scenario of this model depicts recharge at the island center with a general outward flow towards the shoreline. Tidal fluctuations of the water table and mixing of freshwater and seawater are not considered. This model is now generally recognized as over-simplistic, with layered aquifer models more commonly applied (Oberdorfer and Buddemeier 1988). These models more realistically accommodate the differences in permeability between the unconsolidated Holocene sedimentary and reef units that typically form the upper stratigraphy of reef islands and the underlying often highly karstified and porous Pleistocene units with high permeability which often lie beneath. Layered-aquifer models, such as the two-layer (dual-aquifer) model widely applied in investigations of cay groundwater resources (e.g., Griggs and Petersen 1993; Bailey et al. 2009) yield outputs that accord well with field data and show that tidal flow through the Pleistocene foundations has a significant influence on cay hydrology.

Various processes can cement and lithify cay sediments and play an important role in stabilizing cays. Intertidal outcrops of “beach rock” are common on many cays, varying from small discontinuous exposures to extensive features extending several hundred meters alongshore and tens of meters across. These outcrops form where beach sediments are cemented by calcium carbonate (both aragonite and calcite minerals in various crystalline forms) precipitated in the space between sediment particles, consolidating the beach and preserving sedimentary structures and stratigraphy. Beach rock formation has been the subject of long debate, with early workers favoring the hypothesis that evaporation of groundwater supersaturated in calcium carbonate from the

beach face at low tides produced beach rock cements, although it was quickly noted that islands too small or too arid to support groundwater also developed beach rocks thus requiring alternative explanation(s). It is now accepted that beach rocks can form in several ways, including as a result of groundwater and seawater mixing, assisted by biological activity, or where the partial pressure of CO₂ dissolved in sea or groundwater exiting the beach face at low tide as outflow or evaporation is reduced, increasing the pH and encouraging the precipitation of cements in the spaces between beach sediments (see Hanor 1978; Hopley et al. 2007). Whichever processes are involved, beach rock lithification can vary from weakly cemented and friable to dense and highly indurated. Sediments located within the interiors of many cays may also become cemented, especially where a cay supports a large population of seabirds whose droppings percolated downward to the water table and accumulate as guano. As guano-rich rainwater percolates through the soil toward the water table, it is quite acid (pH 4.5–6), but it is quickly neutralized as it mixes with the carbonate-rich groundwater, where upon the precipitation of phosphate cements and the replacement of some carbonates with phosphates can occur. Island sediments lithified by phosphates are commonly referred to as phosphatic cay sandstones and may contain as much as 85% phosphate as cements or bioclast replacement (Baker et al. 1998). Both beach rock and cay sandstones are important in providing stability to cays.

Vegetation

Vegetation plays an important role in the formation and stabilization of many cays. Once a coral cay extends beyond the limits of high tide, it may become colonized by pioneering vegetation, which may either float to the island or be delivered by birds. Pan-tropical plants dominate many cays. Some cays provide classic examples of vegetation succession, with transitions to more advanced vegetation communities occurring as cay sediments transform to soils. Over time, decaying vegetation adds organic matter to the soil, which together with guano produced by nesting seabirds increases soil fertility. Pioneering species tend to be dominated by robust creepers and vines (e.g., *Ipomoea* sp., *Canavalia* sp.) and salt-tolerant succulents (*Sesuvium* sp.) to be followed by grasses (e.g., *Lepturus repens*, *Sporobolus virginicus*) and then shrubs such as *Heliotropium argenteum*, *Scaevola taccada*, or *Pemphis acidula*. Where rainfall allows, cay soils may support significant communities of larger trees such *Casuarina equisetifolia* on the shoreline or *Pisonia grandis* which develops best as interior forest in locations protected from winds. More advanced stages of succession usually occur toward the island interior because it is generally the oldest and most stable location. However, all stages of

vegetational succession may be found on many cays, because many experience phases of erosion and accretion on different sections of shoreline, possibly at different times, providing a mosaic of vegetation communities of different age and complexity.

Classification

Several different classification schemes have been proposed for coral cays (Steers 1929; Spender 1930; Fairbridge 1950; Stoddart and Steers 1977; Hopley 1982), but they are typically classified on the basis of four key criteria: (a) sediment type – sand, gravel, or a mixture of both; (b) location on reef platform – windward or leeward; (c) shape – linear or compact (oval to round); and (d) vegetation cover (vegetated or unvegetated). Unvegetated sands cays are generally small, low, and can be mobile and ephemeral. Vegetated cays tend to be more stable; the presence of vegetation both demonstrates and improves their stability. Cays with a large proportion of their footprint vegetated are typically more stable than those with a lower vegetation cover to footprint ratio. The ends of linear vegetated cays in particular can often be relatively mobile, with sand spits shifting back and forth seasonally on many linear cays (Flood 1986), especially in monsoonal climates (Kench et al. 2009).

Cays composed of coarser (shingle) sediments develop from rubble banks deposited close to and often parallel the windward edge of some reef platforms. As discussed earlier, these banks are typically deposited by storms, and they can be relatively stable features. Some do become vegetated, but they are more inhospitable and colonization may be slow.

On some reef platforms, both windward cays dominated by coarse sediments and leeward sandy cays can develop, with the intervening reef flat often progressively colonized by mangroves which first establish in the lee of the windward rubble cay. Where these reef islands occur on a single reef top, they have been termed “low wooded islands” (Steers 1929). They are especially well known and described from the inner Great Barrier Reef north of Cairns (Hopley et al. 2007) but have also been described from Indonesia (Tomascik et al. 1997) and the Caribbean (Steers and Lofthouse 1940; Stoddart 1965).

Age and Dynamics

Most cays develop over reef flats at or near to sea level, but the reef flat need not be large, extend entirely across the reef platform, nor be constrained by sea level for cay formation to begin (Hopley et al. 2007). Nonetheless, the requirement that cays form on foundations close to sea level means that the maximum ages of cays varies from region to region in accord

with regional variations in Late Holocene relative sea-level history (Perry et al. 2011). The oldest cays are typically located where modern sea level was approached earliest – in the mid-Holocene, where the underlying reefs grew from shallow substrates and thus got to the sea surface relatively quickly, and where the morphology of antecedent substrate on which the reef grew promoted the development of a reef surface on which sediments could accumulate. Establishing precise chronologies for cay development is difficult as radiocarbon dating establishes when the contributing organism died and not the age of the deposit; these two events can be separated by centuries or even millennia in many settings. Nonetheless many cays on the Great Barrier Reef and elsewhere appear to have established around 4000–3000 years ago (Hopley et al. 2007). It is generally believed that this timing is a function of when many reefs reached sea level and had formed sufficiently large reef flats to host stable cays, although the influence of a small falls in relative sea level during the late Holocene have been suggested by some workers as significant (Dickinson 2001). It is entirely possible that sea level falls have been important for the formation and persistence of some cays; however, the thesis that stable or falling sea levels are necessary for cay formation is not supported by evidence from the Maldives, which demonstrates cay formation can occur before reef flats form and as sea levels continue to rise (Kench et al. 2005). Many cays are, of course, continuing to develop; many presently unvegetated cays may be reasonably expected to evolve into larger and more stable cays with time if conditions of hydrodynamic regime, sediment supply, and sea level allow.

Cays are dynamic landforms that continually adjust their morphology in response to changes in hydrodynamic conditions such as wave and currents, as well as changes in sediment supply. Geomorphologists have long recognized the sensitivity of reef island morphology to such drivers (Perry et al. 2011) and noted that processes such as shoreline erosion, progradation, island accretion, sediment washovers, shoreline re-alignment, and even island migration across a reef platform may all reflect such interactions (Kench 2013). Seasonal oscillations in wind and wave regimes may produce relatively predictable morphological adjustments, with investigations indicating that circular islands in the Maldives were most sensitive to seasonal changes in wave climate (Kench and Brander 2006). Longer-period cycles such as ENSO and Pacific Decadal Oscillation have also been linked to adjustments in cay morphology through their influence on both wave climate and sea levels (Dawson and Smithers 2010). Storm events have been recognized as having a particularly large erosion impact on sandy cays but can be associated with significant accretion on cays constructed of coarser sediments (Bayliss-Smith 1988). Residual beach rock outcrops on reef flats clearly document the existence of sizeable cays that have been entirely washed away (Hopley et al. 2007).

Climate changes associated with fossil fuel use including changes in storm regime, sea surface temperatures, and sea levels have generated considerable anxiety amongst those concerned about the future for cays and the societies and ecosystems that depend upon them. Such changes will impact the morphology directly through movement of cay sediments but will also affect the production and supply of sediments from the surrounding reef (Perry et al. 2011). Studies to date suggest that not all cays will respond in the same way, and that predicting the responses of individual cays can be difficult.

Cross-References

- Atolls
- Beachrock
- Coral Reef Islands
- Coral Reefs
- Hydrology of the Coastal Zone

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Chalk Coasts

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The extraordinary whiteness of chalk has given rise to distinct cliffs visible over large distances. From the Fourth Century BC Greek navigator, Pytheas of Massilia, through Julius Caesar and Shakespeare, to modern writers, chalk coasts were given a unique character. They are most commonly known for their place in Shakespeare’s “King Lear,” the literature of Maupassant and the paintings of Monet and the wartime nostalgic song, “White Cliffs of Dover.” The Romans called England “Albion” because of its white cliffs. The earliest regional study of a chalk coast is Girard’s (1907) description of the cliffs of the English Channel. Chalk cliffs occur, however, more widely than these classic locations. Chalk is a CaCO₃ limestone found in the Upper Cretaceous outcrops on the European coast, in northern Ireland, Yorkshire and Norfolk, along the southern coast of England from the Isle of Thanet in the east to Beer in the west (Steers 1946), along the northern coast of France from Calais to the mouth of

the Seine (Girard 1907; Briquet 1930; Precheur 1960) and on the islands of Møn, Denmark, and Rügen, Germany.

Chalk coasts are distinctive for four reasons: the relative purity of the chalk itself, their distinctive landscape, their often vertical rapidly retreating cliffs and their place in the geomorphological literature as type locations for particular landforms. The chalk contains veins of flint, a biogenic silica, which are typically nodular in shape, vary in size (long axis) from 2 cm to over 200 cm and can weigh more than 8 kg. They are the main source of beach clasts on these coasts and may exist in their original form, break into smaller angular particles (quickly rounded at their edges), or as rounded pebbles. These pebbles are often the result of erosion of re-worked landslides and fluvio-glacial deposits. Chalk cliffs have been important, but not unique, suppliers of flint shingle to such large and important features as the ness, spit, and tombolo structures of southeast England, including Orfordness, Dungeness, Hurst Castle spit, and Chesil beach (May and Hansom 2003), and northern France.

Chalk is notable for its whiteness, its purity, and its friability (Hancock 1976) and is formed mainly of calcareous fossil debris. The white chalk is dominated by two main groups of particle sizes (0.5–4 f.L and 10–100 f.Lm). The lower 25–60 m of the chalk is often dominated by chalk-marl facies in which calcareous-rich chalk alternates with marls containing up to 29% clay (Perrin 1971). The dolomitic chalk outcrops of the Normandy coast, the white chalk of Yorkshire and the White Limestone of Antrim are harder than other chalks because of diagenesis by burial and locally high heat

flows. Typically, these harder chalks retreat less rapidly than the other chalk coastlines. Chalk porosity ranges between 35% and 47% (Hancock 1976), but matrix permeability is exceptionally low. Mass permeability is controlled by joints, fractures, and passages enlarged by solution, all of which are very common in the chalk and have a very important role in the development of some of the more complex chalk coasts.

Cliff forms (Table 1) include very steep slopes, slopes which are strongly faceted, and shallower slopes which are associated with large coastal landslides such as Folkestone Warren (Hutchinson et al. 1980). Many chalk cliffs with low mean retreat rates have small slope facets with mean angles between 30 and 42° that are vegetated by chalk down-land grasses and flowering plants. On these slopes, there may be skeletal calcareous soils, but occasionally red-brown sandy clay deposits (“*clay with flints*”) on the cliff-top are washed down the cliffs and form local pockets of soil.

Some chalk coasts include very intricate cave and inlet and stack and arch features. Caves and inlets usually develop along discontinuities in the rock and give rise to large numbers of narrow inlets. At Flamborough Head, for example, there are more than 55 narrow inlets or geos and around the Isle of Thanet headlands were riddled with inlets and caves, some of which were filled in during coast protection works in the 1960s. The classic erosional cave-arch-stack-stump model found in many school textbooks owes much to the example of differential erosion of the sea along intersecting faults and joints north of Swanage. Stacks are rare features on chalk coasts, usually because the friability of the upper cliffs or the

Chalk Coasts, Table 1 Major categories of chalk coast – England and France

Category	Main features	Cliff height (m) ^a	Typical rates of cliff-top retreat (m a ⁻¹)	Length (km)	
				England	France
I	Simple plan and profile, former usually similar to beach plan form. Slope angles exceed 75°. Extensive platforms.	Less than 31	0.13–1.33	48	56
II	As I, with many small less steep slope Facets. Platforms narrower usually less than 200 m.	More than 31	0.05–0.51	28	15
III	Cliff with basal rocky pedestal, many Discontinuities, simple upper profile, but complex plan including caves, stacks arches. Platforms common but may be submerged.	Less than 31	0.09–0.14	5	3
IV	High cliffs with complex profile. Plan Usually simple, but upper and lower cliffs exceed 75. Middle slope typically at mean angles 35–42 degrees. Boulder fields at cliff foot may affect plan in detail and provide protection for periods of several decades.	More than 31	0.11–0.36	9	22
V	High with extensive and complex Undercliff typically with rotational slides. Upper plan related to mass movements, shoreline plan to debris and boulder fields.	More than 31	0.09–0.36	8	18
VI	Lower steep chalk cliffs with upper concave slope affected by landslides. Simple plan.	Less than 31	0.36	1	10
VII	Steep lower cliff with basal rocky pedestal, many discontinuities. Concave upper profile in overlying sands or clays. Lower cliff has very complex plan including geos, caves, stacks, arches.	Less than 31	0.12–0.95	4	0
Total length of chalk cliffs				103	124

Categories I to VI based on Precheur (1960) and May and Heeps (1985), and category VII based on May in May and Hansom (2003)

^a31 m = critical height for vertical chalk cliffs (Hutchinson 1971)

angles of dip are too great to allow classic pinnacle stacks to survive. In contrast, the English cliffs between Beachy Head and the mouth of the Cuckmere River (the Seven Sisters) are one of the best examples worldwide of a cliffed coast where, despite the local effects of jointing, cliff falls and pocket beaches and the larger effects of cutting across a series of valleys, the coastline retains a remarkably straight form. It is a prime example of the efficiency of marine processes in maintaining erosion and being able to remove the supply of sediment from erosion. The strength of the chalk is sufficient to maintain a vertical face unlike many other comparable materials where slumping is more typical.

The average annual retreat of the cliff-top edge of chalk coasts ranges from about 0.05 m a^{-1} to over 1 m a^{-1} (May 1971; May and Heeps 1985), but many chalk coasts are marked by intermittent cliff changes including some very large falls. Before these falls, the base of the cliff may be undercut in a distinct notch (Hutchinson 1971) or blocks may be quarried from it. The resultant talus may act as a groin as these large falls can take several decades to be removed. Typical recorded losses include cliff-top area in excess of $7,000 \text{ m}^2$ (May 1971), 0.5 million tonnes in volume or linear retreat of over 15 m in a single event. In January 1999 a very large fall at Beachy Head on the southern English coast affected a 200 m stretch of a 90 m-high cliff and produced a debris fan over 5 m in height covering some $50,000 \text{ m}^2$. As cliff height and slide volume increase, a “degree of flow” may develop in chalk debris (Hutchinson 1983). In falls from cliffs between 70 and 150 m in height, a “chalk flow” may carry debris across the shore platforms for distances more than four times the cliff height (Hutchinson 1980, 1983). These flow slides may occur because high pore-water pressures are generated through the crushing impact of relatively weak blocks of high porosity, near-saturated chalk (Hutchinson 1984) or as a result of steady-state flows resulting from rapid or undrained loading of the chalk (Leddra and Jones 1990).

The debris produced by falls varies from clay-sized fines ($0.5\text{--}100 \mu\text{m}$) to boulders of over 1–4 m. Wave action and longshore currents disperse the fines rapidly. Shingle-sized debris is commonly rounded within a few days and is worn down to sand-sized fragments in a few months. The larger material may, however, remain *in situ* for very long periods of time and become colonized by algae and molluscs. The boulders may gradually become smaller, but the platforms often retain a substantial debris cover, with boulder arcs marking the outer edges of the chalk flow debris several decades after the original failure. Intertidal chalk platforms can exceed 1 km in width but are more generally up to about 200 m wide. Their form and development (Wood 1968) has been attributed to the effects of the geological structures, storm wave action (So 1965), the presence of flint strata (which may have a very significant armoring effect: May and Hansom 2003), the tractive effects of beach material (Girard 1907) and the

role of boring organisms (Jehu 1918). Recorded rates of platform lowering are about 25 mm a^{-1} , but Foote et al. (2000) show that micro-erosion rates are very variable and are strongly influenced by platform dwelling organisms. Bioerosion may play a significant role in chalk coast erosion, but further work is needed. Apart from parts of the Sussex, Kent, and Normandy coasts where walls have been constructed to reduce coastal erosion and protect towns, chalk coasts are little modified by the effects of coastal engineering works and so remain one of the better studied natural cliffed coasts.

Cross-References

- [Cliffed Coasts](#)
- [Cliffs, Erosion Rates](#)
- [Cliffs, Lithology versus Erosion Rates](#)
- [Coastline Changes](#)
- [Europe, Coastal Geomorphology](#)
- [Shore Platforms](#)

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Changing Sea Levels

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Synonyms

Eustatic changes; Relative sea-level changes

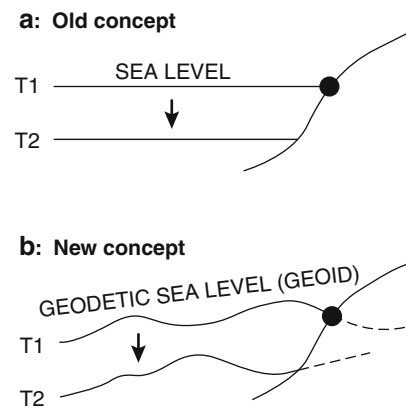
Definition

Changing sea levels refer to variations in the level of the ocean mean sea level regardless of causation mechanism. *Relative sea-level changes* denote the observed changes in sea level as caused by the combination of ocean variables and crustal components. *Eustatic changes* refer to changes in the oceanic level itself (exclusive of crustal and compaction factors). Nowadays, we talk about *regional eustasy*, because of the spatial variability of “eustasy.”

Characteristics

The geodetic sea level – better known as the geoid – is the equipotential surface between the attraction and repelling forces. The actual sea level – known as the dynamic sea surface – closely approximates the geoid surface. The deviations of the dynamic surface from the geoid surface are caused by different oceanographic, hydrographic, and meteorological forces acting upon the distribution of the water masses. Neither the dynamic sea level nor the geoid remains undeformed with time. Sea-level changes refer to changes in mean sea level over time units larger than daily and annual cyclic changes. Longer-term cycles, like the 18.6 and 60 years’ cycles, are measured as sea-level changes.

Our only true benchmarks are the former sea-level positions in the field. The present level of those data points are the combined function both of past changes in sea level and past changes in crustal level. In opposite to all types of crustal deformations, we term all different types of sea-level changes “eustatic” (Mörner 1986). In the old concept of eustasy, all sea-level changes were parallel over the globe (e.g., Fairbridge 1961). This is no longer tenable; sea-level changes can never



Changing Sea Levels, Fig. 1 The old and new concept of “eustasy” (after Mörner 1987) illustrated by the change (arrow) from one sea-level position (black dot and T1) to a new, lower, position (T2). In the old concept (a), any eustatic change in sea level was parallel all over the globe. In the new concept (b), the real sea-level surface is an irregular surface that never remains parallel at a change in sea level

be fully parallel over the globe and are often of compensational, even opposed, character (Mörner 1976). Therefore, we nowadays talk about “regional eustasy” (not global eustasy). The old and new concepts of eustasy are illustrated in Fig. 1.

Water Distribution and Sea-Level Changes

The distribution over the globe of the ocean water masses is defined by the Earth’s rotational ellipsoid, the geoid topography, and various dynamic forces (air pressure, current beat, coastal runoff, etc.). Any changes in any of those parameters will induce sea-level changes on the local, regional, or global dimension.

Sea level can only change within limits determined by its physical causes (e.g., Mörner 2011). The main sea-level variables are listed and quantified in Table 1.

The global water volume is controlled by glacial eustasy, evaporation/precipitation, loss/gain of water, and steric expansion/contraction due to changes in temperature and salinity.

The volume of the ocean basins is controlled by tectonic deformation (ocean floor subsidence, mid-oceanic ridge growth, and coastal tectonics).

The distribution of the water over the globe is controlled by Earth’s rate of rotation, geoid deformation, and various dynamic forces (air pressure, current beat, coastal runoff).

Actual changes in the sea level represent an intricate interaction of multiple factors requiring understanding of various subjects and processes. The dominant controlling factor differs through time and with the timescale considered.

On the million-year timescale, sea level is mainly determined by ocean basin volume, mass distribution, and rate of rotation.

Changing Sea Levels, Table 1 Sea-level variables and quantities (in meters) from Mörrner (2013)

Present tides	
Geoid tides	0.78
Ocean tides	Up to ~18
Present geoid relief	
Maximum difference	180
Present dynamic sea surface	
Low harmonics	Up to max ~2
Major currents	Up to max ~5
Earth's radius differences	
Equator vs pole	21,385
Presently stored glacial volume $\approx$ meters sea level	
Antarctica	~60
Greenland	~6
All mountain glaciers	0.5–0.4
Glacial eustasy	
Last Ice Age amplitude	120–130
Geoid deformation	
Last 150 ka	30–90
Last 20 ka	30–60
Last 8 ka	5–8
Tectono-eustasy (ocean basin volume changes)	
Maximum annual rate	$< \sim 0.00006$
Global glacial isostasy	?
Ocean circulation and rotation beat	
Holocene Gulf stream beat	1.0–0.1
High–low latitude changes (grand solar pulses)	~0.5
Holocene equatorial E–W pulses	~1
Super ENSO events	~1
El Niño–ENSO events	0.3
Thermal expansion	
Open ocean	< 1.0
At 100 m water depth	< 0.035
At 10 m water depth	< 0.0035
At the shore	always ± 0.0
Coastal runoff	
At river outlets	1–2
Evaporation/precipitation	
Indian ocean evaporation low	0.3–0.4
Air pressure	
1 millibar	0.01
Monsoonal range at Bangladesh	0.15
Rotation rate and sea level	
~15 ms rotation (LOD)	1

On the thousand-year timescale, sea level varies mainly due to glacial eustasy and related glacio-isostasy, Earth rotation, and geoid shape as determined by lithospheric mass change.

On the multi-decadal to hundred-year timescale, sea-level changes seem primarily to reflect changes in water–mass distribution related to changes in the oceanic circulation (super ENSO events, PDO, NAO, etc.) caused by variations

in climate, tidal forces and Earth's rate of rotation, and thermal expansion, which differs significantly with water depth (Fig. 2).

On a yearly scale, sea level varies locally because of dynamic factors and El Niño–Southern Oscillation (ENSO) effects.

The Sea-Level Debate

Sea-level changes were recorded and debated from the early eighteenth century (Mörner 1987). The effects of changes in rotation, in ocean basin subsidence, in deflection of the plumb line, and in glaciation were advocated. With the recording of multiple Ice Ages, it became natural to consider cyclic changes in glacial eustatic volume of the oceans (e.g., Daly 1934).

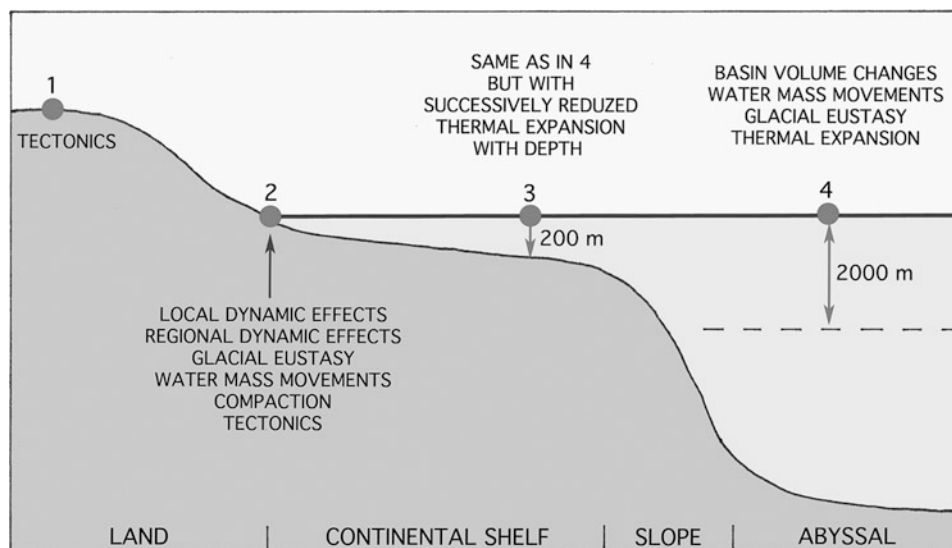
In the 1960s the debate was intensive whether sea level, after the last glaciation maximum, rose with oscillations in mid-Holocene time reaching even above the present level (Fairbridge 1961) or as a smooth line continually rising to its present level (Shepard 1963). The curves of Jelgersma (1961), Scholl (1964), Mörrner (1969), and Tooley (1974) added detailed records from single regions backed up by numerous dates ruling out the occurrence of high-amplitude sea-level oscillations. Assuming that eustatic sea level could be expressed in one globally valid eustatic curve, all differences in observed levels had to be explained in terms of changes in the crustal or sedimentary level. In 1971 Mörrner called attention to the fact that most sea-level curves, despite very large differences in level in mid-Holocene time, converged fairly closely at about 8000 radiocarbon years BP. This seemed to indicate a major cyclic redistribution of the ocean masses in Holocene time. Later, the explanation became obvious; the geoid relief is subjected to deformation (Mörner 1976). Consequently, each region has to define its own *regional eustatic curve* (a global curve became an illusion).

Clark (1980) and Peltier (1998) presented their global glacial isostatic loading models, which called for major sea-level irregularities over the globe.

The “World Atlas of Holocene Sea-Level Changes” (Pirazzoli 1991) gives a good and comprehensive view of the changes in sea level. Besides global differentiation in the eustatic ocean components, the recorded relative sea-level changes are strongly affected by differential tectonics and sedimentary compaction. Despite this complexity, one may distinguish some general regularities: all curves show a general rise up to about 5000–6000 radiocarbon years BP followed by an out of phase to opposed oscillatory trend for the last 5000–6000 years around the present zero level (Mörner 1995). The first part records a general glacial eustatic rise in sea level prior to 5000–6000 BP plus a local to regional differentiation due to tectonics, geoid deformation, and global isostatic compensation. The second part (the last

Changing Sea Levels,

Fig. 2 Variations of main factors affecting relative sea-level changes in a profile from land to the abyssal of the oceans. At point 2, the shore or land/sea interaction, there is a complicated interplay between crustal movements, sediment compaction, and different oceanic variables. The thermal expansion factor is zero at the shore, however. In the open oceans, all the main oceanic variables are actively in operation



5000–6000 years) lacks any significant glacial eustatic component and is instead dominated by the redistribution of water masses around the present mean sea level.

The redistribution of water masses during the last 5000 years seems primarily to be the function of changes in the main oceanic current systems in response to the interchange of angular momentum between the “solid” Earth and the hydrosphere (e.g., Möerner 1995). This includes a multi-decadal E–W swash of oceanic water masses in the Equatorial plane and a N–S interchange of water masses between high and low latitudes during solar maxima and minima (Mörner 2017).

This implies that the Earth was in a totally different mode in the last 5000 years as compared with the previous 15,000 years with different dominant driving forces for the changes in sea level.

To Reconstruct Past Sea-Level Changes

Field observations of past sea-level positions are the only means of reconstructing sea-level changes in pre-instrumental time. The observational object may be a stratigraphic layer, a morphologic feature, or single objects referring to sea-level changes. This object must be fixed in altitude with respect to present sea level (this may range from an absolute level to a relative level). The next step is to interpret the observed field object with respect to the sea level at the time of formation. This implies the study of sea-level relation, sedimentary characteristics, paleoecological inference, paleoenvironmental interpretation, and coastal dynamics. Finally, the object needs proper dating. After all these steps, we have a past sea-level datum or a paleo-sea-level indicator, which can be

used in sea-level curves and sea-level reconstructions. The quality of this datum point depends upon the quality of each individual step in its establishment. The dating of a continuous catching-up coral reef growth may provide an exceptionally continuous record (e.g., Fairbanks 1989).

We have to understand that there is a long chain of uncertainties concerning sea-level changes as they are interpreted by the investigators, the sea-level changes as they are recorded in the field by fragmentary traces, and the sea-level changes as they actually happened in the past.

Reconstructions of past sea-level changes should be based on observational facts in the field. The use of models will always play a subordinate role.

Deep-Sea Records

The deep-sea oxygen isotope records (e.g., Shackleton and Opdyke 1973) have a predominant control from the changes in ice volume and glacial eustatic changes in oceanic water volume. Therefore, they provide a reasonable reconstruction of the quaternary glacial eustatic cycles (Fig. 3). For the last 140,000 years, there seems to be a good agreement between oxygen isotope variations and actual dates of coral reef levels (e.g., Chappell et al. 1996).

Tide-Gauge Records

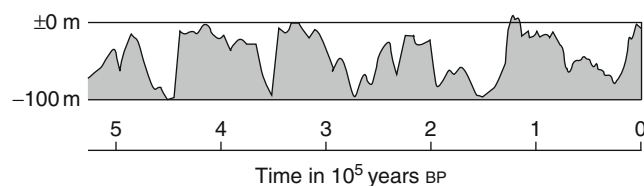
Sea level can be measured by watermarks, mareographs, and tide gauges. The oldest tide gauge is the one in Amsterdam from 1682. From a compilation of global tide-gauge station, Gutenberg (1941) tried to reconstruct the mean global

eustatic component. He arrived at a value of 1.1 mm/year rise in sea level.

Tide gauges record the interplay of changes both in sea level and in land level. Because most tide gauges are located on harbor constructions, often resting on unconsolidated sediments, subsidence due to compaction is a serious problem.

Satellite Altimetry

With the improvement of satellite altimetry (TOPEX/POSEIDON, JASON 1 and 2), it has become possible to record changes in the oceanic sea-level position. This has opened new means of recording sea-level changes. Previously, we were restricted to changes at the sea/land interface (the shoreline). Now we can record the changes over the entire ocean surface covered by satellites. NOAA and the University of Colorado (UC) run databases for the continual recording of mean sea-level rise based on satellite altimetry; while NOAA (2014) gives a mean rise of 2.9 ± 0.4 mm/year, UC (2015) gives 3.3 ± 0.4 mm/year. Both these records have been subjected to “corrections,” criticized by Möerner (2015). If these



Changing Sea Levels, Fig. 3 Main quaternary glacial eustatic cycles as derived from deep-sea oxygen isotope records after Shackleton and Opdyke (1973). A glacial/interglacial sea-level amplitude in the order of 100 m is recorded. It should be noted, however, that improved data sets indicate that the last glaciation maximum was in the order of -120 to 130 m

Changing Sea Levels, Fig. 4

A spectrum of sea-level rates (Mörner 2016a). The five points, further discussed in the text, provide a congruent picture: sea level is globally varying between ± 0.0 and $+1.0$ mm/year (0.5 ± 0.5 mm/year). Only the estimate by the IPCC (2013) is significantly above

“corrections” are removed, the rates become 0.45 and 0.65 mm/year, respectively. This fits well with other means of obtaining actual values of present-day changes in sea level (Fig. 4).

Gravity

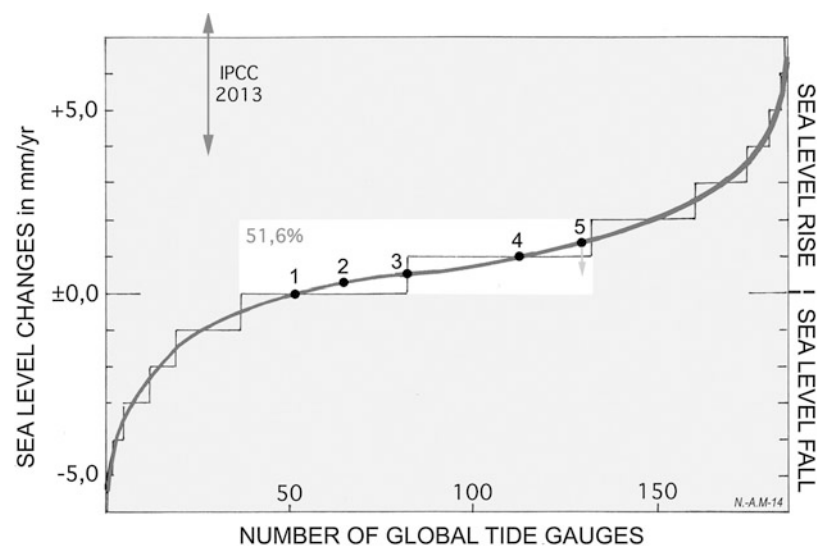
From 2004, there are continual gravity measurements (GRACE), which are used also to estimate the changes in sea level. The measured value is a minute lowering of 0.12 ± 0.06 mm/year.

Present Sea-Level Changes

For the last 5000 years, global sea-level records do not concur with a simple trend but rather with site-specific redistribution of ocean masses. The same applies for the present. Despite this, it seems to be possible to detect a slight general rise in sea level since about AD 1850. No acceleration in this sea-level rise can be recorded in the last decades.

Figure 4 gives a review of the most significant documents of present sea-level changes (Mörner 2016a). There are five main points of information:

1. $+1.14$ mm/year, the mean of 184 tide gauge records scattered all over the globe selected by NOAA (2015) for their global sea-level analyses. This value is too high, however, because many sites used represent subsiding delta sites (Mörner 2015).
2. $+1.0 \pm 0.1$ mm/year, the eustatic component the North Sea, Kattegatt, and Baltic region (Mörner 2015, 2016b).
3. $+0.55 \pm 0.10$ mm/year, the revised satellite altimetry values of Mörner (2015).



4. $+0.25 \pm 0.19$ mm/year, the mean of 170 PSMSL tide-gauge stations having a length of more than 60 years (Parker and Ollier 2015).
5. ± 0.0 mm/year, the value obtained from many global test sites (Mörner 2016b, 2017); the Maldives, Bangladesh, Goa in the Indian Ocean, Tuvalu, Vanuatu, Kiribati, Majuro, Fiji in the Pacific, Surinam–Guyana in NE South America, and Venice in the Mediterranean.

This provides quite a congruent picture of a present global change in sea level varying between ± 0.0 and $+1.0$ mm/year, or 0.5 ± 0.5 mm/year (Mörner 2015, 2016a).

Future Sea-Level Changes

Predictions of sea-level changes to be expected in the future must, by necessity, be based on our understanding of the past and present sea-level trends. Considering the past driving forces, the present trend and rates (Fig. 4), and possible future variability, there are no reasons to assume any disastrous rise in sea level as often claimed (IPCC 2013) but rather a value in the order of a moderate rise of $+5$ cm ± 15 cm in a century (Mörner 2013).

Summary

Sea-level changes are driven by several different factors (Table 1). This calls for a strict application of observationally based facts when reconstructing past, present, and future changes in sea level (Fig. 4). Restriction to models and statistics will never provide data representing the changes in sea level in a meaningful way.

Cross-References

- Coastal Changes, Gradual
- Coastal Dynamics
- El Niño–Southern Oscillation (ENSO)
- Eustasy
- Geodesy
- Late Quaternary Marine Transgression
- Sea-Level Fluctuations Over the Last Millennium
- Tidal Datums
- Tide Gauges

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Cheniers

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Definition and Morphology

Cheniers are relict wave-built coastal plain landforms that occur in sets inland from the shoreline where they originally formed along an active beach as littoral ridges. They are sheltered from daily wave and tidal processes and generally consist of sand, sandy shell, and occasionally sandy gravel. Chenier ridge sequences may develop on prograding deltas or non-deltaic coastal plains. Cheniers form subparallel sets of elongated, narrow ridges (typically up to 3 m high and 40–400 m wide), usually rising slightly above the highest high-tide level. Wave action and storm overwash build the original littoral ridges. The term chenier (Russell and Howe 1935) originated in the French native word (*chêne*) for the majestic evergreen “live oak” tree (*Quercus virginiana*) that lines chenier ridges of southwest Louisiana. Cheniers differ from beach ridge sets (*q.v.*) by the presence of intervening, parallel, low, and level surfaces that represent predominant portion of chenier plains.

Cheniers are separated by wide intertidal or shallow subtidal mudflats, much wider than the ridges, covered by wetland vegetation. Low-lying interior inter-ridge flats may occasionally be submerged by high tides or river floodwaters to form brackish or fresh water wetlands. Intervening areas may even be barren of vegetation or covered at least in part by grasses, salt marsh, freshwater swamp, or mangroves. Misidentification can happen when inter-ridge swales (lows) in a (non-chenier) ridge plain are filled by fine-grained peaty alluvial or delta deposits, covered by recent wetland vegetation. The term chenier plain includes both the inter-ridge flats and the chenier ridges, *where at least two successively formed ridges are present* (Otvos and Price 1979). The latter caveat, unfortunately, has occasionally been disregarded. The term was misapplied to beaches capped by shell-concentrate berm ridges at high-tide level where the beach foreshore faces low intertidal mudflats along the ocean shore (e.g., Neal et al. 2002; Weill et al. 2012). Equally inappropriately, a “chenier” designation was also applied in at least one highly dynamic setting with significant ebb and flood tidal action. Prograding arms of constantly changing, low intertidal sand bar ridges in this small estuary have been migrating over and surrounded small areas of intertidal-subtidal mud flats (Morales et al. 2014).

Successively formed cheniers usually occur in a subparallel seaward sequence as the coastline progrades. Even where there is a substantial shell content, the slopes bordering cheniers tend to be gentle (Byrne et al. 1959). In the fine sandy, shell-free Suriname cheniers, the slope is nearly level (Augustinus et al. 1989). Two of the world’s largest chenier plains are found along the coasts of Louisiana and Guiana (Fig. 1), but smaller chenier plains occur on many coastal plains, often behind embayments.

Chenier coasts are associated with a broad spectrum of tidal amplitudes that range from microtidal (e.g., SW Louisiana) and mesotidal to macrotidal (e.g., Van Diemen Gulf in northern Australia, Gulf of California). Most cheniers are of mid- to late-Holocene age, but cheniers of Pleistocene age occur in the Colorado Delta region of the Gulf of California (Meldahl 1993).

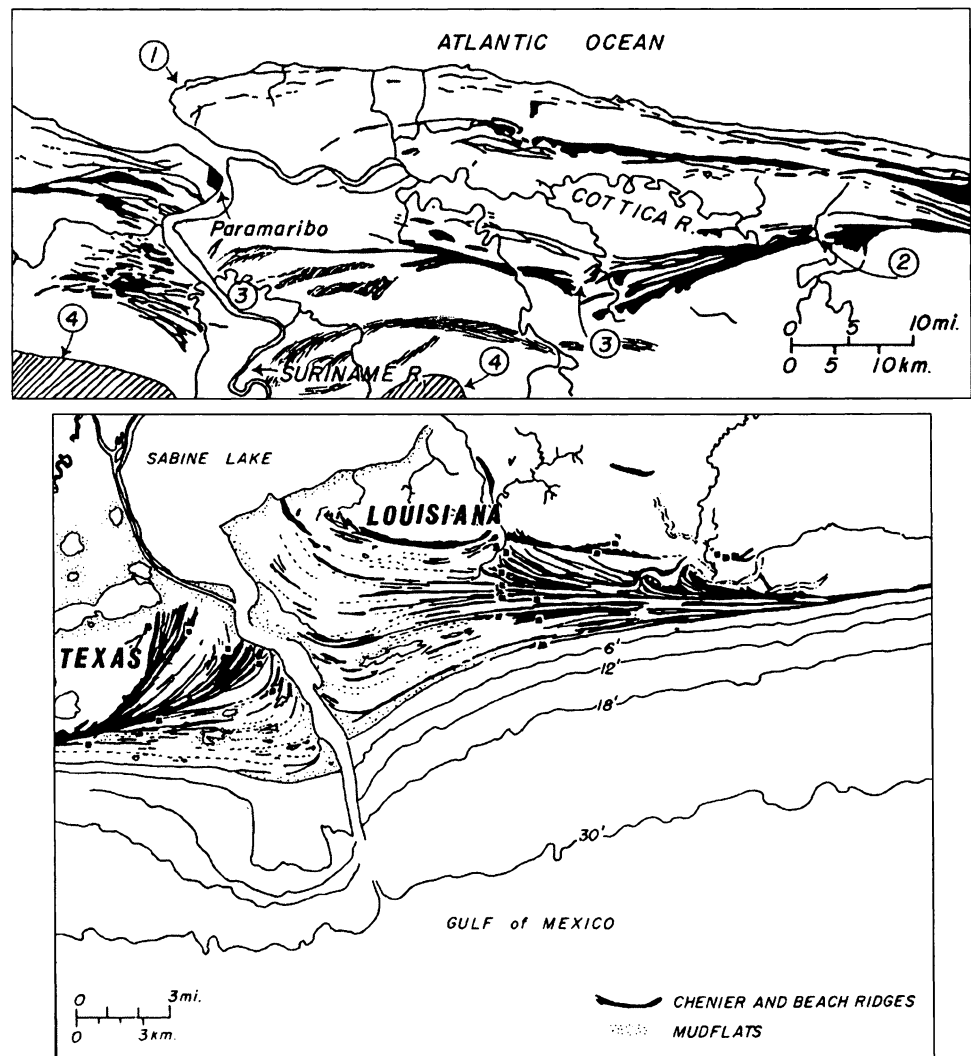
Due to constructive overwash from storm surges, the chenier crests may slightly exceed highest high-tide level. As traced laterally along given chenier ridge sets in Louisiana (Fig. 2), locally variable sand supply and wave conditions resulted in highly variable ridge crest elevations. Given an adequate coastal sand supply and sufficiently strong onshore winds, cheniers may be capped by small foredunes, or at least veneered by eolian sand (e.g., Suriname, Augustinus et al. 1989; SW Louisiana, Byrne et al. 1959). Chenier ridge heights, including ridges, even if their elevation not dune-enhanced, do not necessarily mark former sea levels.

In small bays mudflat progradation, alternating with sand and shell accumulation and *chenier* incorporation into developing chenier ridges occur simultaneously (Woodroffe and Grime 1999). Side-by-side formation of a chenier passing laterally into a sandy beach ridge was described from the northeastern coast of Australia (Chappell and Grindrod 1984).

Regressive and Transgressive Cheniers: Sea Level Relationship

Regressive cheniers that formed successively along the shore of a prograding delta or coastal plains overlie shallow near-shore marine and intertidal deposits (Fig. 3). On tropical shores, these cheniers may extend 1 m or more below the surrounding high-tidal mudflats, down to the buried top of mangrove deposits (Cook and Polach 1973). Except beneath landward-inclined slopes, formed by overwash, seaward-dipping low angle (<15°) lamination and cross-stratification are typical in most of the cheniers. Sandy shell ridges, driven landward over intertidal flats and through mangrove thickets, are transported by intermittent wave swash during high tides. Eventually, they become stranded and stabilized as transgressive chenier ridges (Fig. 3) (Thomson

Cheniers, Fig. 1 Guiana, NE South America, and SE Texas-SW Louisiana chenier trends and chenier plains, (After Byrne et al. 1959, and Vann 1959; from Otvos and Price 1979. Reprinted with permission from Elsevier Science)



1968; Jennings and Coventry 1973; Woodroffe et al. 1983; Chappell and Grindrod 1984; Woodroffe and Grime 1999). In a detailed morphological discussion of Louisiana chenier ridges, McBride et al. (2007) identified originally longshore-prograded regressive ridges and their recurved spit-ends as a separate category. Leading to speculation of higher than present late Holocene marine highstand due to aggradation by tropical storm overwash, chenier berms in Louisiana and elsewhere may exceed the highest high-tide level by 1–2 m.

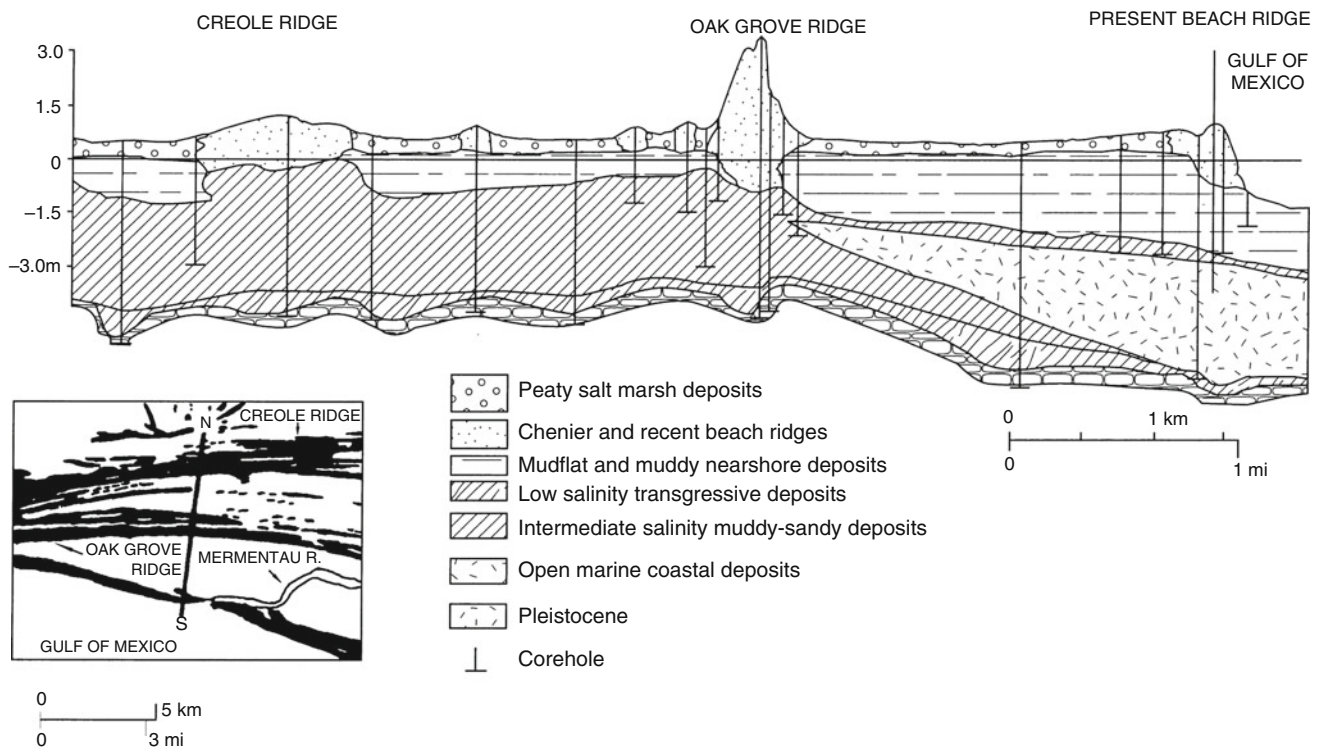
Following the original definition of cheniers as transgressive features (Russell and Howe 1935, pp. 27–34), several authors defined “true” chenier ridges as landward-driven landforms (Otvos 2000). However, stratigraphic and morphological data indicate that most Louisiana cheniers are not transgressive.

Transgressive cheniers are generally emplaced during storm surges, when large waves on a temporarily raised sea level sweep beach sediment inland. With their shallow ridge bases, transgressive chenier ridges display washover-

induced, landward-inclined stratification. They overlie intertidal, low-supertidal mud/marsh surfaces at very shallow depths. Powerful tropical storms often erode and even flatten landward migrating ridges. Even if lesser events effectively translated them landward before and after the storm, the 1974 hurricane surprisingly had no appreciable impact on the landward migration of sand and shell ridges in Shoal Bay, Australia (Woodroffe and Grime 1999). Several articles shed light on details of evolution and regional correlations of cheniers in Australia (Horne et al. 2014) and New Zealand (Daugherty and Dickson 2012; Horne et al. 2015). Chenier plains in northern and southern Australia developed mostly between 2400 and 1300 yr BP and probably during the last 1000 years.

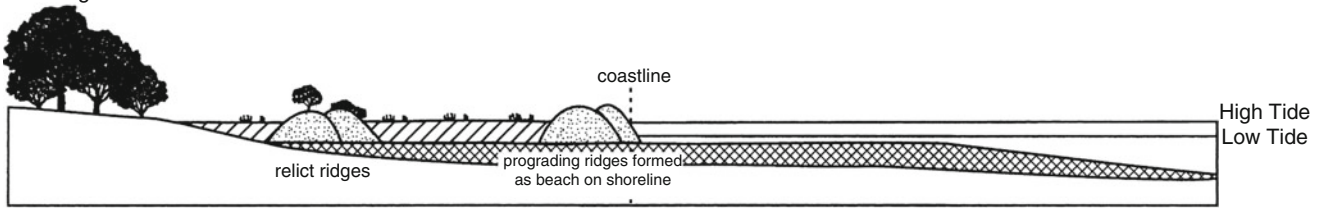
Conditions of Mudflat Development

Alternating chenier ridge and inter-ridge-mudflat development, in Louisiana, has long been associated with periodic avulsion-

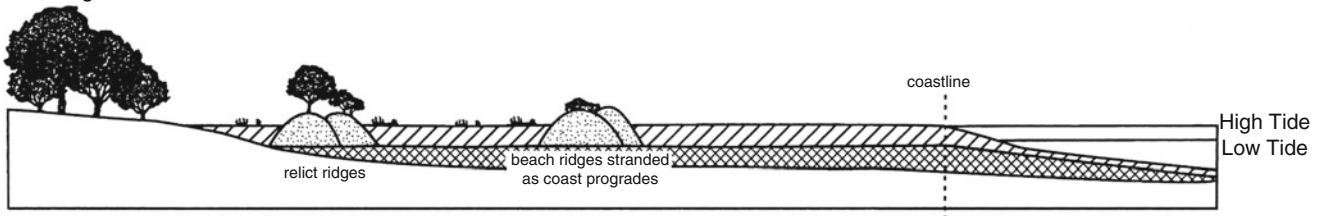


Cheniers, Fig. 2 Cross section across Creole and Oak Grove Ridge cheniers, SW Louisiana (After Byrne et al. 1959; from Otvos and Price 1979. Reprinted with permission from Elsevier Science)

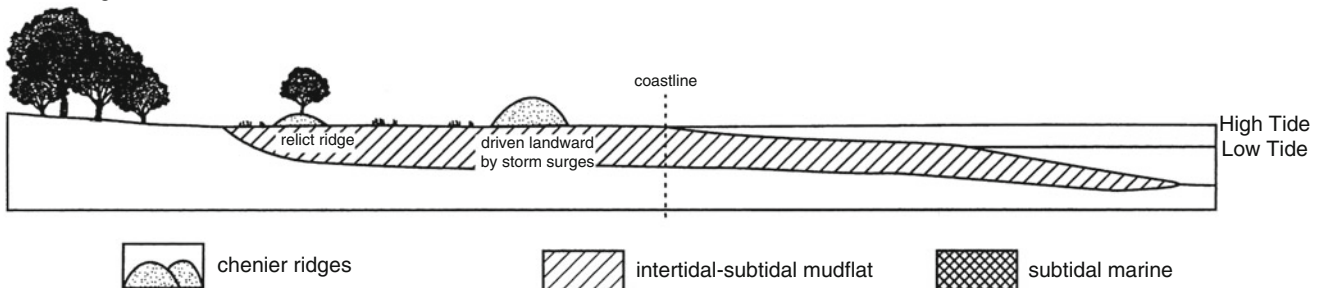
1a . Regressive Chenier



1b . Regressive Chenier



2 . Transgressive Chenier



Cheniers, Fig. 3 Schematic development of regressive and transgressive cheniers (Otvos 2000)

related delta lobe switching. Distal positions of river discharge generally result in (mudflat) shore erosion and shell and sand concentration. This leads to formation of active (“pre-chenier”) littoral ridges along the shoreline. Large-scale mud discharge from nearby (proximal) active lobes, on the other hand, favors mudflat progradation. Roberts and Huh (1995) and Anthony et al. (2011) provided valuable new insights regarding the evolution of Amazon-discharge-related Guiana mud belts, respectively, recently developing Atchafalaya mudflats along Louisiana’s previously severely receded eastern chenier plain sector. Longshore transport from the Atchafalaya delta and onshore stranding of suspended fluid mud during winter cold front water level set-down resulted in extensive mudflat formation. Momentarily dry weather and other climate conditions favor the process. In the course of desiccation and compaction, low-tidal flats prograded along a 20 km stretch of the low-microtidal chenier coast. Progradation averaged 50 m/year in the most actively accreting sectors.

Summary

Cheniers plains, including regressive and transgressive relict littoral ridge sets and intervening, often wetland-covered intertidal mudflats reflect important coastal episodes of stream discharge and shore erosion- and progradation-related changes. The occasionally still misused term “chenier plain” includes both the inter-ridge flats and the chenier ridges, *where at least two successively formed ridges are present*. Chenier ridges are isolated from the active beaches that front the chenier plains; they are insulated from daily littoral processes of wave and tidal action. Ridges are composed of sand, seashell, and occasionally gravel. New findings confirm the role of weather conditions in the sediment sources, desiccation, and consolidation of inter-ridge mud flats. In the large Louisiana chenier plain, delta lobe switching due to trunk stream avulsion played a role in changing sites of mud flat deposition. Two late Holocene episodes of chenier plain development have been identified in Australia’s numerous small chenier plains. (*q.v.*, Beach ridges; Gulf shorelines: Last Eustatic Cycle).

Cross-References

- [Beach Features](#)
- [Beach Ridges](#)
- [Cross-Shore Sediment Transport](#)
- [Deltas](#)
- [Muddy Coasts](#)
- [Sandy Coasts](#)
- [Storm Surge](#)

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Cliffed Coasts

Alan S. Trenhaile

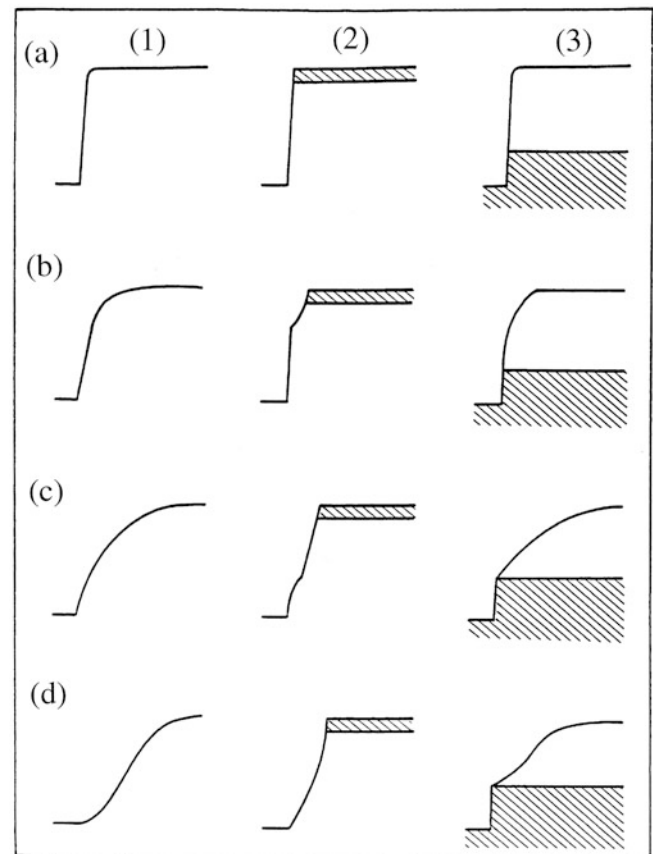
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It has been estimated that there are cliffs, consisting of rock or cohesive clays, around about 80% of the world's oceanic coasts (Emery and Kuhn 1982). Sea cliffs are found in all types of morphogenic and tectonic settings, although they are generally absent from low-lying, sediment-abundant, Amero-trailing coasts, which have plate collision coasts and high mountains on the opposite side of the continent. Despite the morphological importance of marine cliffs, and the aesthetic contribution they make to the littoral environment, there are only a few systematic descriptions of their form and development in the literature (Trenhaile 1987, 1997; Sunamura 1992).

Cliff Morphology

The type of scenery that develops on a cliff coast is the product of a number of factors, including the morphology of the hinterlands, present and past climates, wave and tidal environments, changes in relative sea level, and the structure and lithology of the rocks. Although cliff morphology is difficult to classify on the basis of climate or wave regime, some types of cliff are more characteristic of particular morphogenic regions than others (Fig. 1).

Steep cliffs usually develop in marine-dominated environments and convex cliffs where subaerial processes dominate. Where both process suites are effective, subaerial processes produce a convex slope in the upper portion of cliffs, while marine processes cut steep cliff faces at their base. Coastal slopes therefore tend to be greater in the vigorous, storm wave environments of the mid latitudes than in high latitudes or in the tropics, where frost and chemical weathering are important. Wave action is generally weak in high latitudes because of sea ice, and in some places coastal configuration. Most steep slopes on these coasts are the product of glacial erosion and there are few true marine cliffs. In the humid tropics, the formation of steep marine cliffs is inhibited by fairly weak wave action, protective coral or algal reefs, and extensive coastal plains. Chemical weathering plays an important role in cliff development, and the weathered material is very susceptible to slumping, mudflows, and other mass movements. Coastal slopes are often covered in vegetation for all but the lowest few meters, and steep cliffs may be restricted to headlands and other exposed areas. Steep, bare cliffs are much more common in the arid tropics, particularly in exposed areas where spray, high evaporation and salt



Cliffed Coasts, Fig. 1 The effect of marine and subaerial processes and variable rock hardness on cliff profiles. Columns (1), (2), and (3) represent homogeneous rocks, harder rocks (shaded) at the top of the cliff, and harder rocks at the cliff base, respectively. Rows (a), (b), (c), and (d) represent environments in which marine erosion is much greater than, greater than, equal to, and less than weathering, respectively. (After Emery and Kuhn 1982)

crystallization, alternate wetting and drying, corrosion, and other weathering processes help to steepen coastal slopes and inhibit the growth of vegetation.

There are numerous exceptions to the morphogenic classification of marine cliffs. For example, despite strong waves, there has been only minor modification of glacially sculptured granites in Maine and eastern Canada. On the other hand, despite weak wave action, steep, bare cliffs are common in the limestones of the Canadian Arctic, and they occur throughout the wet and dry tropics where strong waves can be generated by onshore monsoons, tropical cyclones, and Trade Winds.

Geological factors are at least as important as climate and wave regime in determining the relative efficacy of marine and subaerial processes, and therefore the shape of cliff profiles (Emery and Kuhn 1982). For example, weaker material in the upper part of a cliff increases the effects of subaerial weathering, whereas weaker material in the lower part of a cliff facilitates wave erosion. Cliff profiles are also strongly influenced by structural weaknesses, stratigraphic variations,

and the attitude or orientation of the bedding (Emery and Kuhn 1980; Trenhaile 1987, 1997). Very steep cliffs generally develop in rocks that are either horizontally or vertically bedded, and more moderate slopes in seaward or landward dipping rocks. The relationship between cliff profiles and the dip of bedding and/or joint planes is very complex, however, and vertical and seaward-sloping cliffs can be formed in horizontal, or seaward or landward dipping rocks.

Composite cliffs have at least two major slope elements. Multistoried cliffs have two or more steep faces separated by more gentle slopes, whereas bevelled (hog's back, slope over wall) cliffs consist of convex or straight seaward-facing slopes above steep, wave-cut faces. Bevelled cliffs can develop in homogeneous rocks where marine and subaerial processes are of comparable importance, or where weaker rocks overlie more resistant materials (Fig. 1). High composite cliffs in ancient, resistant massifs developed over very long periods of time, however, and they experienced marked changes in climate and sea level during the Quaternary. Wave-cut cliffs were abandoned during the last glacial stage, when frost and periglacial mass movements were probably dominant processes in the ice-free middle latitudes. The cliffs were gradually replaced by convex slopes developing beneath the accumulating talus, and when the sea rose to its present position marine processes removed the debris and, depending on the relative strength of the waves and the rock, either trimmed the base of the convex slopes to form composite cliffs, or completely removed it to form steep, wave-cut cliffs. Modeling suggests that bevelled profiles develop if the debris reached the top of the cliff during the last glacial stage, and multistoried profiles if the debris only rose part of the way up the cliff face (Griggs and Trenhaile 1994).

There are usually shore platforms or beaches at the foot of cliffs, but some plunge directly into deep water. Plunging cliffs developed where the postglacial rise in relative sea level was much greater than the rate of sediment accumulation at the cliff foot. Once sea level had become stable, erosion was inhibited by the lack of abrasives at the water level, and reflection of the incoming, nonbreaking waves. Plunging cliffs are particularly common around basaltic islands in the Southern Hemisphere, where there is slow erosion and little longshore sediment transport.

Cliff Erosion

The erosion of sea cliffs is episodic, site-specific, and closely related to prevailing meteorological conditions. Although recession rates have been measured in many parts of the world, most are little more than rough estimates. Reported rates vary from virtually nothing up to 100 m year^{-1} (Kirk 1977; Sunamura 1992), and it is difficult to identify patterns

that can be correlated with variations in rock type or wave and tidal conditions. A variety of techniques have been used to determine rates of erosion, but the most accurate ones have usually been used over only very short periods of time. Japanese workers have studied the short-term effects of coastal dynamics on cliff erosion. Variations in recession rates were partly explained by differences in the frequency of waves that exceed the minimum height capable of causing erosion, and they increased as the compressive strength of the rocks decreased (Sunamura 1992); however, many other factors need to be considered. In south Central England, for example, the compressive strength of the rocks decreases toward the east, but this is countered by changes in the fracture pattern, which favors increasing rates of cliff erosion to the west (Allison 1989).

Other Features of Cliffed Coasts

Small bays, narrow inlets, caves, arches, and stacks are usually the result of erosion along structural weaknesses, particularly bedding, joint, and fault planes, and in the fractured and crushed rock produced by faulting. These features form in rocks which have well defined and well spaced planes of weakness, yet are strong enough to stand as high, near-vertical slopes, and as the roofs of caves, tunnels, and arches. They are therefore uncommon in weak or thinly bedded rocks with dense joint systems (Trenhaile 1987).

Narrow, steep-sided inlets (geos) often develop along steeply dipping joints or faults in horizontal or gently dipping rocks, and from the erosion of dykes and the collapse of lava tunnels in igneous rocks. The occurrence and morphology of caves are usually controlled by weaknesses associated with joints, faults, breccias, schistosity planes, unconformities, irregular sedimentation, and the internal structure of lava flows. Marine caves are usually quite small, but large karstic caves have been inherited by the sea following cliff recession. It has been assumed that arches and longer tunnels develop through the coalescence of caves driven into the opposite sides of headlands. However, arches near La Jolla, California appear to be associated with coastal indentations and surge channels, that pile up the approaching waves on one side of a point, causing penetration of the weaker rocks at the cliff base (Shepard and Kuhn 1983). Arches can also be formed by the marine invasion of karstic caves. Fountains of spray are emitted through blowholes when large breakers surge into tunnel-like caves connected to the surface along joint or fault-controlled shafts. Marine inheritance of karstic sinks, sinkholes, and tunnels has produced spectacular blowhole systems in some areas.

Most stacks are the result of the dissection of the coast along joints and other planes of weakness, although

some are associated with folding, submergence of tropical tower karst, and differences in resistance associated with solution pipes, induration, or variable lithology. Some stacks develop from the collapse of the roofs of arches, but many form directly from erosion of the cliff face. Spectacular stacks have developed from the dissection of coarse, poorly sorted conglomerates and arkosic sandstones along prominent, well-spaced joint planes in the Bay of Fundy, which has the largest tidal range in the world. Old photographs suggest that these stacks may have life spans, ranging from their initial separation from the cliff to their eventual collapse, of between 100 and 250 years. Notches are ubiquitous at the cliff foot, and they are responsible for the characteristic mushroom-shape appearance of the stacks. Although there is no consistent relationship between the depth of notches on the seaward and landward sides of the stacks, the notches are at higher elevations on the seaward side. The deepest part of most notches is a little below the mean high tidal level, although several are up to a meter or two below it, especially on the landward side of the stacks. Stack morphology and notch depth change in a fairly predictable manner through time, as the stacks become increasingly isolated from the cliff (Trenhaile et al. 1998).

Although it is generally assumed that headlands consist of more resistant materials than those in the adjacent bays, there have been few detailed investigations of variations in rock structure or strength. The occurrence of small bays and headlands often reflects the rather subtle influence of rock structure, including variations in joint density, the orientation of discontinuities, and the thickness, strike, and dip of the beds. Large bays and prominent headlands are more likely to reflect differences in the lithology of the rocks, however, although they could also develop in fairly homogeneous rocks if, as theory suggests, the low cliffs around stream outlets retreat more rapidly than the higher cliffs on the adjacent interfluvies.

The plan shape of crenulated coasts may eventually attain an equilibrium state. This would occur when, because of the increasing effect of wave refraction as coasts become more indented, the more resistant rocks on the exposed headlands are eroded by higher waves at the same rate as the weaker rocks in the increasingly sheltered bays are eroded by lower waves. The plan shape would then be maintained through time as the coast retreated landward. As much of the erosive work could have been accomplished in previous interglacial stages, when sea level was similar to the present day, only minor modification of inherited coasts may have been required to attain equilibrium at present sea level; however, we lack reliable, long-term data on cliff erosion rates that are needed to determine whether the plan shape of rock coasts has attained a state of equilibrium.

Inheritance

In many areas, erosional processes are modifying elements of cliffed coasts that were inherited or partly inherited from the past (Bryant et al. 1990; Trenhaile et al. 1999). For example, raised beaches in south-western Britain contain middle and upper Pleistocene material which suggests that the composite cliffs behind are very old, and dated sediments in marine caves in southwestern Britain, northwestern Spain, off northern France, and in southern Australia have shown that the caves are at least as old as the last interglacial stage (Davies 1983; Trenhaile et al. 1999). The occurrence of these ancient features is consistent with the paleo sea-level record, which shows that interglacial sea levels were similar to the present level on a number of occasions during the middle and late Pleistocene. Inheritance is most likely to have occurred on tectonically stable coasts, and in resistant, slowly eroding rocks. It is much more difficult to determine whether it has occurred in weaker rock areas, where any till or other sedimentary covers, raised beaches, structural remnants, and other evidence which may have existed would have been removed by fairly rapid wave erosion at the present level of the sea.

Cohesive Clay Coasts

Cohesive clay coasts have many of the characteristics of weak rock coasts, and they are often fronted by wide shore platforms. Fine-grained sediments lose their cohesion when they are eroded, however, and the debris is generally transported offshore as suspended load. Therefore no permanent or continuous beach develops, and as the eroded material provides no protection to the cliff, the height of the cliff has no effect on the rate of erosion. Variations in groundwater level, which change the strength and stability of clay materials, is a crucial factor in the failure of cohesive coastal slopes. In temperate regions, erosion of cohesive coasts by marine and subaerial processes is often markedly seasonal in nature, reflecting variations in wave energy, temperature, precipitation, and groundwater pressures. There is a strong relationship between land-slide and flow activity in cohesive sediments and the occurrence of clays with a high proportion of montmorillonite or other swelling minerals. These minerals increase the frequency and mobility of slope movements, and allow them to occur on more gentle slopes. Shallow sliding occurs as mudslides or, if the water content is very high, as mudflows. Clay cliffs are also susceptible to deep-seated rotational landslides, which occur where basal erosion is rapid enough to remove the mudflow debris and steepen the underlying coastal slopes.

Hutchinson's (1973) form/process classification of cliffs in the London Clay of southeastern England is considered to

be generally representative of slopes in stiff fissured clays (see “► [Mass Wasting](#)”). Three types were distinguished, based on the relative rates of basal marine erosion and subaerial weathering:

1. Type 1 occurs where the rate of basal erosion is broadly in balance with weathering and sediment is supplied to the toe of the slope by shallow mudsliding. Removal of the mudslide debris stimulates further sliding, and unloading, softening, and weathering of the exposed clays.
2. Type 2 is found where basal erosion is more rapid than weathering. Waves remove all the material supplied by mudslides and erosion, and undercutting of the in situ clay steepens the profile. This eventually causes a deep-seated mass movement, which is characteristically a rotational slide involving basal failure. The sea removes the slump debris and toe erosion then steepens the slope until another failure occurs.
3. Type 3 coastal slopes develop where there is no basal erosion. When a cliff is abandoned by the sea or coastal defenses are constructed at its base, debris is carried to its foot by shallow rotational slides. Therefore, abandoned cliffs often have a steeper upper slope on which landsliding is initiated, and a flatter lower slope where colluvium accumulates. The slides eventually become quiescent when slope gradients have been reduced to the ultimate angle of stability against landsliding, and hill wash and soil creep gradually convert the bluffs into a series of undulations and then into smooth slopes.

Hutchinson's (1973) three cliff types also develop, in response to long-term changes in lake level, along the northern shore of Lake Erie. Variations in wave intensity and the type of glacial sediments in the cliffs, however, also account for the simultaneous distribution of the three cliff types along this coast (Quigley and Gelinis 1976). Hutchinson's classification is also generally applicable to cohesive cliffs in Denmark, although there are some significant differences resulting from glacio-isostatic recovery and changes in relative sea level (Prior and Renwick 1980). Despite the general applicability of Hutchinson's model to a wide range of cohesive sediments, however, alternate models of cliff retreat are relevant to specific areas.

Cross-References

- [Cliffs, Erosion Rates](#)
- [Cliffs, Lithology versus Erosion Rates](#)
- [Faulted Coasts](#)
- [Mass Wasting](#)
- [Notches](#)
- [Paraglacial Coasts](#)

- [Rock Coast Processes](#)
- [Shore Platforms](#)

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Cliffs, Erosion Rates

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Definition

Sea (or coastal) cliffs are one of the major features characterizing rock or rocky coasts, which are composed of materials ranging from hard rocks such as granite and basalt to a very

soft cohesive substrate such as glacial deposits. The recession of sea cliffs, an irreversible morphological change, occurs on rocky coasts. Although multiple factors are involved in cliff recession, (1) wave action at the foot of cliffs and (2) lithological resistance at the cliff base are crucial when considering the rate of cliff recession, i.e., erosion rates. The two factors significantly affect the areal-temporal variations in erosion rates.

Erosion Rates

Erosion rates of receding sea cliffs are determined by dividing the recession distance by the length of the time interval. Recession distances have been documented by a variety of techniques (Sunamura 1992; Hapke 2004), of which the most common is to measure the change in the position of a cliff top (or edge) on sequential aerial photographs and/or historic maps. In situ measurements using surveys, microerosion meters, and pegs have also been applied. Recently airborne and terrestrial LiDAR (Light Detection and Ranging) techniques have been employed for cliff erosion studies (e.g., Young et al. 2010; Earlie et al. 2015).

Temporal Variations in Erosion Rates

One of the crucial factors for cliff recession is erosion by wave action on the base of a cliff. Erosion occurs only when the assailing force of waves surpasses the resisting force of cliff-forming rocks. On a coast where the rock resistance can be assumed to be time-independent, fluctuations in the wave factor are responsible for the temporal changes in erosion rates. The wave assailing force is obviously dependent on wave energy level in the offshore region. The occurrence frequency of waves that actually cause erosion is important when considering erosion rates: the more frequently large waves attack a coast during a certain period of time, the more severely the cliff recedes to produce a higher erosion rate. A study by Sunamura (1982) on two Pliocene soft-rock cliffs in Japan indicates that the short-term erosion rates are closely related to the frequency of erosion-causing waves.

The wave assailing force at the cliff base is greatly controlled by multiple factors other than offshore wave energy; they are water level, fronting beaches, taluses, nearshore underwater topography, etc. Fluctuations in these controlling factors affect temporal variations in erosion rates. An influential factor on the short-term dramatic erosion is abnormal rise in water level, known as “storm surges” or “surges.” A remarkable water level rise may occur on tidal coasts when storm surges superimpose on high tides. The well-known 1953 North Sea storm surge caused drastic erosion of glacial-sand cliffs at Covehithe, Suffolk, England (Williams 1960). The storm surge with a maximum height

of 2.3 m occurred at high tide yielding a total sea level rise of 3.5 m above mean sea level. A glacial-sand cliff was cut back 9 m due to the action of storm waves during about two hours at the peak of the surge. Water level rise in much longer temporal scales has occurred along the west coast of North America, where monthly mean sea levels have been increased by up to several tens of centimeters during El Niño Southern Oscillation (ENSO) events. Dramatic increases in cliff erosion during the 1982–1983 and 1997–1998 ENSO events have been reported from Oregon (Komar 1986, 1998) and California coasts (Strorlazzi and Griggs 1998; Griggs et al. 2005). On enclosed water bodies such as the Great Lakes, seasonal changes in water levels are experienced due to fluctuations in precipitation and evaporation. A good correspondence between erosion rates of bluffs composed of glacial till and surge-induced extreme water level rises superimposed on seasonal variations has been reported from the Ohio shore of Lake Erie (Carter and Guy 1988).

On some occasions, a beach in front of a cliff decelerates cliff erosion working as a wave-energy dissipater; on other occasions, a fronting beach accelerates erosion working as a feeder of abrasive sediment to waves arriving at the cliff base to enhance the assailing force by means of its mechanical action. Where the cliff material is so soft that only the hydraulic action of waves is sufficient to cause erosion, critical for its initiation is whether waves running up the beach are able to reach the cliff-beach junction. This is controlled by many factors: input wave characteristics (wave height and period), the beach morphology (beach width, beach level, and slope), the sediment grain size, and the water-level factors described above.

The reduction or removal of a fronting beach occasionally leads to a dramatic increase in sea-cliff erosion. This occurred at a sea cliff at Pacific Manor, Pacifica on the central California coast during the 1997–1998 ENSO/El Niño Southern Oscillation (ENSO) event. The cliff, composed of poorly consolidated eolian and alluvial deposits, had receded at a long-term average rate of 0.25 m/year (Lajoie and Mathieson 1985) before the ENSO event, but it experienced 14-m retreat during this short time period (Sallenger et al. 2002). The severe erosion was caused by the attack of storm waves with extreme high runup levels, which reached the cliff base with less energy dissipation due to localized lowering and narrowing of the beach in front of the cliff.

Along the North Norfolk and Suffolk coasts, England, Lee (2008) has examined on the year-by-year basis the recession rate of Pleistocene soft rock cliffs in connection with the volume (above High Water Level) of the fronting beach during a period of 11 year, and provided for each coast the relationship between recession rate and beach volume, plotted with highly scattered data. An envelope of the data cluster shows that, as the sediment volume increases, the maximum recession rate (for a given sediment volume) tends to increase

abruptly and decrease after the volume exceeds some optimal value. A more marked trend is seen on the Suffolk coast.

The effect of a fronting beach on cliff toe erosion has been examined by Sunamura (2015) based on data of extremely short-term erosion rates measured by Carter and Guy (1988) at Helen Drive site (a till cliff) on the Ohio shore of Lake Erie. They have performed a detailed study on cliff-base erosion processes during five years, including measurements of the width of a sandy beach and accurate surveys of eroded distances conducted after storms, which enable us to evaluate the erosion rate on a single storm basis. Sunamura (2015) analyzed such erosion data at seven storm events having occurred under similar water-level conditions (including surge height) and constructed a model. The result of a best fit of the model to the data is plotted by the solid line in Fig. 1, which indicates that the erosion rate abruptly increases with increasing beach width and attains its maximum; as the beach width further increases, the rate drops and finally becomes null. It is seen that the fronting beach with $0 < W \lesssim 5$ m works to augment erosion, while the beach with $W \gtrsim 5$ m has a buffering effect, because the erosion rate is less than 0.3 cm/h, the rate at the time of no beach.

The influence of underwater morphological change on temporal variations in long-term erosion rates has been reported from Dunwich in Suffolk, England (Robinson 1980). Glacial-sand cliffs were cutting back at an annual rate of 1.6 m/year from 1589 to 1753, 0.85 m/year from 1753 to 1824, 1.5 m/year from 1824 to 1884, 1.15 m/year from 1884 to 1925, and 0.15 m/year from 1925 to 1977, the last being extremely low. This drop in the erosion rate can be attributed to the reduction in wave energy arriving at the coast. The energy reduction was due to a significant

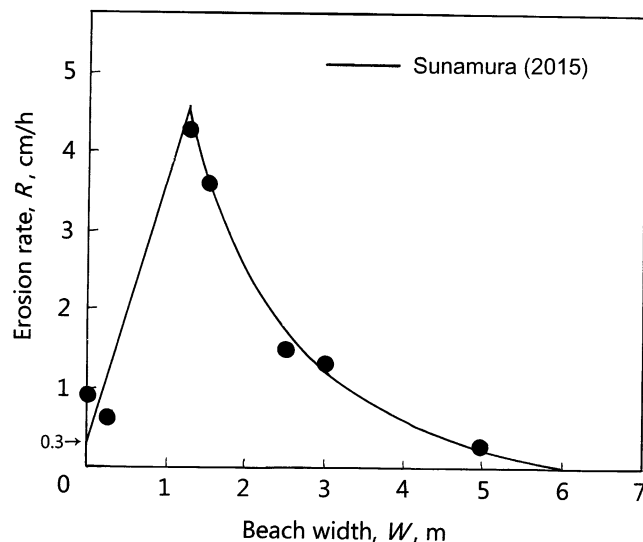
divergence in wave refraction, produced by a changing bottom configuration associated with the steady northward growth of Sizewell Bank which took place during the past century.

Taluses protect the foot of cliffs from wave attack until they are removed by the wave and current action, giving rise to no erosion. Knowledge of the residence time of taluses is necessary to consider the temporal variations in the toe erosion. In the Lincoln City littoral cell on the mid-Oregon coast, the lifetime of a talus at the base of a cliff fronted by a dissipative beach is much longer than where a steeper reflective beach is found (Komar and Shih 1993). The talus remains for several years to decades when protected by a dissipative beach because of the less-active, low-intensity infragravity dominated swash, resulting in a slow recession of the cliff; on the other hand, when a more-reflective beach develops, dominated by swash with a strong incident-wave component, the talus is rapidly removed yielding higher cliff-recession rates. On the chalk coasts in southeast England, most cliff failures yield small amounts of debris, less than 1,000 m³ in volume, which may be removed in a few weeks or months (Williams et al. 2004). Moses and Robinson (2011) have stated that the debris caused by medium-scale failures (1,000–10,000 m³) of chalk cliffs may be removed for months and implied that the decades of erosion are required for the removal of debris supplied from larger-scale failures (10,000–50,000 m³). Lageat et al. (2006) have reported from a chalk coast at Petites Dalles, south of St-Valéry-en-Caux in Seine-Maritime, France, that the debris of 7,947 m³ had been completely swept away in eight months.

Spatial Variations in Erosion Rates

Erosion rates fluctuate from place to place depending on the local variation in the wave assailing force and the rock resisting force. Suppose that several cliffs with rocks of different resistance are exposed to a similar wave climate; then the long-term erosion rate of the so-called “harder” cliffs is lower. The relationship between the erosion rate and the rock resistance has been reported from cliffs on the Pacific coast of eastern Japan (Sunamura 1977) and cliffs on the Mediterranean coast of southern Italy (Budetta et al. 2000).

Assuming that cliffs with similar lithology are subjected to the different wave climate, the long-term erosion rate should be higher where wave conditions are more severe, if other controlling factors are constant. Sunamura (1992) discussed the erosion-rate vs. wave-power relationship using data of long-term erosion rates acquired by Gelinias and Quigley (1973) along the central part (120 km long) of Canadian shore of Lake Erie, where cliffs cut in glacial deposits have developed with no marked alongshore variations in their geotechnical properties, but different exposures to waves.



Cliffs, Erosion Rates, Fig. 1 Relationship between short-term erosion rate at the toe of a cliff on the Ohio shore of Lake Erie and the width of a beach in front of the cliff (Original data from Carter and Guy (1988))

Alongshore changes in beach width give rise to spatial variations in erosion rates, if other factors are uniform. An example has been reported from the south shore of Lake Erie (Carter et al. 1986); cliffs composed of glaciolacustrine clay were receding from 1876 to 1973 at an annual rate of 1.3 to 2 m/year where the beach had a width of 0–6 m, whereas only a few to ten centimeters per year on the shore with wide beaches (>24.3 m). An interesting example illustrating the influence of the morphology and behavior of a fronting beach on the spatial variation in cliff erosion rates can be seen on the central Oregon coast (Komar and Shih 1993; Shih and Komar 1994), where sea cliffs cut into Pleistocene sandstones develop. The beaches developed along the 24 km length of the Lincoln City littoral cell range from dissipative to nearly reflective, depending on the grain size of the beach. The coarser-grained reflective beaches are steeper sloped and respond more quickly to winter storms with larger changes in beach-profile levels than do the fine-grained, gently-sloping dissipative beaches. As a result, the reflective beach is a weaker buffer against wave attack, and the sea cliff is more susceptible to erosion compared with areas where the cliff is fronted by a dissipative beach. A further decrease in buffer protection on the reflective beach is brought about by the more pronounced development of embayments eroded by rip currents, allowing for easy landward penetration of storm waves to the toe of the sea cliff.

Water level changes induced by local tectonic movement may control the wave assailing force, which results in regional variations in long-term cliff erosion rates. The influence of the relative sea level change caused by an earthquake on cliff erosion has been reported from the coast of Oregon (Komar and Shih 1993). A marked spatial variability in cliff erosion has been produced on a 500 km long coast having a north-south orientation. The tectonic movement is associated with the subduction of oceanic plates beneath the continental North American plate. An earthquake-induced subsidence occurred in the year 1700 and has been followed by a gradual aseismic uplift. Along the southern half of the Oregon coast and near the Oregon-Washington border in the north, this tectonic uplift has exceeded the global eustatic rise in sea level during the past century, resulting in a drop in relative sea level that has largely eliminated sea-cliff erosion. Along the northcentral Oregon coast, the present rise in eustatic sea level exceeds the tectonic uplift, and the relative sea level rise continues to be a factor in sea-cliff erosion, although the erosion rates are highly variable due to site-specific factors.

Spatio-Temporal Variations

On a coastal strip where the respective alongshore variations in wave exposure and lithology are considered to be uniform, it is anticipated theoretically that any location along the coast

has receded at the same rate, i.e., there is no spatial variation in erosion rate. Actually, however, there are considerable alongshore fluctuations when considering erosion rates in a short period. The fluctuations are closely associated with multiple factors described before.

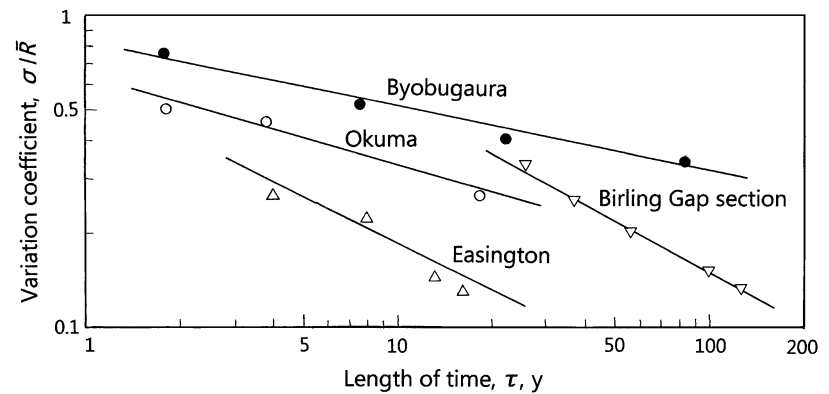
At Easington on the Holderness coast, England, Richards and Lorrimer (1987) have examined spatial and temporal variations in cliff recession rate, during the period of 16 years from 1966 to 1982 at 90 sites with an interval of 3.3 m along a 300 m cliffline. The alongshore variation in recession rate for four subperiods (3–5 years) is respectively presented in four diagrams, which show that recession rates in each subperiod are highly variable because some specific sites having receded extremely rapidly in one subperiod were cut back very slowly in the others. A diagram illustrating the spatial variation in the rates averaged over the 16-year period shows that the variability is much smaller than that seen in the subperiods.

A cliff recession study on wider spatial and longer temporal scales has been performed by Brooks and Spencer (2010) on the Suffolk coast, England, where soft cliffs composed mainly of weakly consolidated Plio-Pleistocene marine sediment are exposed. Recession measurements were carried out, on transects with a 10 m interval, in five subsections along an 11 km coastline. Recession rates for a long term (125 years from 1883 to 2008) and those for six intermediate time spans (11–40 years) were presented in diagrams for each of the five subsections. The result indicated that there were considerable temporal and/or spatial variations in erosion rates in each subsection for the intermediate time spans.

Another example of research with similar spatiotemporal scales can be taken from the East Sussex coast in England. Using the long-term (128 years: 1873–2001) topographical data collected at an interval of 50 m along a cliffline, Dornbusch et al. (2008) have examined the recession of chalk cliffs for a 26 km, approximately east-west oriented coastal stretch and presented a diagram showing spatial patterns of cliffline retreat at the time of 2001 together with those at four intervening times. A 1.5 km alongshore section including Birling Gap near the eastern end of their study area had a long-term, space-averaged erosion rate of about 0.6 m/year, which is more than double the value of the remaining sections of the study area. Analyzing Dornbusch et al.'s (2008) data of this section, Sunamura (2015) has discussed the relationship between the spatial variation in erosion rate and the length of time. The variation coefficient, $\sigma/\bar{R}$, where $\bar{R}$ is the space-averaged erosion rate and σ is its deviation coefficient, is plotted against the length of time, τ , for the Birling Gap section in Fig. 2. Other available data are also plotted; they are taken from two Neogene cliffs in Japan, Byobugaura and Okuma (Sunamura 1973), and from a till cliff at Easington, Holderness, England (Richards and Lorrimer 1987). The decrease in $\sigma/\bar{R}$ with increasing τ found on the four coasts suggests that parallel

Cliffs, Erosion Rates, Fig. 2

The variation coefficient of the recession rate, $\sigma/\bar{R}$, plotted against the length of time interval, τ , for four rapidly receding sea-cliff sites (From Sunamura (2015))



recession of the cliffline will occur over a long period of time on a coast with little alongshore variation in waves and lithology. A longer-term erosion rate at any spot along the coast tends to approach a site-specific value for the coast.

Summary

Some typical examples of temporal and spatial variations in cliff erosion rates are presented. The variations occur depending on fluctuations in the relative magnitude of the assailing force of waves at the cliff base to the resisting force of rocks forming there. The later rock factor is intrinsically of areal variations, while the former wave factor has spatiotemporal changes and it controlled by the wave energy level; water level changes caused by storm surges, ENSO events, and local tectonic movements; the behavior and the size of fronting beaches; the size of taluses; and nearshore bathymetric changes. On a coast with little alongshore variation in the two factors, the parallel retreat of the cliffline will occur over a long period of time.

Cross-References

- [Cliffed Coasts](#)
- [Cliffs, Lithology Versus Erosion Rates](#)
- [Coastline Changes](#)
- [Rock Coast Processes](#)
- [Storm Surge](#)

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Cliffs, Lithology Versus Erosion Rates

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Definition

Major controlling factors for the recession of sea (or coastal) cliffs are the assailing force of waves at the toe of cliffs and the resisting force of cliff-forming rocks. Along a coastal stretch with little regional difference in the wave assailing force, the rate of cliff recession, i.e., erosion rates, is greatly controlled by the rock resisting force, which is closely associated with lithology. Lithological factors include: (1) mechanical strength of intact rocks at the cliff base; (2) overall strength of a rock mass when it contains discontinuities such as joints, faults, and stratifications; (3) weathering susceptibility of rocks; (4) effect of fatigue due to reciprocating wave pounding on rock-strength reduction; and (5) bioactivity on cliff-forming rocks.

Lithological Control of Erosion Rates

When the wave assailing force is greater than the rock resisting force, erosion occurs resulting in augmentation of the instability of an overall cliff profile, eventually leading to intermittent mass movement. Depending on geological structures, stratigraphic features, and geotechnical properties, the mode of mass movement is generally classified into four types: falls (e.g., Duperret et al. 2004), topples (e.g., Kumar

et al. 2009; Santos et al. 2011), flows (e.g., Hutchinson 2002), and slides (e.g., Quinn et al. 2010; Styles et al. 2011). Cliffs recede with one or hybrid types of these failure modes.

Mass movement acts to render the cliff profile more stable—a more gentle sloping profile develops. Simultaneously, debris masses are supplied to the base of the cliff. Waves remove the debris and continue to undercut the cliff base, so that the overall cliff profile once again becomes steep and unstable, and then mass movement ensues. The long-term cliff recession process may be characterized by the occurrence of a cyclic change between steep and gentle profiles, so similar profiles will recur: parallel retreat of cliff profiles will take place with a certain time interval (e.g., Sunamura 1983).

Cliff recession has been documented by a variety of techniques, of which the most popular one is to measure distances, usually at the top (or edge) of a cliff, on sequential aerial photographs and/or historic maps, and erosion rates can be determined by dividing the cliff recession distances by the length of time intervals (see ► “Cliffs, Erosion Rates”). Orders of erosion rates summarized on a lithological basis are generally: 10^{-3} m/year for granitic rocks; 10^{-3} to 10^{-2} m/year for limestone; 10^{-2} m/year for flysch and shale; 10^{-1} to 10^0 m/year for chalk, Neogene and Paleogene sedimentary rocks; 10^0 to 10^1 m/year for Quaternary deposits; and 10^1 m/year for unconsolidated volcanic ejecta (Sunamura 1983). These results show that lithology and cohesiveness of the cliff material greatly control the erosion rate. It should be noted, however, that toe erosion by waves originally causes cliff retreat.

Assuming the occurrence of parallel retreat of cliff profiles on a long-term basis, erosion rates at the top and the base of a cliff are considered to be identical. Erosion rates at the cliff base may be described by the following equation, which has recently been proposed by Sunamura et al. (2014):

$$dX/dt = C[(F_w/F_r) - 1] \quad (1)$$

where dX/dt is the erosion rate, X is the erosion distance, t is the time, F_w is the wave assailing force, F_r is the rock resisting force, and C is a constant with dimensions of $[LT^{-1}]$. This equation holds when $F_w \geq F_r$; otherwise, $dX/dt = 0$. Because a short-term erosion rate is highly variable with time (see ► “Cliffs, Erosion Rates”), erosion rates over long periods of time will be considered here; generally, they show almost stable values, site-specific ones, reflecting the long-term relative magnitude of wave action to rock resistance.

Parameters for Rock-Mass Strength Against Wave Erosion

To find the most suitable parameter for expressing rock resistance to wave erosion, various strength measures have been

tested: compressive strength, tensile strength, cohesive strength, shear strength, penetration strength, and Schmidt hammer rebound value. Of these parameters, compressive strength is found to be positively related to other measures and is a widely used parameter with testing criteria having been well established, so it is suitable to use this measure as a representative strength parameter when considering the intact part of cliff-forming materials (Sunamura 1992). Compressive strength values converted from the result of in situ testings using Schmidt hammers (e.g., Budetta et al. 2000; Stephenson and Kirk 2000a), Equotip hardness testers (Aoki and Matsukura 2007), and needle-type penetrometers (Suzuki and Hachinohe 1995) have been employed in rocky coast research.

Rock resistance against wave force is quite different according to the presence of discontinuities such as cracks, cleavages, joints, faults, and bedding planes, some being inherent in lithology and others being of tectonic origin; especially cracks and joints can also be brought about by weathering if rocks are prone to weather. When waves hit a cliff with joint- or fault-associated openings, the air in the interstices is suddenly compressed, which results in pressure increase. Measurements of dynamic pressure inside a crevice in a vertical cliff set up in a wave flume indicate that a considerable increase in pressure can be produced depending on the size of opening and input wave height (Sunamura 1994). Such pressure increase exerts an excessive outward stress on the sidewall of the opening as well as on its roof. As the opening grows, the wave assailing force inside the opening begins to increase its intensity, which in turn gradually widens and deepens the opening, which causes more increase in the assailing force, finally leading to removal of blocks bounded by joints and faults, an erosional process known as plucking or quarrying (e.g., Naylor and Stephenson 2010).

From their cliff erosion study along the San Diego County coast, California, Benumof and Griggs (1999) have reported that the properties of cliff-forming materials such as joint spacing and width play a more significant role than the wave factor in influencing the long-term erosion rate. To assess the influence of discontinuities on the cliff resistance against wave action, Tsujimoto (1987) employed a non-dimensional parameter, V_f/V_1 , where V_f and V_1 are the sonic velocities measured, respectively, in the field on a rock mass with discontinuities and in the laboratory on an intact rock specimen. With increasing degree of fracturing, V_f decreases, so that the value of V_f/V_1 reduces from unity and approaches to zero. Tsujimoto attempted to evaluate the resisting force F_r multiplying this parameter by compressive strength of an intact rock specimen, S_c :

$$F_r = B(V_f/V_1)S_c = BS_c^* \quad (2)$$

where $S_c^* [= (V_f/V_1) S_c]$ is the overall strength of a rock mass and B is a constant. Budetta et al. (2000) proposed a similar

strength measure: $F_r = J_p S_c$, in which J_p depends on the volume of a rock mass and on joint condition factors such as the roughness, alteration, and size of joints. The measure was tested in the Cilento area of southern Italy.

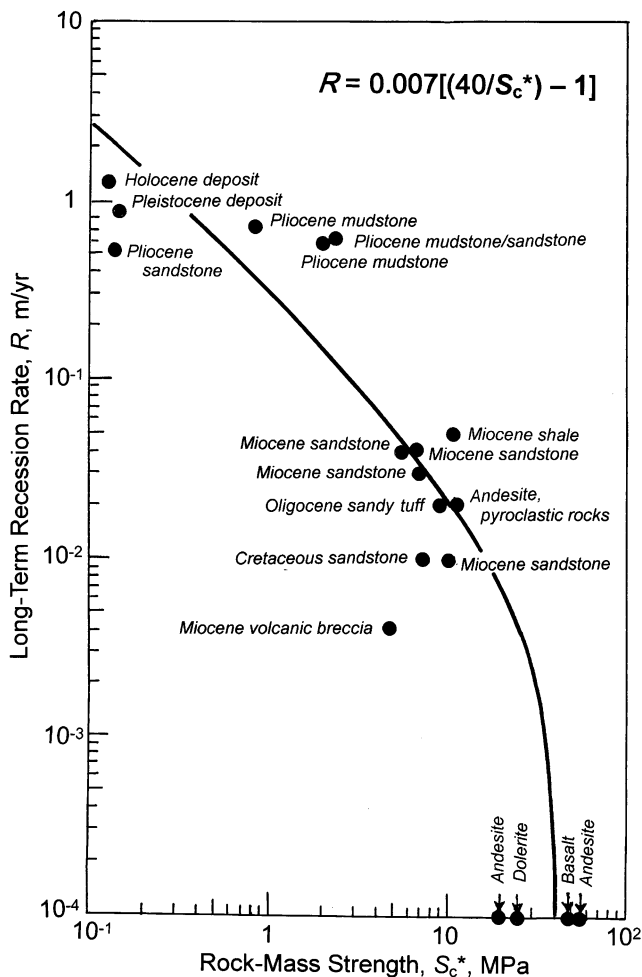
Long-Term Erosion Rates Versus Rock-Mass Strength

Let us suppose that cliffs with rocks of different resistance are exposed to a similar wave climate. The erosion rate equation for such a situation will be written from Eqs. 1 and 2:

$$R = C[(K/S_c^*) - 1] \quad (3)$$

where R is the long-term erosion rate and $K = F_w/B = \text{const.}$ Tsujimoto (1987) collected data on R and S_c^* from 19 sites on open coasts of Japan, facing the Pacific Ocean and the Sea of Japan, with the sites having no marked difference in offshore wave heights. Using his data, the R vs. S_c^* relationship is plotted in Fig. 1, where R -values denote erosion rates averaged over a long period of time more than 30 years. The erosion rates are around 1 m/year for soft rocks represented by Quaternary deposits and Pliocene sedimentary rocks, which have S_c^* ranging from 0.1 to 3 MPa ($S_c = 0.1$ –3.9 MPa); they are on the order of 10^{-3} to 10^{-2} m/year for intermediate-strength rocks shown by Miocene and older rocks with $S_c^* = 5$ –11 MPa ($S_c = 6.3$ –26 MPa); and they are almost 0 m/year for hard rocks of volcanics with $S_c^* = 20$ –56 MPa ($S_c = 30$ –94 MPa), the data being plotted on the x-axis. Morphological features on the coasts composed of soft, intermediate-strength, and hard rocks are respectively Type-A shore platforms, Type-B shore platforms, and plunging cliffs (Tsujimoto 1987; Sunamura et al. 2014). The curve in Fig. 1 is drawn as a result of the best fit of Eq. 3 with the data. The general trend is well described by Eq. 3 with $C = 0.007$ m/year and $K = 40$ MPa, although large scatter in the data is seen. Such scatter could be attributed to: (1) cliff resistance to wave erosion cannot be properly evaluated by S_c^* alone and (2) F_w does not always take on a space-independent, constant value.

One of the factors not incorporated in Eq. 3 is deterioration of the superficial part of a rock mass due to various types of weathering: alternate wet/dry weathering, frost action, salt crystallization, and the combined action of these. Mottershead (1994) examined deterioration through strength testings on Lower Cretaceous sandstone used in sea walls at Weston-super-Mare, Avon, UK. The result indicated that the strength of weathered (honeycombed) rock is approximately 20% of that of fresh rock. Stephenson and Kirk (2000b) have reported that considerable reduction in compressive strength (up to 50%) due to wet/dry weathering had occurred on limestone forming shore platforms at Kaikoura, New Zealand. A similar reduction degree has been found on the surface of a Pliocene sandstone cliff located landward of a shore platform near Shimoda, Japan (Sunamura et al.



Cliffs, Lithology Versus Erosion Rates, Fig. 1 Relationship between long-term cliff erosion rate and rock-mass strength of cliffs on open coasts of Japan. The *arrowed* data denote that erosion rates are almost null (Data from Tsujimoto (1987))

2014). Other factors responsible for strength reduction are (1) cyclic fatigue in cliff-forming rocks due to incessant loading by waves and (2) biological activity of grazing and boring organisms. As for the former factor Adams et al. (2005) have attempted to quantify the fatigue effect on rock-mass strength of a cliff at Santa Cruz, California; for the latter factor, Healy (1968) has reported from his bioerosion study conducted near Auckland, New Zealand, that compressive strengths of Miocene sandstone and siltstone riddled by boring mollusks had only 20–35% of the strength of solid (intact) rocks.

Summary

Of lithological factors controlling the recession of coastal cliffs, the most influential factor is considered to be the rock-mass strength at the cliff base. Under this consideration,

cliff recession data, obtained from Japanese open coasts where there is no significant areal variation in wave climate, are employed to examine cliff erosion rates and the rock-mass strength. Figure 1 shows that the erosion rates decrease sharply with increasing rock-mass strength. Equation 3 with appropriately determined values for C and K can well describe the general trend, although considerable scatter in the data is found.

Cross-References

- ▶ [Cliffed Coasts](#)
- ▶ [Cliffs, Erosion Rates](#)
- ▶ [Coastline Changes](#)
- ▶ [Rock Coast Processes](#)

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Climate Patterns in the Coastal Zone

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Decadal climate changes are associated with the systematic variations in global patterns of atmospheric pressure, temperature, and sea surface temperature that persist from a year or two to multiple decades. These climate changes have major impacts on coastal processes, modifying wave climate, rainfall, river sediment flux, sand transport paths, coastal erosion, sea level, and inundation. Understanding of the types and causes of climate change is in its infancy. *El Niño/Southern Oscillation* (ENSO) patterns that generate storms comparable to extreme seasonal events may occur every 3–7 years and appear to be associated with other longer period changes. The *Pacific/North American* (PNA) pattern of pressure anomalies and its sea surface temperature equivalent, the *Pacific Decadal Oscillation* (PDO), are associated with multidecadal periods of more intense *El Niño* type weather, alternating with decades of more intense *La Niña* weather, the complementary phase to *El Niño*. Similarly, a *North Atlantic Oscillation* (NAO) is associated with decadal changes in atmospheric pressure fields between Iceland and Portugal that alter North Atlantic wave climate and Caribbean hurricane tracks.

Douglas L. Inman: deceased.

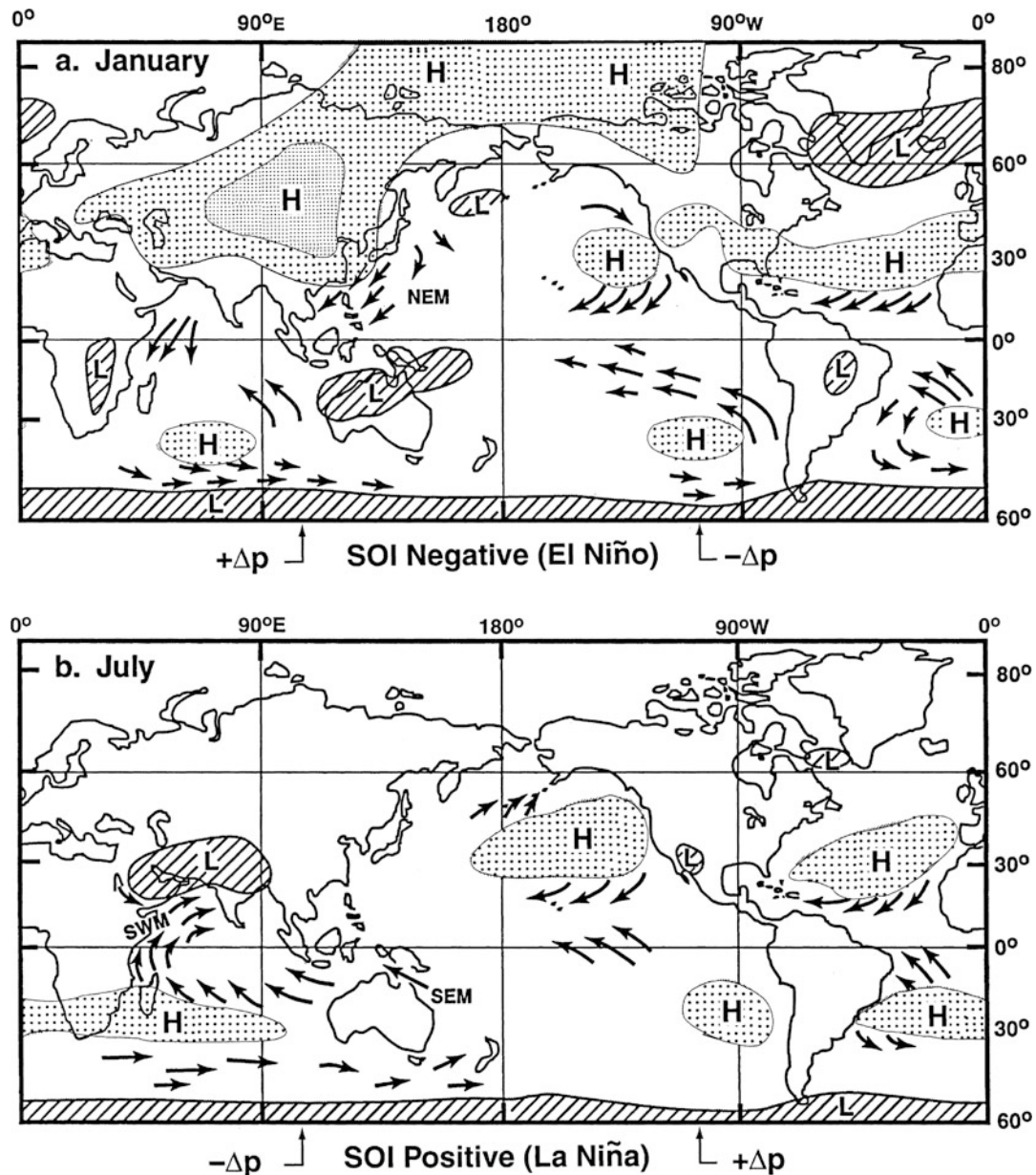
Background

The worldwide connection of weather patterns (teleconnections) was first addressed in detail by Sir Gilbert Walker (1928). He identified weather patterns with three global reversals in atmospheric pressure and temperature, the Southern Oscillation, the NAO, and the North Pacific Oscillation. Nile River floods, monsoon reversals in India, rice crop failures in Japan, and droughts in Australia were associated with these global reversals. Although there were little data at that time, Walker emphasized that ocean temperatures play a most important part in world weather. The advent of detailed global databases and rapid computer modeling in recent times has fueled a renewed understanding of the extent and duration of these climate events. Proxy records obtained from Greenland ice cores, tree rings, coral reefs, fish scales, and ocean sediments show that abrupt fluctuations in climate have occurred throughout the Holocene. These shifts may occur within a few years and extend for decades to millennia (e.g., Soutar and Isaacs 1974; Meko 1992; Cole et al. 1993, 2000; Heusser and Sirocko 1997; Gupta et al. 2003). These past events and the presently occurring climate changes are generally thought to be related to the extreme phases of the ENSO phenomenon.

Hypotheses explaining the causes of unusual ENSO events are numerous and range from the extrusion of hot lava on the sea floor to variations in the intensity of the sun's radiations. Most climate modelers favor a complex set of changes in the interaction of the atmosphere and ocean that cause fluctuations in trade winds, monsoon intensity, and sea surface temperature (e.g., Somerville 1996; Pierce et al. 2000). Although the causes of climate change remain obscure, a clear pattern of atmosphere-ocean events associated with ENSO phenomena has emerged. It is this pattern that gives rise to three-month forecasts of ENSO events. However, ongoing ENSO events are superimposed on a progressive rise in mean global temperature. Consequently, their effects are likely to be less predictable and perhaps more intense than previous events (Inman and Jenkins 1997; Timmermann et al. 1999; Fedorov and Philander 2000).

Short-Term Climate Patterns

The seasonal variations in the exposure of the hemispheres to the sun produce changes in the duration of daylight and the angle of the sun's irradiance. These effects modulate solar heating, resulting in variation of the earth's atmospheric pressure field which in turn induces seasonal climatic effects. Seasonal variations are enhanced by the higher convective effects of land and the greater concentration of landmass relative to water in the temperate latitudes of the Northern Hemisphere (Fig. 1a and b).



Climate Patterns in the Coastal Zone, Fig. 1 Seasonal pressure and winds for (a) January and (b) July. Contours of 1020 mb and 1000 mb are shown around areas of high (H) and low (L) pressure, respectively. Prevailing winds are indicated by arrows: NEM, SEM, SWM designate

northeast, southeast, and southwest monsoons. The pressure anomaly Δp is centered around longitudes 105°E and 105°W for negative and positive Southern Oscillation Indices (SOI, Modified from Inman et al. 1996)

On occasion, the typical seasonal weather cycles are abruptly and severely modified on a global scale. These intense global modifications are signaled by anomalies in the pressure fields between the tropical eastern Pacific and Malaysia known as the ENSO (e.g., Diaz and Markgraf 1992). The intensity of the oscillation is often measured in terms of the *Southern Oscillation Index* (SOI), defined as the monthly mean sea-level pressure anomaly in millibars normalized by the standard deviation of the monthly means for the period 1951–80 at Tahiti minus that at Darwin, Australia

(Fig. 1a and b), lower). A negative SOI (lower pressure at Tahiti, higher pressure at Darwin) is known as an *El Niño* or warm ENSO event, because of the arrival of unusually warm surface water off the coast of Peru at the time of Christmas, hence the term *El Niño*. Warm water also occurs along the coast of California, and both regions experience unusually heavy rainfall. A positive SOI is known as *La Niña* and it signals the occurrence of colder than normal surface water in the eastern Pacific, but stronger trade winds in the Pacific and southwest monsoons in the Indian Ocean with heavy

rainfall in India and on the Ethiopian plateau (Inman et al. 1996).

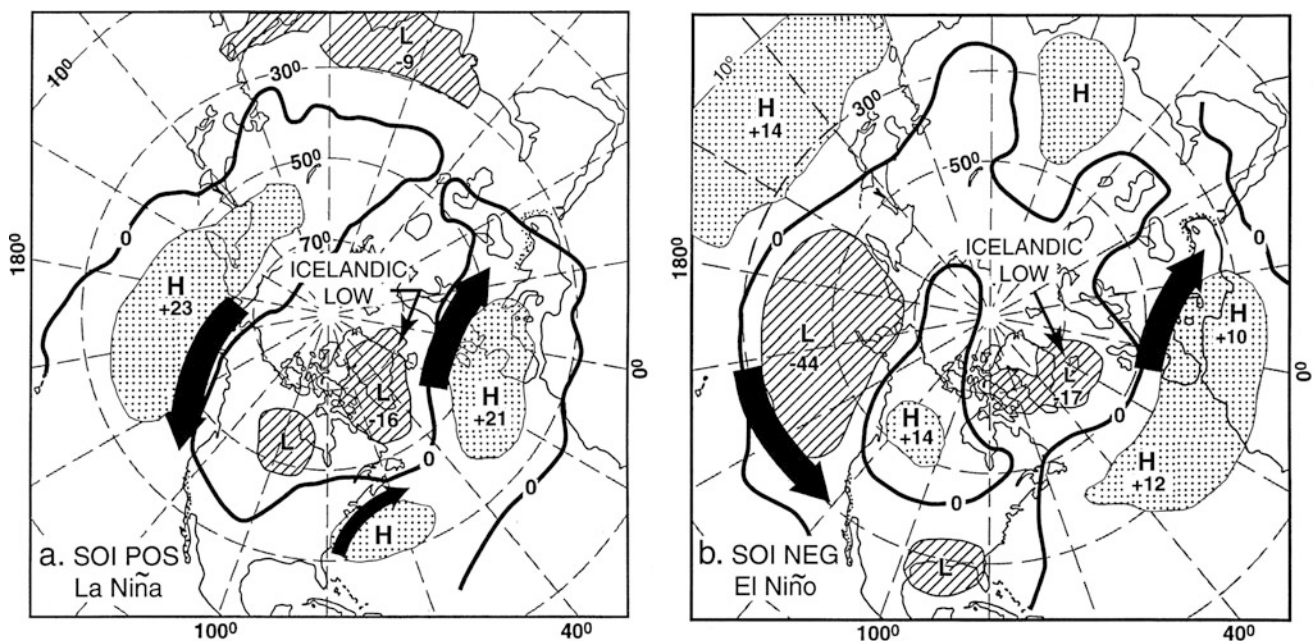
During El Niño events the combination of low pressure over the Pacific Ocean and the thermal expansion of warm water cause an increase in water level along the eastern Pacific Ocean. The water level increases by 20–30 cm during strong El Niño events and is depressed an equivalent amount during strong La Niña events. The water level increases lead to significant coastal inundation and sea cliff erosion during the intense storms that are common to El Niño events.

Decadal Climate Patterns

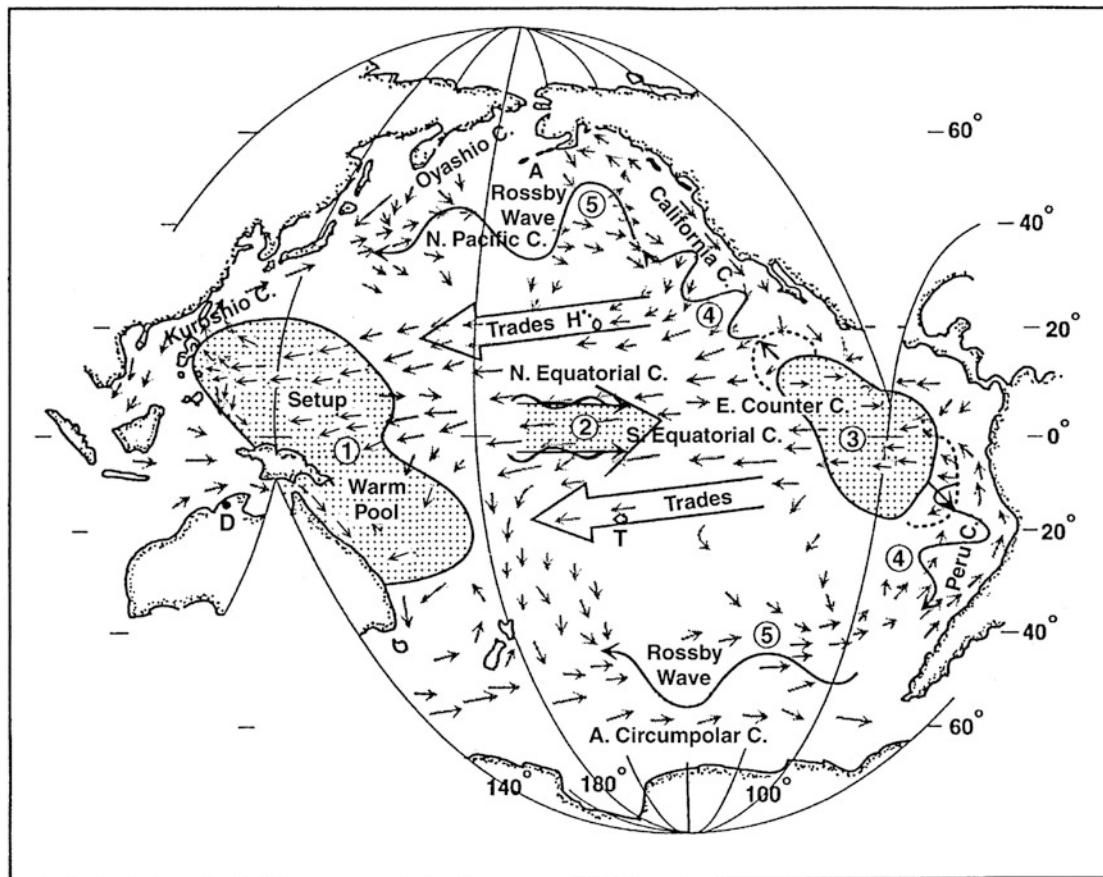
ENSO events occur with dominant spectral peaks at about 3 and 6 plus years. However, these ENSO events are modified by climate changes that occur on decadal timescales of one quarter to one half century. These changes are often discussed in terms of two atmospheric patterns, the PNA and the NAO, and a sea surface temperature pattern, the PDO. Both PNA and PDO are long period, multidecadal analogs of the seasonal variations of global pressure and temperature and are related to Walker's (1928) North Pacific Oscillation. NAO is a periodic intensification or relaxation of the January phase of the pressure variation (Fig. 1a). These decadal climate patterns are disguised (aliased) by the interannual changes because they have the same structure and appear as extreme cases of the interannual patterns. This aliasing has delayed the general understanding and acceptance of these concepts.

The PNA pattern is associated with an atmospheric dipole in pressure anomaly over the Pacific Ocean/North America region whose polarity reversals lead to wet and dry climate along the Pacific coast of North America (Wallace and Gutzler 1981). High-pressure anomaly over the North Pacific Ocean and low-pressure anomaly over the North American continent result in dry (La Niña dominated) climate along the coast of central and southern California, while the opposite polarity in these dipole patterns leads to wet (El Niño dominated) climate (Fig. 2a and b). Inman and Jenkins (1999) show that the twenty coastal rivers of central and southern California have streamflow and sediment fluxes during the wet phase of PNA (ca. 1969–98) that exceed those during the preceding dry phase (ca. 1944–68) by factors of 3 and 5, respectively. The sediment flux during the three major El Niño events of the wet phase averaged 27 times greater than the annual flux during the dry phase.

The PDO is a long-lived sea surface temperature pattern with cool and warm phases similar in structure to the La Niña/El Niño phases of ENSO cycles (Goddard and Graham 1997; Mantua et al. 1997). Although the phase transitions of PNA and PDO may differ slightly in space and time, both events persist for decades and produce the most visible climate signatures in the North Pacific (Francis and Hare 1994; Minobe 1997). The El Niño dominated phase of the PDO cycle is characterized by a weakening of the trade winds that results in an eastward movement (slosh) of the warm pool of equatorial water normally contained in the western Pacific by the trades during La Niña conditions (Cole et al. 2000). The sequence of events leading to and associated with the warm



Climate Patterns in the Coastal Zone, Fig. 2 Storm track enhancement (arrows) associated with 700 mb atmospheric pressure anomalies during (a) La Niña and (b) El Niño dominated climate patterns. (Modified from Inman et al. 1996; pressure data from Redmond and Cayan 1994)



Climate Patterns in the Coastal Zone, Fig. 3 Schematic illustration of sequences leading to a fully developed El Niño event. Numbers refer to (1) warm pool of water setup by trade winds, (2) eastward “slosh” (baroclinic Kelvin wave) of warm water released by relaxation of trade

winds, (3) warm pool travels to eastern Pacific and (4) spreads poleward along continental margins (trapped barotropic Kelvin waves), and excites (5) western traveling planetary (Rossby) waves. D, T, and H designate Darwin, Tahiti, and Hawaii. (Modified from Inman et al. 1996)

(El Niño) phase of the PDO is illustrated in Fig. 3. It has been suggested that the breakdown from a prevailing La Niña to an El Niño occurs when brief bursts of westerly winds near the dateline begin the easterly transport of warm water. The positive feedback from the warm water further decreases the trade winds, beginning the transformation to an El Niño (e.g., Fedorov and Philander 2000).

The stronger trade wind systems during the cool (La Niña) phase of PDO are part of a general spin-up of the atmospheric circulation, which causes the North and South Pacific Gyres to rotate faster. Both effects (wind and current) induce upwelling that maintains cold-water masses along the west coast of the Americas, which sustains the typically cool dry coastal climate of these regions during the La Niña dominated periods of the PDO and PNA. In contrast, the strengthened trade winds of the La Niña period are associated with higher rainfall on the windward sides of Hawaii and other high islands in the trade wind belt. Higher than average rainfall occurred at Hilo and Lihue in the Hawaiian Islands during the La Niña dominated period 1943–68, and lower average rainfall during the subsequent El Niño dominated period

1969–98. However, the El Niño period included episodic Kona storms and the occurrence of hurricane Iniki.

The NAO is an atmospheric dipole that retains its polarity with a low over Iceland and a midlatitude high that enlarges and contracts (Wallace and Gutzler 1981; Hurrell 1995). The Icelandic low remains more or less in place while the midlatitude high-pressure anomaly changes in size, strength, and latitude (Fig. 2). During La Niña the high-pressure anomaly is stronger and centered over the Bay of Biscay and Spain; during El Niño the high expands but weakens and is centered around 30° North Latitude over the eastern North Atlantic and North Africa. Figure 2 shows that during La Niña events, storm tracks end near the Black Sea causing droughts in Israel; during El Niño the storm tracks end in Israel bringing rain and high waves. The biblical 7-year floods on the Nile River are associated with La Niña intensification of the southwest monsoon over the Indian Ocean (Fig. 1). These monsoons bring heavy rainfall over the high Ethiopian plateau that induces flooding on the Nile River.

The dipole systems of the PNA and NAO redirect the tracks of traveling storm fronts, causing changes in coastal wave

climate. The strongest pressure gradients occur along the boundaries of the pressure anomalies of the PNA and NAO patterns and these boundaries indicate the prevailing storm paths. Figure 2a (La Niña) shows large areas of high pressure over the North Pacific and North Atlantic. These high-pressure areas enhance storm tracks as shown by the arrows for waves approaching North America and Europe, with the prevailing North American storm tracts directed at the Pacific northwest and along the Atlantic east coast. In the eastern Pacific, the principal wave energy during La Niña dominated climate (1943–77) was from Aleutian lows having storm tracks, which usually did not reach southern California. Summers were mild and dry with the largest summer swells coming from very distant Southern Hemisphere storms. The wave climate in southern California changed with the El Niño dominated weather of 1978–98. The prevailing northwesterly winter waves were replaced by high-energy waves approaching from the west or southwest (Fig. 2b), and the previous Southern Hemisphere swells of summer were replaced by shorter period tropical storm waves during late summer months from the more immediate waters off Central America. The east coast of North America is out of phase with the west coast for severe weather. Heavy rainfall and numerous, powerful Caribbean hurricanes and northeasters occur typically during La Niña years (Fig. 2a) (see entry on “► [Energy and Sediment Budgets of the Global Coastal Zone](#)”).

Cross-References

- [Coastal Climate](#)
- [Coastal Currents](#)
- [Coastal Temperature Trends](#)
- [Coastal Upwelling and Downwelling](#)
- [El Niño–Southern Oscillation \(ENSO\)](#)
- [Global Vulnerability Analysis](#)
- [Meteorological Effects on Coasts](#)

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Coastal Barrier Preservation and Destruction

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Barrier Destruction During Transgressive Stages of Last Eustatic Cycle

During transgressive phases of the eustatic cycle, the time interval between two consecutive eustatic highstands, the littoral

barrier is either translated continuously landward by a “tank-tread” or “rollover”-style overwash mechanism or at least partially destroyed during submergence beneath the landward-shifting surf zone – the advancing ravinement. In assuming that barrier spits on the shores of extinct Lake Bonneville in Utah, a late Pleistocene pluvial fresh water lake, were destroyed when wave erosion, cutting through spits interrupted littoral drift that previously has nourished them, Gilbert (1885) took the first step toward recognizing this process. Prompt reestablishment of the lake shoreline at a position further landward may have followed spit destruction. In his very brief remarks on the subject, Gilbert did *not* mention an associated lake level rise linked to the assumed cessation of littoral drift and elimination of the barrier spit. However, much later and based on the presence of relict brackish-lagoonal deposits that underlie a veneer of modern ocean sediments off Long Island, New York, Sanders (Sanders 1963) played a pioneer role in recognizing that barrier “overstepping” and landward-shifting process did take place between ~8.0–7.0 ka BP on the Atlantic inner shelf during a rapid phase of early Holocene sea level rise. Caused by reduction in sand supplies that until the critical stage did maintain shore ridges above sea level, barrier island submergence was accompanied by intensive ravinement erosion. Within a matter of centuries, the surf zone moved from ~7 km to only ~2 km from the present Long Island shoreline. Leaving barrier remnants behind on the ocean floor, the overstepping surf zone occupied the landward margin of the old lagoon. It was marked by the steep face of late Pleistocene deposits. During the past 5–6 ka, the shoreface has completed landward retreat to its current position (Rampino and Sanders 1981).

The continental shelf off certain sectors of the Frisian shore between Holland and Denmark experienced similar transgressive processes. Starting at a great distance off the present coast, the coastal barrier system, established by ~8 ka B.P., rapidly transgressed landward until ~5.5 ka B.P. The barrier zone was stabilized between 5.0 and 4.5 ka B. P. (Fruergaard et al. 2015). Beachrock (“hardground”) and eolianite cementation hindered erosion of submerged sandy littoral ridges near Sardinia (DeFalco et al. 2015). Cemented sand ridges mark abandoned shoreline positions at 100 m and 60 m depths on the narrow Indian Ocean shelf of South Africa (Salzmann et al. 2013). Submerged rocky headlands separate the drowned sandy coastal sectors. Evolution of the South African barrier complex, from formation to drowning may have taken only ~1 ka. Each of the overstepped barrier ridges formed within a few centuries. Steep antecedent shelf gradient fostered ravinement erosion of the multiple relict strandline landforms. Erosional removal of shoreline lithosomes coincided with rapid sea level rise and subtropical diagenesis (Salzmann et al. 2013). Submerged erosion-resistant gravelly barrier remnants and their occasional landward migration over the seafloor were reported by Forbes et al. (1991), Hartstein and Dickinson (2000), and Mellett et al. (2012). However, at certain locations, such as the southeastern

coast of Brazil (Cooper et al. 2016) where ocean wave-exposed unconsolidated barrier lithosomes tend to offer little resistance to surf erosion and ravinement, local sheltering may occasionally protect overstepped fossil barriers against total destruction. In general, a very gentle seafloor slope, small tidal range, low wave energy, and decreased sediment supply; breakdown in the longshore drift system and anchoring of large barrier sand volumes between bedrock “shoulders”; and above all rapid sea level rise resulting in increased accommodation spaces contributed to preservation of drowned barrier ridges. Slow sea level rise and low and/or moderate land subsidence associated with sufficient sand and/or gravel supply to the barrier system, results in storm overwash-driven continuous ridge retreat without permanent submergence under the ocean.

Episodes of regionally synchronous and rapid sea level rise are often related to global meltwater pulses. Five overstepping-flooding events, two contemporaneous between different sites were recorded between 9.6 and 6.8 ka B.P. in the east Texas continental shelf (Anderson et al. 2014, 2016). As rising sea level expanded the mid- and late Holocene lagoons landward, a 30–50 km wide lagoonal estuarine system developed in the rear of protective barrier chains. Traces of several drowned littoral ridge sets, submerged in the course of the Holocene transgression were noted across the northern shelf of the Gulf of Mexico (Anderson et al. 2014; Kindinger et al. 1994). Utilizing high-resolution multibeam echosounder surveys across the outer continental shelf and upper slope on the northeastern Gulf, Gardner et al. (2005, 2007) presented images of several drowned fluvial deltas and linear barrier remnants on the seafloor.

Barrier Island Erosion and Elimination Processes: Late Holocene-Recent Highstand

Erosion of Nondeltaic Barrier Islands

The very moderate, most recently accelerating “highstand” sea level rise since 4 ka B.P., the frequently recurring major tropical and winter storms, and human interference with the littoral and fluvial sand delivery system may help to explain why many barrier sectors on the U.S. Atlantic and Gulf coast are experiencing pronounced recession. Once the landward moving shoreline bypassed offshore sand sources, sediment supply from the transgressive ravinement has declined (Anderson et al. 2014). Thus, updrift sectors of Mississippi–Alabama barriers suffered particularly intensive recession, exacerbated by the more recent, wholesale reduction of the long, narrow, and low barrier spit-like sectors of Dauphin, Petit Bois, and Ship Islands to subtidal levels by tropical cyclonic storm erosion during Hurricanes Camille (1969) and Katrina (2005). Extreme storm-induced geomorphic changes also caused profound changes in the plant communities of the islands (Carter et al. 2018). Storm passes and cuts frequently opened up across very narrow central and western Dauphin Island. Due to recurrent storm cuts

and their subsequent healing between 1882 and 1928, the island's length has fluctuated between 6.0 and 21.6 km. Petit Bois Island and Petit Bois Pass, east of the island underwent similar rapid changes in width (Otvos and Carter 2008, 2013). In addition to storm destruction of its supratidal portion, the near-demise of East Ship Island during Hurricane Katrina in 2005 in part may be also attributed to the naturally altered tidal dynamics within the Dog Island Pass system. The Pass became a sizable sink for the westward-directed longshore sand flow (Twichell et al. 2013). However, judging by the permanence of the adjacent shoal belt, supratidal sand loss from the Mississippi-Alabama barrier system apparently exceeds the sand loss suffered from the more stable subtidal shoal platform lithosomes that underlie and laterally adjoin the islands.

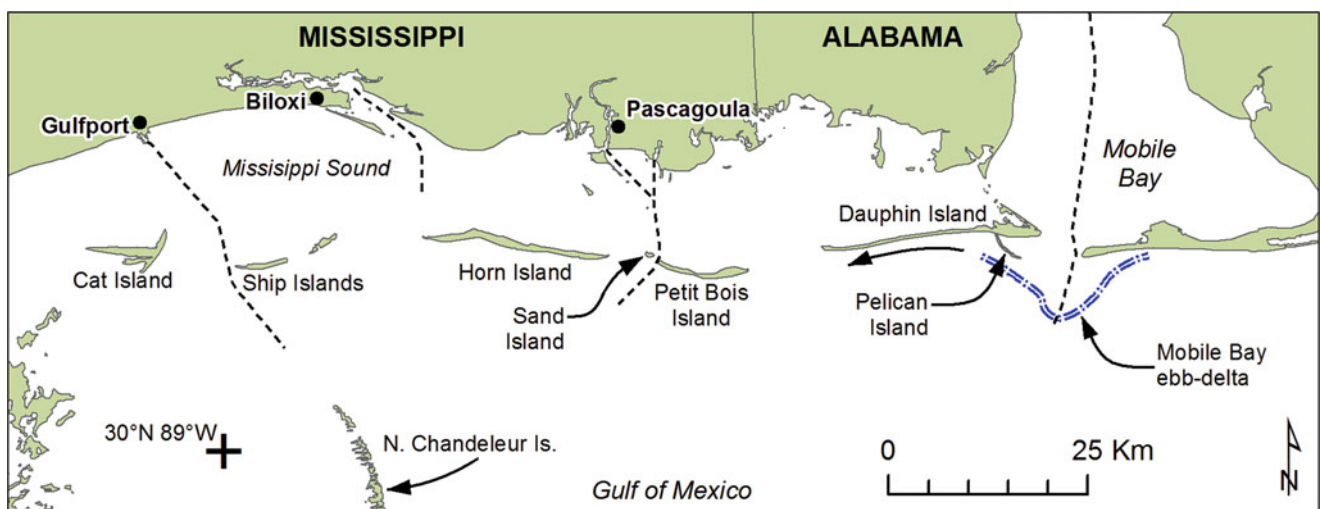
Trunk channel avulsion in the eastward prograded Mississippi River delta prior to ~ 2 ka B.P. resulted in expanding new delta lobes. Delta expansion resulted in elimination of the western one-quarter of the original Louisiana-Alabama barrier island chain (Fig. 1; Otvos and Giardino 2004). Delta lobe formation initially interrupted littoral drift, first by shoaling the adjacent seafloor, then by encroachment of the subaerial delta plain on the barrier islands. Finally, delta sediments buried a number of barrier islands. Ending further progradational growth of Cat Island, this process permanently severed the island-nourishing westward-directed littoral sand transport from West Ship Island.

Erosion of Mississippi Delta-Related Barrier Islands

Numerous late Holocene barrier islands that formed during the last ~ 2 ka in the periphery of the Mississippi Delta Complex have been impacted by intensive wave destruction coupled with significant sediment compaction-associated subsidence. These islands are composed of thin, fine-sandy deposits that render them exceptionally vulnerable to wave

erosion. The cumulative area of the transgressive Chandeleur chain, ~ 73 km long and covered by fields of extensive, medium-high sand dunes in mid-nineteenth century, following devastation by Hurricane Katrina and lesser tropical cyclonic storms, was reduced from 44.5 to 4.7 km² (Fearnley et al. 2009; Otvos 2018). Small, highly fragmented island remnants, mostly in the north, with exception of small new dune fields, barely reached above sea level in the years immediately following Hurricane Katrina. Delta islands were eliminated not by “submergence” or “drowning,” a process that may leave part of their lithosomes intact on the seafloor, but by in-place, *in situ* destruction by storm and also fair-weather wave erosion, including subsequent ravinement near present sea level. The repeated destruction of the Isle of Caprice (Dog Keys) between Horn and Ship islands, Mississippi, following its repeated emergence, was one non-deltaic example of this method of island elimination (Otvos 1981).

Despite its remarkable recent regeneration (Otvos 2018; Encyclopedia entries ► “Barrier Island Formation and Development Modes”; ► “Gulf shorelines, Last Eustatic Cycle”), the decades of the Chandeleurs as a major island chain on the long run may be numbered (Fearnley et al. 2009; Rogers et al. 2009; Flocks et al. 2009, 2015). During and immediately following Hurricane Katrina the chain lost most of its stabilizing intertidal marsh to erosion and overwash. It was estimated that ~ 0.2 km³ sand volume, comparable to the amount used in the construction of a (soon failed) protective sand “berm” ridge after a disastrous oil spill, would be required to restore and maintain for 50 years the remaining islets at the pre-Katrina (2005) position of former North Chandeleur Island (Moore et al. 2014). Depending on storm climate pattern and sea level rise in coming decades, such nourishment may or may not be adequate and/or economically affordable even in a greatly regenerated island chain.



Coastal Barrier Preservation and Destruction, Fig. 1 Mississippi–Alabama barrier island chain. (Dashed line: navigation channels). Sand Island was renamed “West Petit Bois Island”.

The island fragments, in fair weather periods slightly accreting, eventually will become a sandy subtidal shoal zone, driven westward by the rising sea level and steadily advancing Gulf surf over the delta-lagoon complex. Continued compactional subsidence of the delta region, combined with an assumed increase in the rate of global eustatic sea level rise, characterizes this local transgression.

Summary

Barrier island overstep, coupled with ravinement erosion of the submerged islands, along with flooding surface development landward, characterized episodes of sudden eustatic sea level rise and localized increases in RSL relative sea level. A distinction must be made between the natural processes of island elimination primarily by drowning (submergence) and the in situ erosional destruction of islands at current sea level. As the shoreface rapidly shifts landward and the arcuate barrier chain that protected the lagoon collapses, open marine sediments settled over previously deposited sediments of wide late Holocene lagoons and bays. A very gently inclined seafloor, small tidal range, low wave energy, and decreased sediment supply; breakdown in longshore drift system and anchoring of large barrier sand volumes between bedrock “shoulders” (Cooper et al. 2016), and above all rapid sea level rise resulting in increased accommodation spaces contributed to preservation of drowned barrier ridges. Slow sea level rise, accompanied by sufficient sand and/or gravel supply to the barrier system results in storm overwash-driven continuous ridge retreat without ridge drowning. Fragile deltaic barrier islands, composed mostly of fine-grained sand are the most susceptible to storm erosion. Burial under advancing delta lobe has also played a role in partial extinction of an island chain. Accelerating sea level rise, insufficient littoral sand supply, anthropogenic alterations of the coastal and fluvial systems are among factors that contribute to erosion during the recent marine highstand phase with minimal sea level rise since, ~4 ka B.P. Late Quaternary barrier islands in the Gulf of Mexico offer several good examples of island emergence and destruction. (*q.v.* Encyclopedia entries ► “Beach Ridges”; ► “Barrier Island Formation and Development Modes”; ► “Gulf shorelines, Last Eustatic Cycle”).

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Coastal Boundaries

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Synonyms

Coastal borderlines; Coastal borders; Coastal frontiers; Coastal limits; Coastal margins; Delimiting boundaries

Definition

Coastal boundaries delimit a geographical space and are defined by the distances of one or more areas on the coast, the sea or on its interface, on the order of a few hundred meters to many kilometers, average. Coastal boundaries can also extend and go inland by land, from the interior of the watersheds to the limit of a national jurisdiction into the sea. Most criteria for establishing coastal boundaries depend on several factors such as the set of specific problems, the uses of the territory, or the physical and geographical aspects that are relevant to each stretch of coast. Within it, the interaction between sea, land, and atmosphere is produced differently by the occurrence and direct effect of natural events or processes of anthropic origin. Inside coastal boundaries, exclusive forms of fragile ecosystems will be developed, and specific economic, social, and cultural relations are present. Sometimes, because of the complexity and drastic changes that occur in this space, coastal boundaries need to be systematically monitored to favor proper planning and supervision of the coastal marine space and its integrated management. When establishing limits in coastal municipalities, special attention should be paid to the local social component that interacts with the coastal resources.

Introduction

Coastal boundaries all over the world show a great variety of ranges (Garbert 2005). According to the management and scientific points of view, the extent of boundaries varies in relation to the nature of the management issue (Kay and Alder 1999). Some authors of various countries take into account different aspects to determine coastal boundaries: the physical–natural criteria (Barragan 2003), the impact of natural processes (Milanés et al. 2017), and the coastal morphology and human coastal uses in marine and land based (Balaguer et al. 2008). Some nations establish political–administrative limits for coastal boundaries (GORC 2000; MMA 2001), while others take into consideration the climate change scenarios, arbitrary distances, or other different aspects. That is why the creation of international laws and methods as specific ways determining boundaries is so important and has increased in popularity among different countries.

A critical review of the legal and methodological frameworks for establishing coastal boundaries in Integrated Coastal Zone Management (ICZM) and land-use planning (L-UP) is carried out further down, since its establishment is based on different criteria or areas of various disciplinary approaches. Three important comparative aspects were taken into account for the analysis of the establishment of coastal boundaries: (1) technical terms, conceptual approaches, and existing legal frameworks, (2) different criteria used to define coastal boundaries, and (3) current methodologies and guidelines of ICZM programs and L-UP instruments. Different distances established for coastal boundaries are also discussed, as well as essential terms and legal frameworks to define coastal boundaries.

Essential Terms, Conceptual Approaches, and Legal Framework to Define Coastal Boundaries

There is a great variety of technical terms in the scientific literature for establishing coastal boundaries, such as “shore” (Oertel 2005; McKechnie 1979), “shoreslines” (PNUBCL 1994; GORC 2000; Oertel 2005; Castro and Alvarado 2009), “coastline,” “sea level” (Oertel 2005), “neap tides” (Teschemacher v. Thompson 1861), “coastal zone” (Hildebrand and Norrena 1992; Andrade and Castro 1997; Pujadas and Font 1998; Schlotfeldt 2005; Andrade et al. 2008), “zona de influencia costera” (Pujadas and Font 1998; Henríquez et al. 2011), “plataforma insular” (GORC 2000), “edge of the continental” (French 2005), “administrative boundaries” (GORC 2000; French 2005), “submerged lands boundary,” “offshore boundaries,” “tidelands,” “Outer Continental Shelf” (Grabert 2005), “uplands” (Shalowitz 1962; Graber 1980, 2005), “federal navigational servitude” (Graber

1980, 1981, 2005), “territorial sea” (CTSCZ 1958), “contiguous zone” (CTSCZ 1958; Graber 2005; Law of the Sea Convention 1983), and “exclusive economic zone” (Graber 2005), among others. Table 1 and Figs. 1 and 2 show some specific terms and descriptions used in an interchange carried out by members of the scientific community and policy makers to establish world coastal boundaries (see Table 1).

It is very important to be aware about terminological contradictions present in many of the definitions of coastal boundaries since some of the expressions used are not clearly explained and tend to be confusing. Furthermore, the meaning of some words is changed using synonyms, as well as the limits and distances of different coastal nations. For example, in 1935, the US Supreme Court updated the term “mean high-water line” with the counterpart term “line of mean high water” (Graber 2005). The technical terms low-water line (Shalowitz 1962; Maloney and Ausness 1974; Graber 1980, 2005) and mean high water (MHW) are sometimes called foreshore (Oertel 2005; French 2005). Occasionally, the term “tidelands” is also called foreshore, or wet-sand area.

In *Borax*, the Court equated the common-law term “ordinary high-water mark” with the technically precise phrase “line of mean high water.” “Boundary” is also referred to as “mean high-water line” and “mean high-tide line” (*Borax, Ltd. v.* 1935; Graber 2005). Something similar occurs with the three term indistinctively used to determine the coastal influence boundaries: “zone of coastal influence,” “zone of indirect influence,” or “zone of direct influence.”

Specific policies to establish marine and terrestrial boundaries with the aim of planning or managing coastal zones are present in this context (Milanés and Pérez 2012; Botero et al. 2016). In this international legal frameworks, many important terminological contradictions can also be observed. Some normatives use technical terms like: “delimitation of coastal zones” (GORC 2000), “demarcation of marine terrestrial zones for public use” (LEC 1988; IGN 2004), “coastal delineation” (Steer et al. 1997), “delineation of coastal bordering” (Martins 2010), “delimitation of technical brands” (Barragán 2003), and “area for public use” (Macias 2000). Additionally, the terms “restriction zone,” “exclusion zone,” and “protection belt” are used in countries like Brazil, Poland, and Denmark.

In Cuba, new concepts such as “Zones Under Integrated Coastal Management Regime” (Salabarría and Brito 2011), “Primary Environmental Coastal Unit for Integrated Management (PECUIM),” and “Basic Environmental Coastal Unit for Management and Land Planning (BECUIMLUP)” (Milanes et al. 2017) are being used for demarcation and delimitation of coastal zones in ICZM and L-UP. Meanwhile the terms “integral units for planning and land-environmental spatial planning,” “coastal and/or oceanic environmental units” (also called “coastal environmental unit”) and “integrated management unit” (MMA 2001; CCO 2007; MADS 2013) are used in significant coastal policies in Colombia. In

majority of cases, definition of “homogeneous units” was done by using map-crossing with geographic information systems.

The term “demarcate” is used to refer to the process of locating a boundary on the ground, (Graber 2005). According to Milanés (2014) and Milanés et al. (2017), it represents the coastal zone boundary of the selected area for applying ICZM programs. Graber (2005) also used the technical terms “delineating” for showing on a map public–private coastal property boundaries and “define” to refer to the legal definition of a coastal boundary.

“Delimitation” represents the marine–land boundary for applying L-UP (Milanés 2014; Milanés et al. 2017). Furthermore, “delimitation” is an interdisciplinary exercise carried out with the purpose of providing a cartographic representation of the territory, on which the planning and subsequent management are to be implemented. Also “zoning” is used when delimiting coastal boundaries (CicinSain and Knecht 1998). According to these authors, zoning is the process of ordering the spaces that make up the coastal edge of the coastline. A broader view of the same term is offered by Henriquez et al. (2011) who stressed that it is a participatory process with the aim of establishing multiple uses in the territory as well as those restrictions for its administration. This exercise is displayed in one or several maps to facilitate the decision making of the government as well as private inhabitants who take specific actions on the coastal edge.

As some concepts and terms used to define coastal boundaries are not greatly different, they can lead to misunderstanding and confusion when used by academicians, scientists, and other specialists. Therefore, continuous and clear explanations are necessary for a better understanding of the real meanings the different terms convey. Similarly, each technical term gives specific distances when referring to coastal boundaries at land and sea. Table 2 shows some of the terms used in the coastal policies, ordering plans, laws, and decree laws reviewed (see Table 2).

As Table 2 showed, particular coastal policies define boundaries establishing a range from la línea de base recta in land up to la zona económica exclusiva de 200 nautical miles in sea (LEC 1988); other legal frameworks take into consideration the different types of coasts (GORC 2000). “Protection zone,” used in Cuban Coastal Decree-Law 212, refers to land ordering or planning of coastal zones considered different limits, such as mangrove swamps, lagoons, beaches, low-lying sea coast, and hard rocky or cliff-lined coast, among others (see Fig. 3a–d).

Criteria Used to Establish Coastal Boundaries

The first serious analyses about the different criteria used for coastal zones boundaries were carried out during the 1990s.

Coastal Boundaries, Table 1 Technical terms and corresponding descriptions scientifically and legally used for coastal boundaries

Terms	Descriptions
Shore	A line, not a region (see Fig. 1) Area adjacent to a water body (Oertel 2005) Land near or at the edge of a water body (McKechnie 1979)
Shorelines	Portion of the territory comprising fiscal beaches, beaches in general, bays, gulfs, straits, and inner channels, including also all area of the national seawater Portion of the coastal zone with jurisdictional-administrative boundaries defined by the corresponding norms and comprising all territorial beach spaces (PNUBCL 1994) Reduced zone restrained to special regulations, not giving possibilities of doing systematic analysis and planning of all coastal zone Restricted coastal planning space useful for sectorial regulations regarding the use of littoral resources (Henríquez et al. 2011) Area not large enough to be used for territorial ordering of the coastal zone (PNUBCL 1994; Castro and Alvarado 2009) Line demarcating the edge of a water body Line of coincidence between the coast and the mean sea level (GORC 2000) Specific boundary located between land and water, generally longer than the associated coastline length, line demarcating specific boundary between shore and water (Oertel 2005)
Coastline (also known as coastal edge)	Approximate boundaries at relatively large spatial scales (see Fig. 2) Line connecting the most extended promontories into the sea (Oertel 2005)
High-water line; high-water mark; or line of mean high water	A federal rule of the United States defines the public-private boundaries in most of the states, although some states use the term mean low-water line or other lines (Shalowitz 1962; Maloney and Ausness 1974; Graber 1980, 2005) Area comprising the upper zone of the beach, up to where the waves reach due to the high tide, exceeding, therefore, the land line (PNUBCL 1994) The 18.6-year mean of all high-tide elevations is called mean high water (MHW) (Oertel 2005) Intersection of the mean high water (MHW) with the shore Boundary determined following the California rule: using only the high “neap tides” (Borax, Ltd. v. 1935; Graber 2005) Boundaries determined by the coast agency and geodetic survey (now named NOS). Data based on 18.6 years and now about 19 years (Graber 2005)
Low-water mark or low-water line	Intertidal zone, ignoring all seaward areas Only useful where marine-based activities are of minor importance (French 2005 p. 317) The 18.6-year mean of all low-tide elevations is called mean low water (MLW) (Oertel 2005 p. 323) In the United States, six Atlantic coastal states do not follow the English common-law rule and use the mean low-water line or variations like public-private coastal boundary (Maloney and Ausness 1974; Graber 1980, 2005)
Foreshore	Portion of the shore where wave energy dissipates on the land Portion sculptured by dispersion and shifting of granular particles due to the wave energy and littoral currents. The resultant patterns of erosion and accretion turn the foreshore in one of the most dynamic areas of the shore (Oertel 2005 p. 325)
Backshore	Landward portion of the shore primarily molded by aeolian processes (Oertel 2005)
Sea level	Generally defined as the highland low-water elevations and the generic “shoreline,” rather than with the high or low tide (Oertel 2005 p. 325) Not constant elevations when high and low tide Changes in the magnitudes of gravitational and centrifugal forces pulling on the water surface Changes in the gravitational forces resulting from motion and change of the relative sun, moon, and earth alignments The 18.6-year mean of all sea-level elevations is called mean sea level (MSL) (Oertel 2005)
Coastal zone	Zone of hundreds of meters or kilometers, extended from inland coastal watersheds to the limits of national jurisdiction in the offshore. Its definition depends on set of issues and geographic factors relevant to each stretch of coast (Hildebrand and Norrena 1992) (see Fig. 2) Area of defined width and limited length for adequate planning ordering and comprising 12 miles of territorial sea and inland, including all ecosystems, landscapes adjacent micro basins, systems of coastal wetlands, as well as all important cultural recreational, social, and economic area having interdependent littoral functions (Henríquez et al. 2011) Zone having specific terrestrial and marine characteristics Geographical space of variable length in contact with the lithosphere, hydrosphere, and atmosphere having, therefore, special characteristics regarding richness, fragility, and complexity (Andrade et al. 2008) Interphase where processes of high physico-biological dynamism occur, resulting in special characteristics of high fragility and in usual valuation (Andrade and Castro 1997; Schlotfeldt 2005; Andrade et al. 2008) Zone with inland boundaries established according to the structure and configuration of the different types of coasts. The external boundary at sea coincides with the edge of the insular platform of the territory, normally 100–200 mts. Depth (GORC 2000) Boundary established with a minimum of 500 mts. Inland, and defined according to the line of average high tide. At the marine zone, the maximum boundary is 12 nautical miles (GO 2001)
Zone of coastal influence. Also called	All terrestrial or marine space influenced by some coastal or non-coastal activities having positive or negative and direct or indirect impact on coastal zones (see Fig. 2)

(continued)

Coastal Boundaries, Table 1 (continued)

Terms	Descriptions
“indirect zone of influence” or “zone of direct influence”	In some cases, the zones are determined by the territorial functions and problems (Pujadas 2008) Spatial definition length and width varying according to the territorial function or problem object of analysis, coinciding, in some cases, with coastal zone definition, but not in others (Henríquez et al. 2011) The zone demarcation may include definition of hydrographic basin or subbasin associated to the coast and influencing on the sea contamination (Milanés 2012, 2014; Henríquez et al. 2011)
Insular platform	Marine bottom not very deep comprising the coastline and the deep change of the external edge (GORC 2000)
Edge of the continental	Geological limit varying among nations (French 2005) Most of the sea area included, covering some fisheries and sediment stores. Sediment is frequently dredged for coastal recharge
Administrative boundaries	Used internationally and very useful. There are various types of distances, for example, three in the United Kingdom: (1) 3 nautical miles, limit of controlled waters; (2) 6 nautical miles, limit of exclusive fishing rights; (3) 12 nautical miles, limit of territorial waters (French 2005, p.317)
Submerged land boundary	Lands below the mean low-water line (Graber 2005) The Submerged Lands Act, 67 Stat. 29 (1953), proposed not extending the coastline of any state more than 3 geographic (nautical) miles (one marine league) in the Atlantic and Pacific Oceans or more than 9 geographic (nautical) miles (three marine leagues) in the Gulf of Mexico (Submerged Lands Act 1953; Graber 1981, 2005). The law had the general effect of solving the federal–state boundary at either 3 or 9 miles offshore
Offshore boundaries	The extent of the territorial sea, the contiguous zone, and the exclusive economic zone (Grabert 2005)
Uplands	Lands above the mean high-water line Sometimes considered as littoral lands or fastlands. According to Grabert (2005), the owners of those lands are littoral owners (see Fig. 1)
Tidelands	Lands between the mean high-water and mean low-water lines (see Fig. 1) Sometimes described as lands over which the tide ebbs and flows Adjacent to uplands on the landward side and to submerged lands on the seaward side (Graber 2005)
Federal navigational servitude	Federal government paramount authority to control and regulate the navigable waters of the United States under the commerce clause (Graber 1980, 1981, 2005)
Outer continental shelf	In the legal framework, this term refers to the submerged lands subject to federal jurisdiction immediately seaward of the 3- or 9-mile belt of submerged lands owned by the states (Graber 2005) The congress of United States passed the Outer Continental Shelf Lands Act authorizing the federal government to lease oil, gas, and other mineral resources within the area of submerged lands between the seaward limit of the states’ submerged lands and the edge of the continental shelf (Outer Continental Shelf Lands Act 1973; Graber 2005)
Contiguous zone	Zone of the high sea adjacent to the territorial sea Zones where coastal nations are authorized to prevent infringement of their customs, fiscal, immigration, and sanitary regulations (CTSCZ 1958; Graber 2005) This zone does not extend more than 24 miles beyond the baseline used for measuring the width of the territorial sea under international law (Law of the Sea Convention 1982)
Exclusive economic zone	200-mile-wide exclusive economic zones where coastal nations can explore, exploit, and manage living and nonliving resources (Garbet 2005) (see Fig. 1) Usually comprising 200 nautical miles (371 km) seaward (Inman and Jenkins 2005)

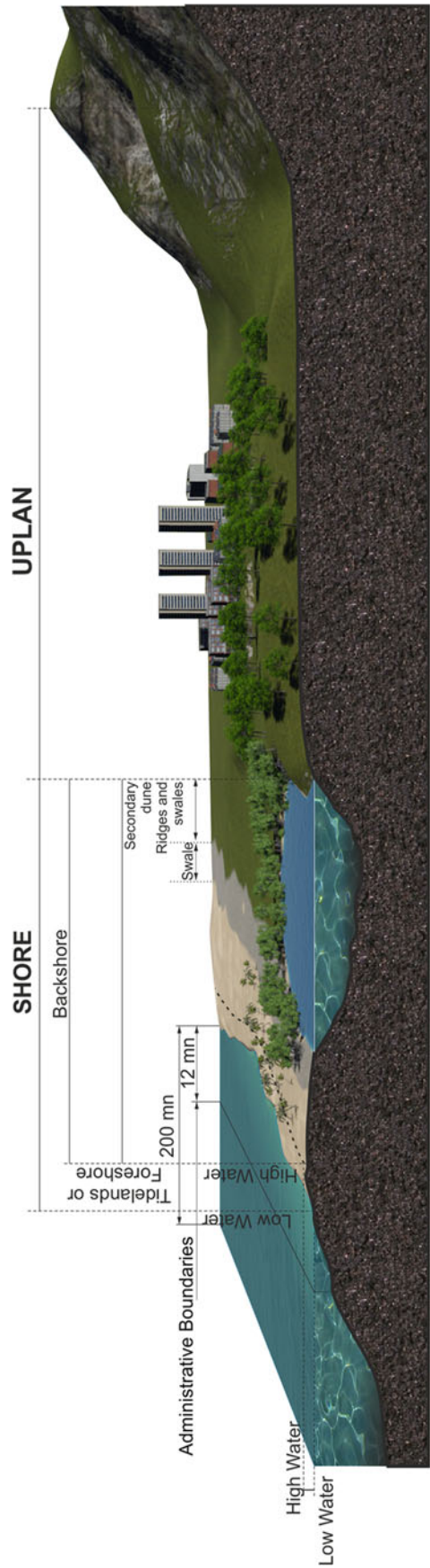
The first American law of coastal management proposed to divide the different criteria into four categories: (1) arbitrary criteria, (2) physical–natural criteria, (3) economical–organizational criteria, and (4) jurisdictional and administrative criteria (CZMA 1972). Biological criteria are also used to define limits (Kay and Alder 1999; Abogado and Mendez 2003), as well as the historical memory of past natural events (Barragán 2010; Milanés 2014; Milanés et al. 2017). Table 3 showed the most representative criteria used for coastal zone boundaries.

For effective establishment of coastal boundaries, the criteria shown in Table 3 must be taken into consideration in a flexible and holistic way. When establishing boundaries at land and sea, the best thing to do is to overlap each of those criteria by several cartographic layers and in this way to get a coherent and effective delimitation of the corresponding areas.

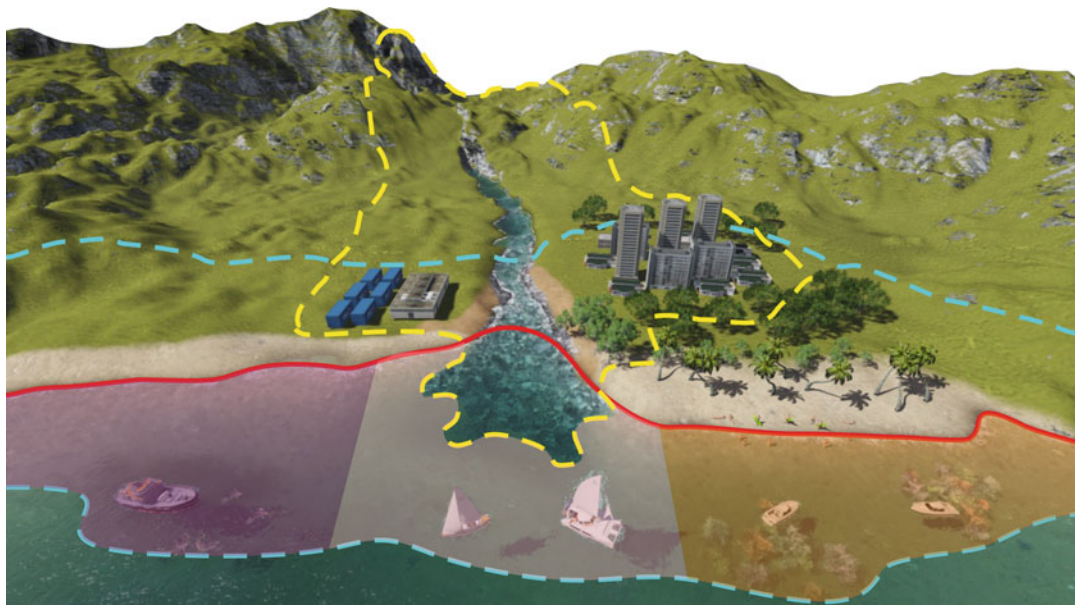
It is equally important to know the purpose of the task to be carried out before starting the establishment of boundaries.

There are occasions in which a single criterion is useful enough to solve a particular case. Whatever criteria adopted, establishment of coastal boundaries must fulfill the following aspects: (a) clear and comprehensible delimitation, with possibilities of being cartographically represented; (b) evidences of the coastal zone reality and its natural, economic, social, cultural, political, etc. factors; (c) areas of environmental and economic importance; (d) development tendencies of the corresponding areas, as well as the amount of resources; (e) solution to the multiple use–use and use–resource problems (Abogado and Mendez 2003; Milanés 2014).

To get homogeneity and functionality when doing delimitation, the most effective way, among so many options, is to take into account the natural features and the socioeconomic



Coastal Boundaries, Fig. 1 Cross section showing the technical terms of a coastal territory in the marine and terrestrial parts (Modified from Oertel 2005)



Coastal Boundaries, Fig. 2 Aerial section of an area showing the coastal classification for L-UP and ICZM. Legend: ——— Zone of coastal influence. ——— Coastline or coastal edge. ——— Coastal zone boundary

criteria, trying to adjust the boundary proposal to the territorial political-administrative division (Henríquez et al. 2011). There is not a single criterion with the possibilities of being universally used in isolation for establishing coastal boundaries; neither all requirements to get effective delimitation of coastal zones can be fulfilled. Sometimes, using only one criterion would mean simplicity. Therefore, the use of several criteria is required to get realism and reliability when establishing coastal boundaries. The holistic approach followed when doing boundary delimitation results in getting different spatial and functional dimension of the coastal zone with every criterion used. Different coastal boundaries can be established throughout time, depending on the criteria and methods used when developing those tasks.

Current Methods and Tools Used for Establishing Coastal Boundaries in ICZM and L-UP

Establishment of coastal boundaries depends on the purpose of the task, for example, (1) ICZM programs (GESAMP 1996; Kay and Alder 1999; Planas 2012; Clark 1996); (2) land-use planning (MMA 2004; Padrón 2002; Pantaleón 2001); (3) marine spatial planning; (4) management programs of protected areas at land and sea, including their different categories (Gerhartz et al. 2008); (5) environmental planning (Obllurys et al. 2009); (6) zoning of uses and risks (Ochoa et al. 2000; Salzwedel et al. 2002; Abogado and Méndez 2003; Henríquez et al. 2011); (7) river basin management (Planas 2012); (8) defining scenarios of danger, vulnerabilities, and risks (AMA 2014); and (9) combined purposes when defining management coastal areas, zones for territorial ordering, protected areas, and hydrographic basins (Milanés 2014).

There is noticeable absence of works aimed to establish proactive methods to define coastal boundaries for Integrated Coastal Zone Management initiatives (Balaguer et al. 2008). Nevertheless, depending on the interest and availability of data, some countries have defined coastal boundaries in different ways for the same program. Specific methodologies are focusing their analysis on delimitation of coastal environment for ICZM and L-UP (see Table 4).

Comparative synthesis of ICZM and L-UP tools and procedures showed several differences when establishing coastal boundaries. Some methods and procedures propose the inclusion of all coastal resources and activities of interest (WB 1993; Salabarría and Brito 2007; Planas 2012; GESAMP 1996); however, other institutions do not make any reference to systems or approaches to establish coastal boundaries (PNUMA 1992; WCCR 1996).

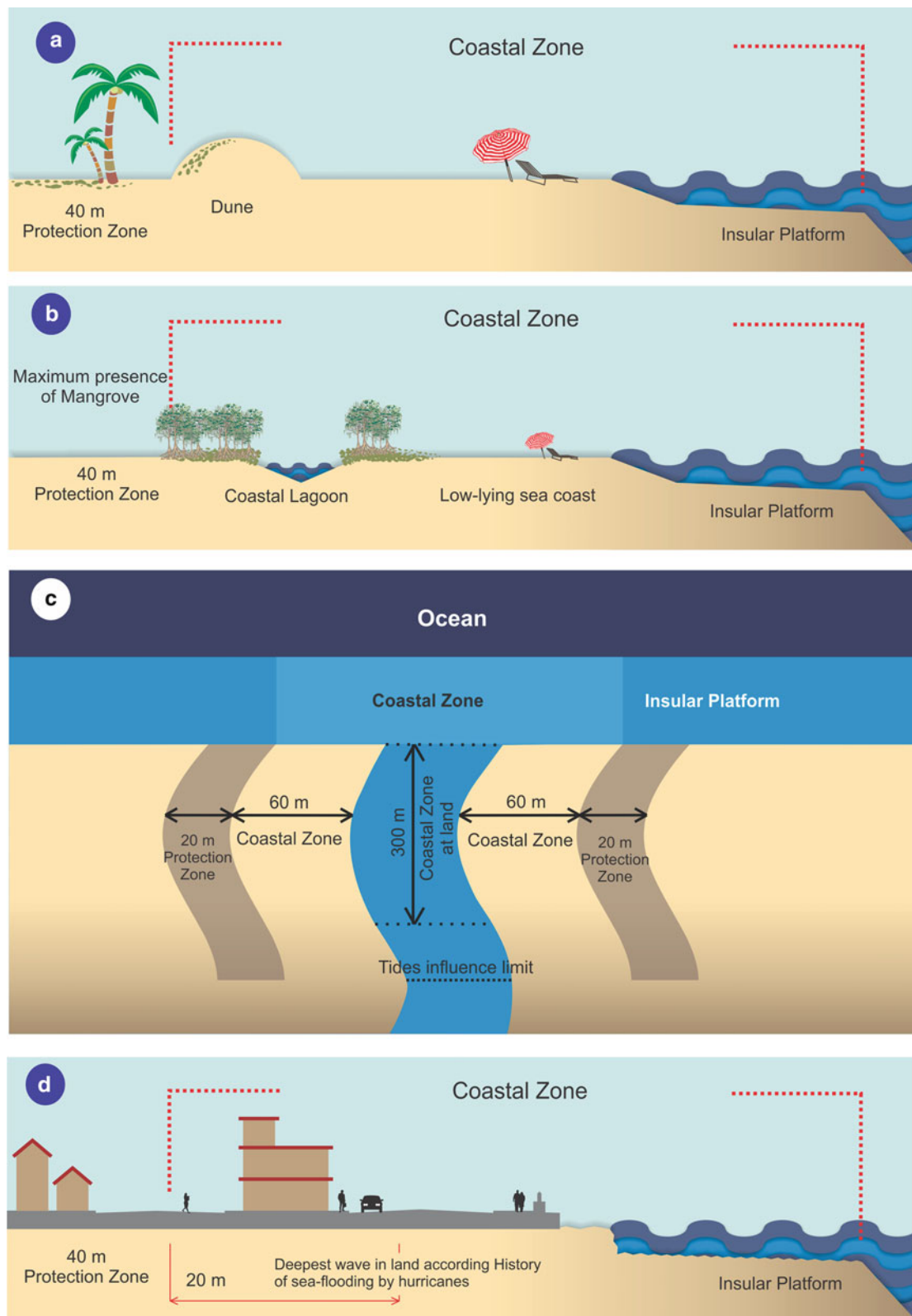
Even though the similarity in the methodological processes of the tools, every insular and coastal region is different, their ICZM and MLUP tools must be put into practice following the corresponding regional, provincial, municipal, or local features. That is the reason of never using a method or procedure as effective model in a specific country, without previously doing the corresponding analysis and adequate procedures.

Conclusions

The review of the legal framework makes evident that no terms, distances, or standard criteria for establishing coastal boundaries have been defined. Understanding of the

Coastal Boundaries, Table 2 Technical terms and corresponding policies for coastal boundaries

Term	Coastal policies, laws, decree laws, or plans	Country	Distances for establishing boundaries at land and sea
Marine–terrestrial public domain	Coastal Law 22/88 (LEC 1988)	Spain	From the line of straight base toward the continent at land and comprising 12 nautical miles of the territorial sea and 200 nautical miles of the exclusive economic zone at sea
Protection area	Jurisdictional regimen of marine public domain (Decreto lei N° 309/93)	Portugal	Protection area of 50 mts., related to the maximum line when high tide in equinoctial waters Examples of tools used: Maps of edge ordering, also called special maps of territorial ordering (Martin 2010) Regional ordering plans, as the one applied in the coastal zone of the Algarve region, where an area of 500 mts. and another one of 500 to 2000 mts. are proposed depending on the municipal plans for territorial ordering (MAOTD 2006)
Orla Costeira	Plan of federal action for coastal zone (<i>Plano de Ação Federal para a Zona Costeira</i>), (MMA 1998)	Brazil	Proposed distances for coastal zone protection vary from 50 to 100 mts
Coastal edge	National politics of coastal edge uses (Decreto Supremo N° 475 1994)	Chile	At land, 80 mts. from the coastal line when high tide and 12 nautical miles at sea
Coastal environmental units	National Environmental Politics for sustainable development of oceanic spaces and coastal and insular zones of Colombia (MMA 2001) National Politics of Ocean and Coastal Spaces (CCO 2007) Decree 1120 of 2013 (MADS 2013)	Colombia	Coastal zones are demarcated according to the following units: National scale: 3 large oceanic and coastal regions Regional scale: 12 coastal–oceanic environmental units Local scale: several integrated management units
Area of public domain	Law of Coastal Zones (GO 2001)	República Bolivariana de Venezuela	Minimum of 80 mts. at land considering the average high-tide line
Protection zone	Decree-Law 212 Management of Coastal Zone (GORC 2000)	Cuba	Terrestrial and marine space adjacent to the coastal zone, minimizing the negative effects of anthropic actions Depending on the coast type, distances vary in ranges of 20, 40, and 80 mts. (GORC 2000) (Fig. 3a–d)
Neap tide	Supreme Court’s 1861, <i>Teschemacher</i> decision	The United States	California’s so-called neap tide rule originated in the Supreme Court’s 1861 <i>Teschemacher</i> decision (Teschemacher v. Thompson 1861). The Court echoed Lord Hale’s seventeenth-century treaty relating “nepe” or “neap tides” to “ordinary tides” for boundary purposes Using the 19-year mean of the high “neap tides” (Graber 2005) Usually departing from a baseline, referring to the low-water line along the coast as marked on officially recognized large-scale charts Following an international law, each coastal nation is entitled to assert a territorial sea immediately seaward of its inland waters as wide as 12 miles from the baseline (Graber 2005) Obligation of allowing free passage to foreign vessels In the past the territorial sea comprised only 3 miles wide, but on December 27, 1988, the boundary changed to 12 miles (Proclamation No. 5928 1988; Graber 2005)
Territorial sea	Convention on the Territorial Sea and the Contiguous Zone, 1958, art. 3 Proclamation No. 5928 1988		
Ordinary high-water mark	English <i>Chambers</i> . Lord Hale’s 1666–67 Attorney General v. Chambers 1854	England	In the past, distances were established according the sovereign’s foreshore or tidelands and the adjacent private littoral lands (Graber 2005) Line of the medium high tides between the springs and the neaps, excluding both the neap and spring tides (Attorney General v. Chambers 1854; Graber 2005) In 1854, English Chambers decided to establish the “ordinary high-water mark” by excluding both the neap and spring tides, using only the intermediate, or medium, tides (Graber 2005)



Coastal Boundaries, Fig. 3 Limits of the “coastal zone” and of the “protection zone” established in the Cuban legal system according to the types of coast. (a) beach; (b) coastal lagoon; (c) river mouth; and (d) anthropic coast

Coastal Boundaries, Table 3 Characteristics of the criteria used for establishing coastal boundaries

Criteria	Characteristics
Arbitrary	Width of the zone is determined by the selected x distance Use of the line of high tide or other similar distances Deficiency in defining the inner limit because of difficult coincidence with the administrative units Limits not related to the most outstanding natural features of the coastal zone and its economic relevance nor to the urban development patterns and the corresponding areas of influence No inclusion of the social and cultural factors nor the corresponding ecological systems of the zone (Henríquez et al. 2011) Very useful when used with the critical biophysical and socioeconomic elements identified at the corresponding zone (Abogado and Méndez 2003) Inclusion of local conditions, such as topography, land use, or habitats (French 2005)
Fixed distances	Usually the measurements of boundaries at land and sea are calculated using the high-water mark Using the ocean component of a coastal area, usually applying the limits of governmental jurisdiction, for example, limits of the territorial seas (Kay and Alder 1999) Permanently fixed by, for example, litigation or boundary line agreements Used on rocky coasts for all practical purposes. Possibilities of changes in tidal data or in future coastal erosion (Graber 2005)
Biological features	Inclusion of flora and fauna elements, for example, the landward limit of a coastal vegetation complex or the seaward limit of a fringing reef (Kay and Alder 1999) Identification of inner limit at land spaces where living organisms depend on marine systems for their vital functions. At sea, limits of many kilometers depend on the most dynamic elements (Abogado and Mendez 2003)
According to uses and particular coastal management issues	Advantage of focusing attention on particular issues (Kay and Jackie Alder 1999). For example, sources of marine pollution would require the definition of an area of attention that includes the river basins. A coastal boundary defined for this purpose would be much larger than one defined to manage other smaller damages at beaches and dunes (Coastal Committee of New South Wales 1994) A great range of coastal uses, marine based or land based, is considered Variation in the distribution of coastal use; therefore, boundaries must vary accordingly. Necessity of determining the main coastal activities, for example, touristical or fishing activities with coastline impact at land and sea (French 2005) Delimitation done according to the kind of problem to be solved Possibilities of solving regional problems and decreasing the working area to solve local problems (Abogado and Mendez 2003)
Hybrids	Use of specific indicators to identify the components of each system Mix one type of coastal definition for the landward limit of the coastal area and another for the seaward limit Fixed jurisdictional limits over nearshore waters; this is a relatively common practice by some governments Inclusion of vertical dimension as part of the overall legislative frameworks, not explicitly covered by specific coastal policies, for example, the ones about minerals below lands and waters, as well as the corresponding atmosphere, generally covered by laws and governmental regulations (Kay and Alder 1999)
Physico-natural features	Inclusion of basic characteristics of the physico-natural scenario, for example, landward limits of Holocene dunes or seaward limits of submarine platforms (Kay and Alder 1999) Use of the spatial limits of the different ecosystems and their structures and functions (Barragan 2003) Necessity of great scientific experience and knowledge about the territory object of study Great complexity of the techniques to be used Importance of the territorial hydrological coherence for demarcation of basins located perpendicularly to the coastline (Planas et al. 2016; Milanés et al. 2017)
Political-administrative issues	The most generalized used for landward limits of local municipalities located frontally to the ocean (Kay and Alder 1999) Advantages for facilitating the decision-making processes and the territorial control mechanisms Advantages regarding the available information about administrative units with possibilities of being compared Limits easily recognizable, cartographed, and well administrative worked Presence of two main characteristics: (1) multiplicity and variety of zone divisions; (2) rigidity, for example, some municipal zones comprise areas not related to coastal systems or presence of ecosystems out of the municipal limits (Henríquez et al. 2011)
Historical memory	Inclusion of past natural threats like hurricanes Integrated recognition of present and future coastal threats and risks (Milanés 2014) Modeling of temporal or permanent flood scenarios in case of combined threats caused by extreme hydrological events and sea-level rise due to the climate change (Milanés et al. 2017) Inclusion of security zones in case of extreme natural events, for example, seaquakes (Barragan 2010)
Practical issues	Limits selected by hierarchical organization of the variables involved in the analyzed problem, for example, if demarcation is done to regional scale, large environmental areas, as sandy or rocky coasts, must be identified and not dunes, beaches, or cliffs (Abogado and Mendez 2003; Henríquez et al. 2011)

Coastal Boundaries, Table 4 Tools, variables, and criteria used in methodologies and International Guidelines of ICZM and L-UP for establishing coastal boundaries

Tools	Criteria and variable used for established coastal boundaries
International methodologies and guidelines	
Methodology for macrozoning the continental coastal zone (Ochoa et al. 2000)	Use of arbitrary criteria, defining 5 kms. starting from the head of the estuaries to the edges of beaches as the most representative areas of the ecosystems and coastal uses At sea, 30 mts. of the bathymetric curve as referential area for fishing shrimps
Zoning the coastal edge. Methodological guideline for communal level: experience of the Bio Bio Region (Salzwedel et al. 2002)	Limits vary according to the characteristics of every commune Analysis done in an area of 4000 mts. length, 3000 mts. seaward, and 1000 mts. landward, starting from the coastal line
Methodological proposal for coastal zone delimitation (Abogado and Méndez 2003)	Use of the decree equivalent to the Venezuelan coastal zone law, where Article 4 refers to the political-administrative and physical-natural criteria established at the plan for ordering and Integrated Coastal Zone Management The following fixed or arbitrary distances are used: Terrestrial area no smaller than 500 mts. measured perpendicularly from the vertical high-tide line up to the coast Area at sea no smaller than 3 nautical miles and not exceeding the territorial sea limits (12 nautical miles)
Project of integrated marine edge management (<i>Projeto de Gestão Integrada da Orla Marítima -Projeto Orla Costeira-</i>) (MMA 2004)	Distance between 50 to 100 mts. for protection of the coastal zone
Guideline of coastal zoning for territorial ordering (Henríquez et al. 2011)	Use of practical criteria for terrestrial delimitation Analysis of the heterogeneity of the physical-natural, socioeconomical, and cultural characteristics of the coastal zone No inclusion of the marine area
International Guidelines for ICZM	
Global status of mangrove ecosystems (IUCN 1983, 1994)	Difficulty in establishing specific demarcations due to the natural processes, the socioeconomical base, and the local development of every nation Recommendation for establishing limits in harmony with the areas of integrated inter-sectorial planning Use of criteria in correspondence to the marine-protected areas and to establish coastal boundaries according to the zone naturality and the biogeographic, ecological, economic, social, and scientific importance of the corresponding areas and also according to the international, or national, significance and practicality/feasibility of the zones
Program 21. Chapter 17, Protection of all kinds of Oceans and Seas. Report of the United Nations about environment and development (PNUMA 1992)	No recommendations for establishment of boundaries
The Noordwijk Guidelines for Integrated Coastal Zone Management (WB 1993)	It is proposed that the management area to be delimited should include all coastal resources of interest and all activities that have the capacity to affect the resources and waters of the coastal area – from water mirrors up to the area of 200 miles
The Noordwijk Guidelines for Integrated Coastal Zone Management (WCCR 1996)	Recommendations about setting limits are not provided The case studies consulted provide a wide variety of boundaries between land and sea offered in; however, the details of the selection of areas are not included there
ICZM Methodologies	
Contribution of science to Integrated Coastal Area Management (GESAMP 1996)	Scale analysis based on boundary problems Programs usually cover particular geographical areas of the different countries In most cases only specific areas of entire ecosystems are demarcated. For example, bays, estuaries, and lagoons, among others, no matter being shared by two or three countries Variations in the size of the management area: large or small No scale analysis is done, nor specific limits are established at land and sea
Environment and development at littoral areas. Introduction to planning and integrated management PGIAL (Barragán 2003)	According to analysis of the physical-marine conditions Analysis of environmental problems according to geographical, functional, and arbitrary or metric criteria
Concepts and methodological guideline for Integrated Coastal Zone Management in Colombia, Manual 1: Preparation, Characterization, and Diagnosis COLMIZ (Alonso et al. 2003)	Environmental units demarcated according to the Colombian coastal politics (MMA, 201) Demarcation of integrated management units adjusted to criteria of local institutions and communities, expressed at different workshops carried out in the study area
Statement of zones under Integrated Coastal Management in Cuba (Salabarría and Brito 2011)	No specific criteria for demarcation, although the problems to be solved may be taken into account No use of boundaries established in the Cuban Decree 212

(continued)

Coastal Boundaries, Table 4 (continued)

Tools	Criteria and variable used for established coastal boundaries
Local indicators of sustainability for environmental energetic management at the coastal zone of the southeastern region of Cuba (Planas 2012)	Two types of coastal boundaries are used: Following the Cuban Decree 212 and stating that the extension will depend on the interest of the management itself and management actions Considering the criteria of proposed landscape units and local environmental sustainability indicators
Method for demarcation and delimitation of coastal zone boundaries DOMIZC (Milanés 2014; Milanés et al. 2017)	It is an integrated methodological proposal for demarcation and delimitation of coastal boundaries according to ICZM and L-UP perspective Demarcation of Primary Environmental Coastal Unit for Integrated Management (PECUIM) and Basic Environmental Coastal Unit for Management and Land Planning (BECUIMLUP) are carried out at land and sea with integrated criteria Delimitation at sea to consider: (a) diving or preservation zones; (b) presence of coralline barriers or coral patches; (c) presence of protected marine areas; (d) presence of cultural–subaquatic patrimony; (e) presence of fishing zones; (f) presence of zones of mineral deposit exploitation Delimitation at land use the variables: (a) history of sea-flooding; (b) deepest sea-flooding in land according to threat, vulnerability, and risk studies; (e) sea-level rise caused by climate change; (d) combined threats; (f) presence of physical infrastructures; (g) natural consolidated vegetation; and (f) population identity and sense of property
Cuban methodologies and guides for land-use planning and spatial planning	
Methodology guide for special plan for territorial ordering in touristic areas (Pantaleón 2001)	Along the limits of the tourist area, those of its area of influence are defined Land and sea tourist attractions and their relationship with other functional elements such as transportation facilities, sources of supply, and nearby human settlements in which a significant part of the population lives or will inhabit are linked
Guide for land-use planning (Padrón 2002)	It is applied in coastal municipalities The Cuban Decree 212 is considered
Methodology for threats, vulnerabilities, and risk in the territorial scale (AMA 2014)	For the deepest sea-flooding in land, methodologies that consider threats, vulnerabilities, and risk studies are used
Methodology for protected areas management (Gerhartz et al. 2008)	The terrestrial part through terrain accidents or recognizable anthropogenic elements is delimited On the marine side, straight lines determined by prominent points on the coast Marine-protected areas are zoned according to protection interests which are followed
Methodological guide for environmental planning in Cuba (Obllurys et al. 2009)	Environmental units according to homogeneous patterns are defined: these are established by the natural and socioeconomic functions of the territory, by crossing them into a geographic information system No delimitation is required in the marine environment

terminology used to establish limits and the aims, or purpose for carrying out that task, presents several outputs.

Over the last century, significant progress has been achieved in the implementation of policies, laws, and decree laws internationally developed for establishing coastal boundaries. In this way, the impact caused by the Cuban Decree-Law 212 has been well received and considered as a reference in some Latino-American countries.

Coastal boundaries determination, depends upon the purpose for which this is made, for example Depending on each case, correct establishment of boundaries, either at sea or at land, five aspects must be considered: (1) strategic activities for coastal development, making sure the different activities are compatible; (2) areas of potential development and sustainable productive activities; (3) areas of ecological and patrimonial values; (4) areas of preferential coastal use; and (5) involvement of social actors located on the coastal edge.

A multidimensional and multidisciplinary view is necessary for absolute understanding of coastal boundaries. When analyzing coastal boundaries, several integrated criteria must be

taken into account: ecology; ecosystem biodiversity; existing problems and conflicts on the protected areas; the inhabitants' sense of property, history, and culture of the coastal communities; as well as the political–economical and technical–productive infrastructures located at coastal zones. The previously mentioned integrated criteria show the holistic character of the processes for establishing coastal boundaries.

The great diversity of criteria used to establish coastal boundaries makes the use of a universal method very difficult. Some methodologies are not entirely acceptable because of their complexity. Some of them have involved relatively primitive and inaccurate techniques, as well as variables and procedures so sophisticated that updated cartographic maps and geographic information systems have been frequently necessary.

Cross-References

- [Coastal Management Practices](#)
- [Coasts, Coastlines, Shores, and Shorelines](#)

- **Geomorphology and Sea Level**
- **History, Coastal Protection**

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Coastal Changes, Gradual

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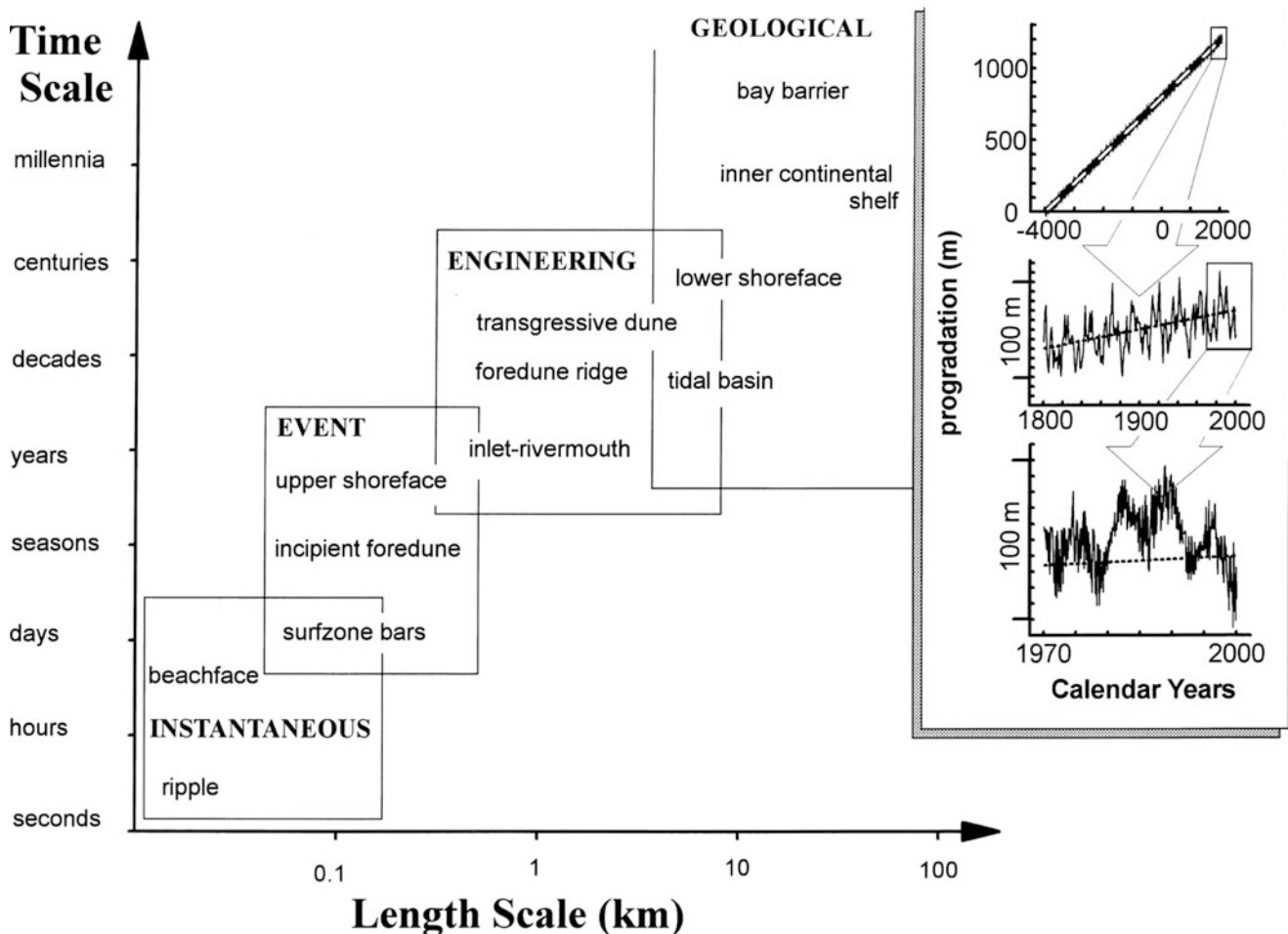
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The concept of gradual coastal change can be seen as embracing different contexts, both spatial and temporal. It is conceivable that gradual change may include those depositional and erosional events that have taken millions of years to produce present-day coastal landforms at one end of the spectrum, while at the other end slow and imperceptible accretion to beaches after storms may represent another form of gradual change. Therefore, a wide-ranging time frame may be used covering geological, engineering, and arguably event scales as depicted in Fig. 1.

This figure demonstrates how high frequency, low magnitude processes at shorter time scales can be superimposed on longer-term oscillations and trends at scales of millennia and longer term as reflected in depositional and erosional land forms.

Cenozoic Geologic Scale

Coastal geomorphologists have a long-standing interest in the evolution of coastal landforms from Tertiary to Quaternary



Coastal Changes, Gradual, Fig. 1 Spatial and temporal scales of coastal geomorphic change depicting examples of features formed at different scales and how oscillations in magnitude and frequency of

processes may be superimposed on longer-term trends (e.g., progradation of a strandplain with beach cut and fill cycles superimposed as at Moruya, New South Wales, Australia, see Thom and Bowman 1981)

times. Many coastal regions show an historical imprint or inheritance of depositional or erosional processes reflecting changes in shoreline position as sediments accumulate or cliffs erode. In glaciated areas, the impact of gradual coastal accretion or erosion from Tertiary times is rarely apparent. Similarly, where tectonic and volcanic activities prevail there is little evidence of shoreline features of pre-Quaternary age. However, tectonic uplift as seen, for instance, along the Huon Peninsula in Papua New Guinea, can provide a mechanism for preserving the remnants of gradual coral reef accumulation over hundreds of thousands of years which correlate very well with glacial to interglacial climatic and eustatic cycles (Chappell and Shackleton 1986).

Classic thinking in coastal geomorphology, epitomized by the work of Douglas Johnson (1919), dramatically demonstrated the nature of gradual change following the application to coasts of the Davisian cyclic model in geomorphology. Assumptions built into this model may be queried, but the basic notion of change driven by the interaction of marine

forces (especially waves) and the land still underpins any conceptual framework for understanding how coasts evolve over time. The importance of sea-level change during the Quaternary is now seen as a major factor in coastal geomorphology, a phenomenon of relatively limited significance in the Davis/Johnson model. Shepard, Russell, Fairbridge, Armstrong Price, Steers, and Guilcher (and others) are of a generation that assessed coastal landforms in different regions of the world from a Quaternary perspective. They showed how erosional, and especially depositional landforms, over time, have changed their character under different sea-level conditions. Many coastal regions experienced post-depositional modification and transformation after primary wave, tidal and wind processes have performed their work (Thom 1975).

The Australian coast contains many examples of gradual change in coastal environments during the Cenozoic. These features have formed since dislocating tectonic forces that ripped the continent apart from Gondwana remnants

essentially stopped operating about 60 million years ago. On the east coast, the narrow continental shelf has been subjected to very slow marginal subsidence. This has facilitated two geomorphic outcomes: shelf-edge accretion of a sediment wedge ca. 500 m in thickness, and an inner-shelf abrasion surface up to 10 km wide and ca. 1,000 km long (the East Australian Marine Abrasion Surface – EAMAS). The surface is now veneered with late Quaternary sediments and is entrenched by infilled paleo-valleys carved by rivers draining to lower sea levels. The surface culminates on the landward side with discontinuous marine cliffs cut into Mesozoic or Paleozoic rocks often over 100 m in height and representing truncated spurs and ridges of drowned valleys and plateaux (as seen around Sydney). We are uncertain of the precise age of initiation of the surface and cliffs, but it is easiest to conceive of these features as time-transgressive evolving progressively over millions of years during periods of relatively higher sea level.

Another large-scale Australian example that reflects gradual accumulation of sediments are the barrier-lagoonal deposits formed over millions of years within the lower Murray-Darling basin. In this region, relative sea level since the early Tertiary has fallen. Sands, including heavy minerals such as ilmenite, form horizontally stacked barriers which in Quaternary times have been partially reworked into terrestrial dune fields. Similar “stepped” or extensive strand plains initiated in the Tertiary and extending through the Pleistocene into the Holocene have formed in southwest Western Australia near Perth, in parts of southeast Brazil, and along the Atlantic coastal plain of the United States.

Holocene Geologic Scale

At the other end of the geologic scale are coastal phenomena which are the result of processes operating at or close to present sea level over the last 6,000 years or so. Within this time frame, erosion and/or deposition can initiate and reconstitute sand barriers, tidal flats, river mouths, tidal deltas, coastal dunes, and rock platforms at the macro scale, and a myriad of landforms at a micro scale. Use of techniques of drilling and dating have enabled age structures and sedimentary histories to be reconstructed from Holocene deposits in coastal areas of many parts of the world.

In areas of postglacial rebound beach deposits dated 6,000 years BP may occur 100–150 m above present sea level (e.g., Gulf of St Lawrence), whereas for areas marginal to the great deltas, like the Mississippi, deposits of similar age may have gradually subsided below sea level and be buried by younger sediments (see Carter and Woodroffe 1994). In contrast, those coasts experiencing a much more passive neotectonic regime, such as eastern Australia, witness beach/

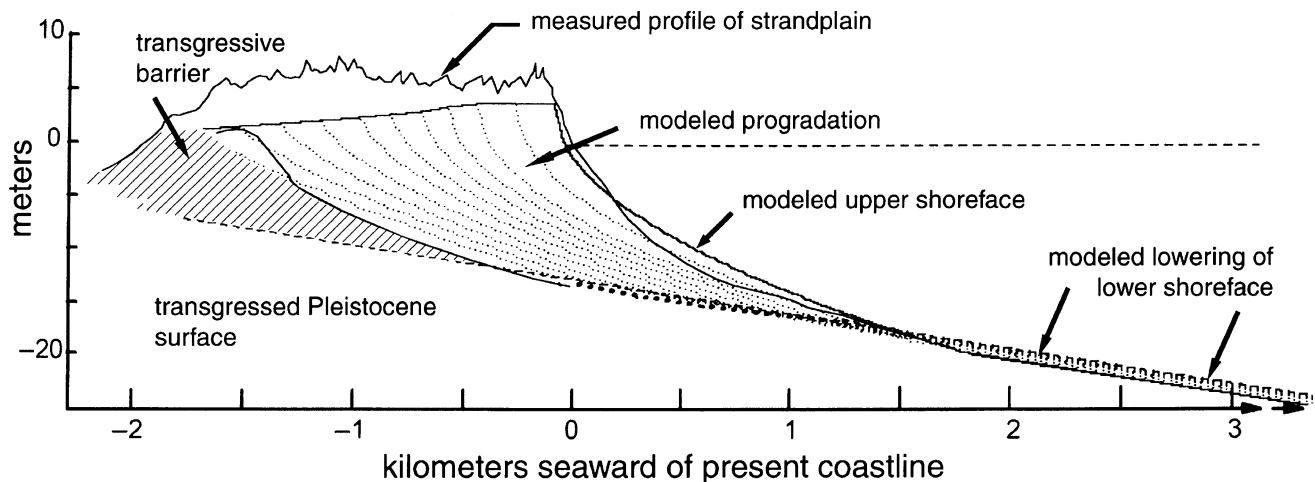
foredune ridge accumulation at about present sea level not only in the Holocene, but also in the last and penultimate interglacials (Roy et al. 1994; see Hopley 1994, for a review of gradual coral reef accumulations in similar tectonic settings).

One conclusion which can be drawn from an analysis of Holocene sand barriers is how gradual processes of accretion are periodically interrupted by erosion or reduced rates of accretion. The result is a strand plain with distinct linear ridges (beach or foredune) parallel or sub-parallel to the present shoreline. Changes in sediment budgets are critical in the gradual evolution of strand plains. As pointed out by Cowell et al. (1999) sand may be supplied from the shoreface as well as littoral sources in order for accretionary processes to work. Closed sediment compartments highlight the effectiveness of gradual onshore transport of sediment from the lower shoreface in certain areas (see Fig. 2; Roy et al. 1994). These areas are in contrast to those “leaky” compartments dominated by littoral sand transport associated with river mouth (inlet) or headland bypassing.

Engineering Scale

Even closer to the present are those features which result from erosion or deposition over periods of 2 years or more. Increasingly, coastal geomorphologists are recognizing the impact of oscillations in climatic conditions leading to changes in beach volume, beach alignment, inlet openings/closures, river-mouth bar migration, dune vegetation recovery, etc. However, some of these changes may reflect “self-organization” of morphodynamic conditions involving positive and negative feedback effects independent of climatic forces (see Cowell and Thom 1994, citing field studies in the United States, Netherlands, and Australia)

Recovery of the subaerial beachface in width and thickness following a period of sudden storm wave erosion is well documented in coastal literature (Komar 1998). Storm impacts in winter followed by gradual summer accretion are characteristic of many coasts in North America (e.g., California) and Europe. The severity of impact may be enhanced during El Niño years as in Oregon, and that gradual beach recovery may take longer after such severe winter events (see Komar 1998 for description of El Niño storm impacts). Contrasting situations occur in the southwestern Pacific where beaches may accrete through many years (winter and summer) influenced by El Niño only to erode drastically in La Niña or storm-dominated periods (Thom and Hall 1991). The role of the Pacific Decadal Oscillation (PDO) as a factor in driving beach cycles of erosion and accretion is now seen as a significant phenomenon which may explain the



Coastal Changes, Gradual, Fig. 2 Simulated strandplain development during the Holocene at Tuncurry, New South Wales, Australia (after Cowell et al. 1999). The simulation involved a constant sea level for 6,000 years and an across-shore sand supply from lower to upper

shoreface averaging 1.1 m^3 per meter of shoreline together with a steady increase in depth of the lower shoreface at a rate of 0.002 m/year decreasing linearly with distance to 10 km seawards of the present coastline forming 80% of sand now in the strandplain.

timing of events that do not easily fit ENSO patterns. El Niño and La Niña can be seen as phenomena lying on top of the large-scale temperature and sea-level height distributions determined by the PDO.

Understanding processes of gradual beach accretion is of significance in beach management. Monitoring beach change by photogrammetric and/or field survey methods yields valuable information on how quickly all or part of a beach may recover following a stormy period. Beaches and foredunes in New South Wales took 5–10 years to reach pre-storm positions following the last major episode of erosion in 1974 (Thom and Hall 1991). During phases of “gradual and imperceptible” accretion, landowners with land title tied to an “ambulatory” boundary (such as mean high water mark) may seek a redetermination of their land title to claim accreted land. Under the English “doctrine of accretion” exported to many other countries, the boundary remains fixed if sudden erosion cuts back into this new land. A consequence is for landowners to seek to defend their properties with seawalls and bulkheads against erosion. Beach amenity can suffer as a result of these actions (Titus 1998).

Gradual change in coastline position and form, in tidal flat morphology, in the nature of inlet openings, and in the vegetative cover of coastal dunes are phenomena that affect the lives of millions of people. Understanding the dynamics of gradual changes in coastal geomorphology requires monitoring and historical reconstruction using a variety of techniques and analytical procedures. The use of simulated modeling, as seen for instance with the Shoreface Translation Model developed by Cowell et al. (1995), highlight a way of achieving such an understanding when coupled with a knowledge of geomorphic history and processes (see Fig. 2).

Cross-References

- [Changing Sea Levels](#)
- [Coastal Changes, Rapid](#)
- [Coral Reefs, Emerged](#)
- [El Niño–Southern Oscillation \(ENSO\)](#)
- [Holocene Coastal Geomorphology](#)
- [Sea-Level Fluctuations Over the Last Millennium](#)
- [Submerging Coasts](#)
- [Uplift Coasts](#)

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Coastal Changes, Rapid

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Introduction

There are many definitions of what might be called rapid events occurring on coastlines from sea-level movements caused by global warming and/or cooling (10s to 1000s of years) to tsunamis that occur in seconds. In this entry the types of rapid changes discussed will be of the more instantaneous nature from *tsunamis* (*q.v.*) to *storm surges* (*q.v.*), the type of damage done and what the events leave as a trace in the geologic record. These phenomena were deemed sufficiently important by UNESCO/IGCP (International Geological Correlation Program) to form an IGCP project (IGCP 367) that lasted for 5 years (1994–98), where over 500 people from 70 countries were involved working on almost every aspect of rapid change in the coastal zone. Much of this entry will be drawn from the five volumes produced from that project (Scott and Ortlieb 1996; Shennan and Gehrels 1996; Murray-Wallace and Scott 1999; Scott 1999; Stathis et al. 2000).

Climatic Events

Storms

By far, the most dramatic of climatic events are hurricanes or typhoons as they are called in the Pacific. These events produce profound changes on coastlines and particularly man-made structures in the coastal zone. The effects are of two types, wind damage and water damage from elevated tides and torrential rains causing coastal flooding. At the center of a hurricane winds can reach over 200 km/h and storm tides can be as much as 8 m above normal. The amplitude of the storm tide depends on the barometric pressure (the lower the pressure, the higher the tide), the wind speed which

causes the water to pile up onshore and the natural state of the tide (i.e., high versus low tide stage). The worst scenario for high storm tides is low pressure and high winds coinciding with a normal high spring tide.

The type of damage done by wind consists of knocking down of forests as occurred in hurricane *Hugo* in South Carolina (USA) in September of 1989 when the eyewall swept through a national forest and simultaneously knocked down most of the forest in a few seconds. Together with that it also destroyed many coastal man-made structures. Probably the most damage and deaths were caused by the storm surge, which at the eyewall location was 6 m above normal. However, even 50 km from the direct hit, the storm surge was almost 3 m. A purely human problem was sorting out which phenomenon caused the damage, the wind or the water, because many people did not have government flood insurance, which meant that their losses were not covered by their commercial hurricane insurance.

Maybe not too surprisingly the traces left in the naturally occurring geologic record are much less profound. At the eyewall hit of *Hugo*, there was a 10 cm sand layer but at a location 50 km north there was no sedimentological trace and in the extensive marsh areas there was not even a thin sand layer (Collins et al. 1999), the natural systems can act as a buffer to storm events, perhaps a lesson to the human population. The hurricanes do leave a trace in nontidal *coastal inlets* (*q.v.*) of either small sand layers or reworked microfossils, which can be used to build a chronology of storms on some coastlines (Liu and Fearn 1993; Collins et al. 1999). Also strong surges and sustained winds (either from hurricanes or long lasting winter storms such as “nor’easters” on the Atlantic coast of North America) can create washovers and change inlet positions in *barrier island* (*q.v.*) systems.

Global Warming

This effect causes a general sea-level rise on scales from 10s of years to 1000s of years both by melting of glacial ice caps and steric expansion of the ocean, which may amount to 30–40 cm. Depending on the coastal gradient this may or may not be significant.

It is difficult to measure these changes (Pirazzoli 1991) because they are usually on the order of a few centimeters. Mostly they have been detected by *tidal gauges* (*q.v.*) for the last 50 years (Pirazzoli 1991) but these measurements are sometimes questionable. Some *biological markers* are now capable of measuring very small changes (± 5 cm) in sediment records (Scott and Medioli 1986) such that it is possible to measure rapid sea-level changes in the late Holocene (e.g., Scott and Ortlieb 1996). It is these mid-Holocene measurements that may help us to understand the suggested *sea-level rise* (*q.v.*) to be caused by anthropogenic *global warming* (*q.v.*).

Seismic Events

Earthquakes

Sudden land movements on coastlines can either instantly emerge or submerge coastlines and any structure that might be in the way. This is particularly a problem around earthquake prone coasts such as the Pacific Rim and most of the northeastern Mediterranean.

The type of records preserved are often in coastal marshes because they remain depositional and their strong vertical zonations help to show sharp vertical surface changes, that is, sudden change from forest to marsh if the land drops or vice versa if the land emerges (Atwater 1987; Atwater et al. 1995; Shennan et al. 1999). There are outstanding records of these phenomena all up and down the west coasts of both North and South America. In rocky coastline areas, *biological markers* maybe uplifted far above their natural range, providing a good indication of rapid change and the magnitude of that change (Ortlieb et al. 1996).

Extended records of these events have been documented in the northwestern part of the United States and southern British Columbia, Canada that show a 300–500 year periodicity for magnitude 8 or 9 (on the Richter Scale) earthquakes (Atwater 1987) over the last several thousand years. Until recently, it was difficult to plan based on a 300-year return time but Shennan et al. (1998, 1999) found in two separate places that there appeared to be a precursor event between one and three years prior to these events that could be detected in coastal forest sequences that submerged during a major seismic disruption. The most compelling evidence is from the traces left by the Great Alaska Earthquake of 1964 that was the second largest seismic event ever recorded at 9.2 on the Richter Scale. The evidence from this quake is convincing because unlike the prehistoric quakes in Oregon (Shennan et al. 1998), in Alaska, we know exactly when the event occurred and we know absolutely that the forest layer was above tidal influence before the March 26, 1964 quake. What is seen is a small change in both diatom and foraminiferal assemblages in the forest *before* the quake that indicate a slight subsidence, maybe 30 cm, that was not detected by anyone living there or by macro-vegetation (Scott et al. 1998; Shennan et al. 1999). So it is possible that there may be a technique now that we can use as an early warning system for these mega-quakes.

Tsunamis

Tsunamis (q.v.) are perhaps the most spectacular and frightening of the coastal effects caused by earthquakes. They are large wavelength, fast moving waves resulting from the rapid land slumps or uplifts caused by earthquakes in offshore areas. When they are in the deep water of the open ocean they are hardly noticeable but when they reach the shallow coastal regions the energy is converted to wave

height, which in many cases can reach as high as 20 m. The most recently documented cases at this writing were the 1994 Java (Dawson et al. 1996) and the 1998 Papua New Guinea tsunamis where several thousand people were killed (McSaveney et al. 2000).

Tsunamis also leave highly detectable traces in coastal deposits in the form of sand layers contained in marsh or nontidal *coastal pond* sediments. However, these traces cannot be reliably separated from hurricane layers, so if both phenomena occur in the same area it could be confusing to sort out which is which. As with storm traces, the best evidence for offshore transport is the presence of marine microfossils. Many times the tsunami will leave no trace if it travels over hard substrates.

The occurrence of tsunamis is limited largely to the Pacific Rim where they can occur either very close or very far from the actual seismic event; an example of a distant catastrophic tsunami was the killing of several people and destruction caused by a tsunami in Crescent City, California (USA) resulting from the 1964 Alaska earthquake whose epicenter was several thousand kilometers to the north.

Sedimentation/Erosion Events

Natural Changes

There are rapid events in coastal areas caused simply by rapid sediment events that may be a result of flood deposits, rapid deforestation caused by forest fires, or any type of event that changes the sediment budget in a dramatic sense to cause a shift in morphology of the coastline. In the case of a forest fire, large amounts of erosion as a result of deforestation would place sediment into *coastal inlets (q.v.)* causing rapid infilling or conversely, if a sediment source is reduced it can cause the shoreline to retreat rapidly in the absence of any other effects (e.g., sea level or storms).

Anthropogenic Causes

By far the most spectacular rapid sedimentation events are caused by human intervention. They are almost too numerous to mention but almost every marine *delta (q.v.)* in the world is being influenced by sediment starvation which causes rapid land loss as the delta continues to subside with no replenishment of sediment (e.g., the Nile and Mississippi Deltas). The starvation results from dams either that trap the sediment upstream (the Nile) or diversions of the rivers, which do not allow the river to switch back and forth (the Mississippi).

Other less spectacular but damaging rapid sedimentation events are caused by careless placement of structures on the coast that impede *sediment transport (q.v.)* rapidly changing the balance of erosion/ sedimentation.

Summary

Rapid coastal changes can take many forms; only a few of the most obvious examples are discussed here. Often the geologic record of these events is difficult to detect, especially in dynamic coastal environments, where tidal activity reworks most of the record. The records are best in nontidal *coastal ponds and inlets* where an extreme event brings in allochthonous sediment that is then sandwiched in quiet water, often organic deposits and are relatively simple to detect and date.

Cross-References

- [Barrier Island Landforms](#)
- [Changing Sea Levels](#)
- [Coastal Changes, Gradual](#)
- [Cross-Shore Sediment Transport](#)
- [Deltas](#)
- [Longshore Sediment Transport](#)
- [Meteorological Effects on Coasts](#)
- [Longshore Sediment Transport](#)
- [Storm Surge](#)
- [Tide Gauges](#)
- [Tsunami Deposits](#)

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Coastal Climate

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Three factors are important in an explanation of coastal climate. The first is the juxtaposition of two surfaces, land (or in some cases, ice) and water, of very different properties. The second is the geographic location of the coast with respect to continental masses and global-scale atmospheric and oceanic currents of air and water, respectively. The third is the topography of the coast and its hinterland. We must also take into account temporal and spatial scale issues so as to order the different aspects and to maximize our understanding of coastal climates. Scale will be used as an organizing framework in the following account.

Several useful data and information sources exist for an entrance into the topic of Coastal Climatology. One is an encyclopedia entry under the same topic name by Walker et al. (1987) in *The Encyclopedia of Climatology*. A second is the Master Environmental Library that is an online data source for environmental data for coastal modeling applications (Allard and Siquig 1998).

Time and Space Scales

Climatologists recognize different types of climate according to their time and space scales. The local wind currents that redistribute sand around a beach sand dune and the different

temperatures on the sunny and shaded side of a stand of beach grass would be in the realm of *microclimatology*. Different climates formed by the different surfaces and local climatic interactions as one might traverse the sand of a barrier island, the water of its related lagoon, the inland coastal marshes, and the forest or agricultural land of the immediate hinterland would constitute a study of *topoclimatology*. A sea breeze might penetrate inland some 75 km from the shore on a given day and would represent a phenomenon studied in *mesoclimatology*. Climatic variable value gradients parallel to the coastline are usually much steeper than those that are perpendicular to the coastline. The geography of the US Pacific Northwest (PNW), for example, gives rise to well-marked gradients in values of climatic variables both in the latitudinal (north-south) direction but also in the longitudinal (west-east) direction. This is not uncommon for coastlines that have some north-south alignment. The location of a coast in relation to global latitude, continental geography, hemispheric-scale winds and storm tracks in the atmosphere, and surface water currents in the ocean, and the state of teleconnective indices such as the Southern Oscillation and North Atlantic Oscillation, would help explain its *macroscale climatology*. The passage from micro to macroscale defines a scale hierarchy. Most scale hierarchies in Earth Sciences, including coastal climates, have at least three characteristics. First, events on the larger scales tend to set the context for those on the scale below. Second, explanation of events on one scale is best given by reference to events on the scale immediately above and below the scale and phenomenon in question. Third, at least at the lower end of the hierarchy, there tends to be a direct relationship between time and space scales. For example, ephemeral events, like an eddy in the wind, happen on small time and space scales while some other events, like a hurricane, sometimes take many days to run their course and have the potential to affect a relatively large geographic area.

Microscale Coastal Climates

Wind is the principal climatic variable at the microscale in coastal climates and its interaction with topographic and biologic factors has long been studied. Three factors are important at this scale. First, wind at coastlines tends to come from a predominant direction often from offshore. This has the effects of distributing sand and sand dunes in particular ways, and of sometimes shaping the vegetation canopies. The actual type of sand dune formed has been suggested to be a function of the supply of sand, the density of vegetative cover, and the average wind speed (Strahler and Strahler 1992, p. 406). Second, the capacity of wind to carry sand is an exponential function of its velocity. Thus when wind speeds are decreased, by, for example, the presence of vegetation or a fence on a beach, then sand

and other particles being transported are rapidly deposited. Vegetation and wind thus have a synergistic interaction in stabilizing sand movement. Third, typically wind velocity increases logarithmically with height. This fact needs to be taken into account when designing high-rise buildings in coastal environments. Pilkey et al. (1998, p. 254) have noted that at any given height pressure from the wind is greater near the shore than it is inland.

Topoclimatology

The coast, interpreted broadly, may be formed of many different surfaces including water, sand, soil, rock, and a variety of vegetated and urban surfaces. Each one of these surfaces has a different energy and moisture balance, and depending on its roughness, a different degree of momentum transfer with the atmosphere. These differences might not be too important on the topoclimate scale on a coastline with a steep cliff topped with relatively homogeneous vegetation because the climatic differences would be overwhelmed by both smaller- and largerscale processes. However, the differences are important on a coastline typified by a barrier island, lagoon, and hinterland series on days when the large-scale wind is not too strong. The differences are important because they will form a set of distinctly different microclimates that will relate, at least, to the flora and fauna of the area.

Salt transport by wind is also important at this scale. The concentration of salt particles near the surface in coastal climates is directly related to the wind speed (Bigg 1996). A high number of salt particles makes the coastal air very hazy. The salt particles attract water vapor and play an important role in raindrop formation. Damaging salt spray can sometimes have a pruning effect on coastal vegetation. How far inland or how high up a cliff this effect can occur is determined by wind flow and its interaction with topography on the scale in question. Soils near coasts are typically high in salts transported by wind from the ocean.

Mesoclimatology: The Land–Sea Breeze Circulation

It is the different properties of the water and land surface that ultimately lead to the land-sea breeze phenomenon – one of the most studied features of all aspects of coastal climates. The circulation is developed at mid and low latitude coasts on days when larger-scale wind systems are not predominant. Anticyclonic weather in summer time provides an ideal context. The circulation is found both on maritime and freshwater coasts. At the surface of the Earth, the land-sea breeze circulation takes the form of winds coming from water surface during the day and from the land during the night. The surface

of a water body usually manifests only a small daily range of temperature. Oke (1987) has outlined the four reasons for this. Solar radiation can easily penetrate a water surface and thus heats a deep layer and not just the surface itself. Convection currents and mass transport in the water permits any heat gain or loss to be spread throughout a large volume of water. Evaporation from the water surface cools the water surface and helps promote vertical mixing within the water itself. Water has a high thermal capacity compared with soil and thus takes about three times the amount of heat to raise its temperature by 1 °C as for the same volume of soil. Thus there is a reduced flow of sensible heat from the water surface to the air compared with that over the land during the day. This leads to a reduced warming rate of the air over the water compared with that over the land surface. As described by Oke (1987), greater sensible heat flow over the land in the morning heats the air columns more rapidly and to greater heights than over the water. The atmospheric pressure well above the surface of the land is higher than at the same level over the water. The resulting horizontal pressure gradient results in airflow at upper levels toward and out over the water surface. This flow ultimately produces a higher atmospheric pressure at the surface over the water that, in turn, forms a cross-coastline flow of air from the water to the land. This surface level flow is known as the *sea* or *lake breeze*. The *land breeze* starts in the evening because of the greater contraction and cooling of the air columns over the land compared with the sea.

Oke reports that the daytime sea breeze circulation has greater vertical and horizontal extents and larger wind speeds than the land breeze because it is driven by solar forcing. Typically, the sea breeze might have a velocity of 2–5 m/s, a depth of 1–2 km, and extend inland as far as 30 km. The land breeze characteristically has a velocity of 1–2 m/s and a smaller extent than the sea breeze. If the sea breeze is forming in a location where there is cold upwelling water, then the air formed over this water may be cool enough to continue the sea breeze circulation all through the night and not have a land breeze develop at all.

The sea breeze brings with it air that is cooler and more humid than the air found over the land. This cool air plows underneath the warmer air over the land and forms a localized cold front known as a *sea breeze front*. The sea breeze front gradually advances inland during the day and is associated with the development of sea breeze cumulus clouds. These clouds are often taken by the counter (seawards) flow above the surface and dissipate because they are decoupled from their moisture source. Purdom (1976) showed that some parts of the front displayed higher concentrations of convection than in others. He found that the shape of the coastline in some places created stronger upward vertical velocities. Kingsmill (1995) documented for summer months the collision of a sea breeze front traveling from the east side of the Florida peninsula with a gust front traveling from the west.

Interestingly, he found that convective activity was most prominent just prior to the collision of the two fronts rather than at the time when the two collided. Oke (1987) has suggested that if the sea breeze travels far enough inland then the influence of the Coriolis force will affect its direction and the breeze will end up traveling parallel to the coast. A fairly deep penetration of the sea breeze front inland was observed in April 1999 at Wokingham, which is 75 km inland from the southern coast of England. The passage of the front was accompanied by an increase of the dew point temperature of 4 °C and a change of wind direction from 70° to 210° (Burton 2000).

The climatology of sea (lake) and land breezes is highly variable depending on the particular location being considered. A 15-year study of Lake Michigan found that lake breezes tend to occur more frequently along the eastern shore of Lake Michigan than along the western shore (Laird and Kristovich 1998). In addition, there was a gradual increase in the number of lake breeze events from May through August, then a slight decrease into September. The distribution of the frequency of land breezes is similar to that of the lake breezes from May through September, however, they occur about 10% less often than the daytime lake breeze circulation. Typically, sea breeze activity is observed on the south coast of England on 75 days during the year (Burton 2000).

The land-sea breeze circulation has many practical implications. Oke (1987) quotes one study in which tobacco leaves near the coastline of Lake Erie were damaged by ozone that had been formed over the lake from lakeshore city pollutants and then transported back over the shore and tobacco crop by the lake breeze. In another study, quoted by the same source, blister rust disease impacted pine trees 15–20 km from the Lake Superior lakeshore. This impact was thought to be due to spores from bushes located between the diseased trees and the lake and transported out over the lake at night, back aloft by the counter airflow of the land breeze, and then descending upon the pine trees. Michael et al. (1998) showed that during spring and summer, sea breeze circulations can strongly influence airport operations, air-quality, energy utilization, marine activities, and infrastructure in the New York city region. Here, the investigators found the geographic configuration of region presents a special challenge to atmospheric prediction and analysis. The New Jersey and Long Island coasts are approximately at right angles to each other; additionally Long Island Sound separates Long Island from the mainland of Connecticut. The various bodies of water in the region (Atlantic Ocean, Long Island Sound, New York Harbor, Jamaica Bay, etc.) have different surface temperatures. In addition the urbanization of the New York area can modify atmospheric flows.

Apart from land-sea breeze phenomena mesoscale wind fields in general are very important in influencing the coastal and coastal ocean processes. Considerable attention is being

given, for example, to the effect of wind on driving and directing freshwater plumes from rivers entering the ocean. This has many implications for cross continental shelf transport (Lohrenz 1998).

Macroscale Climatology

The location of the coast in the world determines in large part its overall climate. A first order classification of coasts on this scale would include tropical windward coasts, tropical leeward coasts, mid- and high-latitude windward coasts, mid- and high-latitude leeward coasts, and ice-influenced coasts. Each of these categories would be modified according to whether or not the hinterland of the coast contained mountains. Examples of climate data from such categories are given in Table 1. Walker et al. (1987) quote other ways of classifying coastal climates.

Under the classification suggested here the extremes for high precipitation are seen in the tropical and mid-latitude

windward coasts with mountains such as, respectively, on the island of Maui and other islands in Hawaii, and the PNW of the United States. In the case of the north coast of Maui, Hawaii, onshore moist tropical air is lifted up and cooled against the mountains giving rise to some of the world's highest precipitation values. Similar orographic effects occur at the coast of the PNW but here it is moist, mid-latitude Pacific air that is uplifted often during mid-latitude storms, which provides intense precipitation. The extreme for low precipitation is found on tropical coasts on the western sides of continents, as for example, on the coast of Peru. Air flows down the mountain sides of the Andes mountains in Peru and northern Chile and thus is warm and dry when it reaches the coast. More importantly this coastline, and others in this category, has a cool ocean current offshore. Upwelling of cold water gives rise to fog that is often the only source of precipitation in this climate. Ice-ocean boundaries are not considered to be coastlines by some (Walker et al. 1987) but since they form a distinctive interface on the Earth's surface they will be briefly mentioned here. Thus the "coasts" with

Coastal Climate, Table 1 Climate data for selected coastal areas

Tropical	Location	Latitude (Deg Min)	Longitude (Deg Min)	Elevation (m asl ^a)	Mean annual temperature (°C)	Mean monthly temperature Range (°C)	Annual precipitation (mm)
Windward							
Mountains	Hana, Hawaii	20 48 N	156 13 W	2	23.4	21.7–24.9	2060
No mountains	Belem, Brazil	123 S	48 29 W	14	25.9	25.4–26.5	2770
Leeward							
Mountains	Arica, Chile	18 28 S	70 22 W	29	18.5	15.4–22.2	1
No mountains	Bonthe, Sierra Leone	7 32 N	12 30 W	8	26.8	25.1–28.2	3718
Mid-latitude							
Windward							
Mountains	Hokitika, New Zealand	42 43 S	170 57 E	5	11	6.4–15.1	2721
No mountains	Bordeaux, France	44 51 N	00 42 W	48	12.3	5.2–19.6	900
Leeward							
Mountains	Tokyo, Japan	35 41 N	139 46 E	36	14.7	3.7–26.4	1563
No mountains	Boston, USA	42 22 N	71 01 W	9	10.8	–0.9 to 23.2	1086
High-latitude							
Windward							
Mountains	Tromso, Norway	69 39 N	18 57 E	114	3.3	–2.7 to 12	1119
No mountains	Spitsbergen, Norway	78 04 N	13 38 E	9	–3.8	–11.9 to 5.0	354
Leeward							
Mountains	Cape Tobin, ^b Greenland	70 24 N	21 58 W	42	–7.4	–16.3 to 2.7	Not available
No mountains	Resolute, Canada	74 43 N	94 59 W	64	–16.2	–31.8 to 4.6	131
Ice-influenced	Nord, Greenland	81 36 N	16 40 W	35	–16.5	–31.0 to 3.9	Not available

Source: Data are from Liljequist (1970) except for Hana data which are from the Western Region Climate Center

^aasl, above sea level

^bNote that Cape Tobin also represents an ice-influenced coast

the lowest temperatures are those formed by the meeting of ice masses and the sea such as the Greenland and Antarctic ice cap coasts. Such coasts are often marked by strong, katabatic winds blowing from the ice cap. Katabatic winds are winds formed from cold, relatively dense air flowing down a topographic feature. The climate of the other categories of coasts is intermediate between the extremes described here.

One of the most important features of coastal macroclimate is the storm. Both tropical storms (hurricanes) and mid-latitude storms or cyclones heavily impact the geologic, biologic, and human dimensions of a coast. Sometimes it is the presence of the coast itself that helps cause or enhance the storms. Cyclogenesis is enhanced, for example, at the east coast of the United States as upper-level disturbances cross the Appalachian mountains and then cross the strong baroclinic zone of the atmosphere near the coast. At the coast, the cyclones flow parallel to the landmass in winter and a strong cooling of air on the landward side leads to the formation of macroscale fronts (Rotunno 1994).

Considerable time is spent researching the frequency and magnitude of coastal storms. Waves generated by mid-latitude storms, for example, are responsible for much of the coastal erosion that occurs along the Atlantic coast of the United States. Dolan et al. (1988) found that at this location, the five-month period December through April had 63% of all the storms for the area. They also found a systematic increase between 1942 and 1984 in the number of storms occurring in May and October resulted in the lengthening of the storm season and a more abrupt transition between the winter and summer storm regime. The authors were able to construct a map of the mean cyclone track and frequency of all the mid-latitude storms in their data set that produced significant waves at Cape Hatteras. The storm track runs Southwest to Northeast following the coast of the Carolinas and continuing out to sea at Cape Hatteras. Locally the storms following this track are called Nor'easters and they often do considerable damage to the coast. The secular variation of such storms is very important. Hayden (1981) discovered that in the early years of the twentieth century there was a trend toward increased cyclone frequency over marine, as opposed to continental, areas of the Atlantic US coast. He also found decadal changes in coastal cyclogenesis with a maximum in the record up to 1978 occurring in the 1950s.

It is possible that decadal changes in the frequency of storms whose predominant wind directions is different can lead to marked geographic effects. At Hog Island, Virginia there is a clear dynamic coastline change pattern: erosion occurs on the north when accretion occurs on the south, or vice versa. This kind of cyclic process of coastline changes may repeat every 190 years on Hog Island. Currently, the southern part of Hog Island just ended the erosion processes, while the northern part just ended the accretion processes. It is hypothesized that the changing geography of the island on

this timescale may be related to the frequency of storms that either have winds with a predominant northeasterly or southeasterly component.

Damage to coasts is also done by Tropical Cyclones that go by different names in different parts of the world – hurricanes, for example, in the Atlantic and typhoons in Southeast Asia. For a tropical cyclone to be called a hurricane there must be a sustained wind of 33.5 m/s (75 mph) or more. Such systems can have diameters exceeding 300 km. Very large typhoons in the western North Pacific can have diameters up to 3000 km and are sometimes called super-typhoons. Hurricanes are placed into five categories in the United States. Category 1 has wind speeds of 119–153 kph; category 2, 154–177 kph; category 3, 178–209 kph; category 4, 210–249 kph; and category 5, has wind speeds in excess of 249 kph. In the following discussion, I will use the term hurricane as being synonymous to tropical cyclone.

Hurricanes are features of coastal climate mainly during the late summer and autumn seasons since they depend upon high Sea Surface Temperatures (SSTs) as one factor in their initial formation. Hurricanes impact the tropical, subtropical, and mid-latitude coastlines that are mainly on the east coasts of continents or coasts that face south in the Northern Hemisphere and vice versa. Hurricanes usually weaken after passing the coastline and moving inland because they depend upon the warm moist ocean surface for their energy. Besides high winds the storms are noteworthy for bringing large quantities of rain to the coast. Sometimes tornadoes are spawned as well. Hurricanes represent an interesting interaction between meso- and macro-climatology since their formation and track once formed is often a function of the macroscale features such as upper atmosphere wind fields. Much coastal damage is done by the interaction of the hurricane with the ocean. The resulting formation of storm surges brings extra high flows of ocean water over the coastline. The storm surge of a category 5 hurricane could be 5.5 m (18 ft) above normal sea level.

As with mid-latitude cyclones, interdecadal variability in frequency is well marked in hurricanes. Hurricane frequency in the Atlantic decreased from the 1890s to the 1920s and 1930s, increased during the 1940s to 1970s, and then decreased in the last part of the twentieth century (Riehl 1987). The latter decrease coincided with the development of much coastal property in the Southeastern coast of the United States.

Ice-influenced coastlines have their own interesting climatological features besides the katabatic winds already mentioned. An interaction between macro and mesoscale events gives rise to strips of open coastal water surrounded by ice-covered ocean. The open water areas are called polynyas and they typically open and close over the time period of a few days. They are partly formed by coastal winds pushing the ice offshore and are associated with a complex set of water and atmosphere density and flow interactions (Chapman 1999). The polynyas, sometimes called leads, are large sources of

sensible heat to the atmosphere in winter with heat fluxes locally exceeding 500 W/m^2 . In summer, the comparatively low albedo of the open water compared with that of nearby ice allows the absorption of solar radiation that warms the water. The fractional lead coverage in the Arctic Ocean as a whole is 1% in winter and greater than 20% in late summer (Curry and Webster 1999).

There are also phenomena of dynamic climatology existing where coastlines are backed by mountain areas. Rotunno (1994) describes how Kelvin waves may propagate along the basin-wall-like sides of an ocean basin such as at the Pacific coast of North America. Here also strong alongshore jet streams have been documented as a result of the mountains. Phenomena that appear similar to flow separation in classical fluid dynamics have been found to occur in the lee of capes and other coastline salients.

Temporal Climate Variability in Coastal Climates

All coastal climates are subject to climate variability on a variety of timescales. The presence of the ocean water has a moderating effect on the climate of the nearby land at all the timescales. Climate variability at coastlines is therefore usually less than that in interior continental landmasses but can still be significant especially at the longer timescales. We may examine an example from the PNW of the United States to demonstrate the kinds of climatic variability that can occur at, or near, a coast over a variety of timescales.

Major changes in the climate of the PNW from glacial to interglacial have been described by Warona and Whitlock (1995). The extreme glacial climate, a little inland from the coast, had a mean annual temperature that was likely to have been 7°C lower than today with the January temperature 14°C lower. Conditions were also probably drier than today with the Polar Front jet stream flowing well to the south. After the last retreat of the ice the climate was warmer. Between 10,000 and 5600 BP (Early Holocene) temperature is believed to have increased and effective moisture decreased as summer droughts became a climatic feature associated with an expansion of the subtropical high pressure in the eastern Pacific Ocean. On the next timescale down, some cyclic coastal climatic behavior has been suggested. Sediment cores from anoxic basins reveal a 60–70-year cycle of warming and cooling over the past 1500 years (US GLOBEC 1994). The last cycle in this pattern, which ecologically is associated with sardine dominance over anchovy and vice versa, is seen with warming around 1925, cooling at 1948, and a return to warming about 1977. Ware and Thompson (1991) regard the 60-year cycle in Southern California data as being consistent with their belief that pelagic fishes are responding to a 40–60-year oscillation in primary and secondary production that, in turn, is being forced by a long-period oscillation in wind-

induced upwelling. Trenberth (1993) reviews coupled General Circulation Model (GCM) output and suggests, under current projections of CO_2 increase, the western coast of North America is expected to warm by about $1.0\text{--}1.5^\circ\text{C}$ by the year 2030 and that northern regions will warm more rapidly than southerly regions. The hydrologic cycle is expected to intensify by about 10% leading to greater amounts of evaporation and precipitation. Trenberth (1995) has also suggested that the relative frequency of the El Niño mode in the past decade “may well be one of the primary manifestations of anthropogenic global warming.” There are also interdecadal scale variations in PNW coastal climate. Mantua et al. (1996) described a Pacific Decadal Oscillation (PDO) in SSTs that has many implications for salmon production. It has been shown that the PDO manifests itself in coastal climate by giving rise to four distinct climatic periods. The periods are 1896–1915, wet and cool, 1915–46, dry and warm; 1947–75, wet and cool; and 1976–94, dry and warm. The dominant source of variation on the quasi-quintennial scale is related to the El Niño–Southern Oscillation (ENSO). Low Southern Oscillation Index (SOI) values (El Niños) are correlated with reduced precipitation (Redmond and Koch 1991), snowpack and streamflow in the PNW (Cayan and Webb 1992). The PNW also tends to display warmer winter temperatures and drier than usual winters under El Niños conditions (Redmond and Koch 1991). At the opposite end of the oscillation Kahya and Dracup (1993) are among several investigators to demonstrate that high precipitation and high streamflow values in the coastal PNW are often associated with La Niña years. Superimposed on ENSO-scale climatic variability is the variability on the interannual and shorter timescales such as the passage of individual mid-latitude cyclones and anticyclones.

Future Research

There is an unlimited amount of work to do on coastal climates. Working through the scales we may suggest the following areas of important research. First, it would be interesting to document the differences in the surface energy budget values across the different surfaces of the subcomponents of the shoreline – barrier island, marsh, lagoon, etc. Second, it is important to find out what are the interactions between the mesoscale and macroscale events with respect to coastal climates. How, for example, do sea breeze fronts affect other weather systems occurring inland? Much more has to be found out about the effect of varying SSTs on weather systems passing over them. This is particularly the case where SSTs show a large range of different values over a short distance. The Gulf Stream ocean current and areas of coastal upwelling water are cases in point. There are many local dynamic features that need investigation such as the trapping

of atmospheric waves under marine temperature inversions and low-level coastal atmospheric jet streams. Interdecadal variation in the climatic variables of coastal climate must be established and, if possible, explained. This is important because of the huge amount of coastal development that has happened and will continue to occur worldwide. Finally, the affect of potential global climate change on coastal climates over the next 100 years must be studied. This topic has not received too much attention so far because the global climate models tend to operate at a coarser spatial scale than one that allows a coast to be specifically considered.

Cross-References

- [Climate Patterns in the Coastal Zone](#)
- [Coastal Temperature Trends](#)
- [Desert Coasts](#)
- [El Niño–Southern Oscillation \(ENSO\)](#)
- [Geographical Coastal Zonality](#)
- [Ice-Bordered Coasts](#)
- [Meteorological Effects on Coasts](#)
- [Storm Surge](#)

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Coastal Currents

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Introduction

Coastal currents are coherent water masses in motion that are found in the region between the coastline and the edge

of the continental shelf. Coastal currents are important because the coastal zone is the place where most nutrients, pollutants, and sediments are introduced into the ocean, and where most larvae are generated and dispersed. Coastal currents are responsible for the transport and dispersal of these biological, chemical, and geological tracers in the water. For example, predictions of the advection and diffusion of spilled oil is dependent upon knowledge of the currents near the coast. Coastal currents often are considered as being made up of two components, alongshore, or parallel to the coast, and cross-shore, or perpendicular to the coast. Away from the influence of inlets or river mouths, cross-shore flows are typically weaker than alongshore flows, yet the larger gradients in cross-shore properties make the cross-shore flows better at redistributing those tracers. In addition, if a current is periodic, such as a tidal current, even though it may be quite strong, it may not result in significant net transport. A non-periodic event-driven current such as from a storm may result in a significant net transport. Some currents that are persistent and result in significant transport of water are named after an adjacent landmass, such as the California Current and the Kuroshio Current. In general, currents are classified by the processes that form them (Wright 1995). Some of the most important currents that exist near the coast outside of the surf zone include (1) wave-driven currents, (2) tidal currents, (3) wind-driven currents, and (4) buoyant plumes.

Types of Coastal Currents

Wave-Driven Currents

Periodic water particle movements are associated with the passage of wind-generated surface gravity waves. The back and forth currents due to waves have a relatively short period, up to about 20 s. The horizontal orbital velocity of water motion is a function of wave amplitude, wave period, wavelength, and water depth (Komar 1998). According to linear wave theory, the water motion under waves in deep water is circular, with horizontal motions being comparable to vertical motions. The diameter of the circular motion decreases exponentially with water depth, so that in deep water, the influence of the surface wave can no longer be felt. In intermediate water depths the motions are elliptical, becoming smaller and flatter as the bottom is approached, and in shallow water the water motions are back and forth. In deep water, where linear wave theory is best applied, there is no net transport of water due to the passage of a wave. In shallow water, however, linear wave theory breaks down. According to nonlinear Stokes theory, the velocity magnitude is increased and the duration shortened under the crest of the wave, and the velocity magnitude is decreased and the duration lengthened under the trough. The result of this asymmetry in orbital motion is a

non-periodic current, or mass transport in the direction of wave propagation, which is called Stokes drift. This mean current can be large for large amplitude and short period waves, but is a small fraction of the instantaneous wave speed, only a few centimeters per second for long period swell.

Tidal Currents

Tides and tidal currents in the ocean basins are produced by the interaction of a rotating earth with the gravitational forces of the moon and the sun (see “► Tides”) (Mofjeld 1976; Bowden 1983). These oceanic tides impinge on the continental shelf and produce sea-level oscillations and currents in the coastal zone at tidal frequencies. The dominant fluctuations are either semi-diurnal (twice a day), diurnal (once a day), or a combination of the two. For example, on the East Coast of the United States the tides are semi-diurnal, whereas on the Gulf Coast they are diurnal. The difference is due to the interaction between the forcing and the natural frequencies of the tidal basin. Most coastal environments also have longer variations in the tides and tidal currents with periods of about 2 weeks. These fortnightly variations result in stronger spring tides and weaker neap tides. Tidal heights (up to a few meters) and tidal currents (up to several tens of centimeters per second) are periodic, thus they result in little net transport, despite excursions of several kilometers per tide. The horizontal component of tidal currents on the continental shelf tends to change direction and amplitude during each tidal cycle. An arrow denoting the current's speed and direction would trace an ellipse with each tidal period. If the direction of the current is restricted by a boundary, as near a coast, then the ellipse will become more elongated, with the long axis aligned parallel to the coast. The speed of a tidal current may increase when the current goes around a headland, passes through a restriction, or passes from deeper to shallower water.

Wind-Driven Currents

The drag of the wind on the sea surface results in the formation of waves and in the generation of a surface current. The wind-generated current is strongest at the surface and decreases with depth. The magnitude of the surface current depends on the wind speed and on the roughness of the sea surface due to waves, which effects the transfer of energy from the wind to the water. Based on field observations, the magnitude of the surface current is approximately 3% of the wind speed and the direction of the wind-driven current is nearly aligned with the direction of the wind, but not exactly. Due to the rotation of the earth, the Coriolis force causes the wind-generated current to veer to the right of the wind in the Northern Hemisphere and to the left of the wind in the Southern Hemisphere. Various theories predict that the surface current will be directed between 10° and 45° to the right (left) of the wind in the Northern (Southern) Hemisphere, and

that the current vector will rotate further to the right (left) with increasing depth below the surface. This rotating current direction and decreasing speed with depth is known as the Ekman spiral. The total transport averaged over the entire Ekman spiral is at right angles to the direction of the wind and is known as Ekman transport. The turning of the upper layer is reduced, however, in shallow water where the water depth is less than the Ekman depth.

Upwelling and Downwelling Circulation

When the wind blows parallel to the coast and the wind-driven Ekman transport is away from the coast, then the water near the coast will set down causing a lowering of sea level. The sloped sea level will induce an onshore-directed pressure gradient, which when balanced against the Coriolis force will lead to a shore-parallel geostrophic flow in the same direction as the wind beneath the surface Ekman layer. In addition, water near the coast will rise, or upwell, to replace the surface water that is transported offshore. Upwelling circulation is very important because it brings cold, nutrient-rich water from depth to the surface, which can enhance biological productivity. Many of the largest producing fisheries in the world occur in regions of coastal upwelling. Downwelling flows result when the wind is in the opposite direction, causing onshore-directed surface Ekman transport. In the downwelling case, sea level rises against the coast as surface waters are brought from offshore to onshore. In some cases, there can be an interaction between the shore-parallel wind-driven flow and the upwelling or downwelling waters that enhances the coastal flow. As an example, Korso et al. (1997) described how cold upwelled water along the coast of Oregon and southward-directed wind-driven surface currents interact resulting in an increased coastal jet occurring along the upwelled front.

Buoyant Plumes

Currents arise in the coastal ocean from the flow of freshwater out of rivers and estuaries. Currents result from both the difference in the height of the water between the river and the ocean and from the density difference between the fresh water and the salty ocean water. Upon entering the ocean, the positively buoyant plume will spread laterally on top of the ocean water and thin. After leaving the confines of the river or estuary, the plume will also be acted on by the Coriolis force. The degree of turning of the plume due to the earth's rotation depends on the width of the river mouth, the thickness of the plume, and the Coriolis parameter, which varies with latitude (Garvine 1987). The boundary of the plume where the water density changes rapidly is known as a front. Fronts are important because they mark zones of flow convergence. Plume currents can be affected by existing tidal currents and by the wind. For example, Stumpf et al. (1993) observed a buoyant plume move at speeds of up to 11% of the wind speed. The

enhanced speed of the plume relative to normal surface drift is due to reduced friction between the buoyant plume and the underlying ocean water.

Coastal Currents and Boundaries

Regardless of their origin, coastal currents are effected by their proximity to a boundary, either the seafloor or the atmosphere. Boundaries tend to modify ocean currents through the action of friction. The regions, or layers, in which currents are modified are called boundary layers. Boundary layers near the bed and near the surface can be 5–20 m thick, so on the inner and middle continental shelf, the dynamics of the boundary layers can make up a significant percentage of the water column. In the bottom-boundary layer the flow depends not only on the near-bottom current, but also on the presence of surface gravity waves. This *wave–current interaction* (*q. v.*) tends to modify the near-bed flow. Currents in the bottom-boundary layer can also be complicated by the presence of density stratification and bed morphology (see “► [Shelf Processes](#)”).

In a narrow zone immediately adjacent to the coast, near-shore currents result when wave asymmetry and breaking cause significant mass transport of water in the direction of the shore. Important nearshore currents include undertow, rip currents, and longshore currents (Komar 1998) (see “► [Surf Zone Processes](#)”).

At the outer boundary of the coastal zone, open-ocean currents flowing along the continental slope can at times influence coastal currents. For example, ocean currents such as the California Undercurrent off the US west coast can influence shelf currents by impinging onto the shelf, by entraining water from the shelf, or by inducing flow on the shelf through pressure gradients between the two water masses.

Measurement Techniques

Ocean currents are studied experimentally, through field measurements, and theoretically, through the use of mathematical models. Currents are measured with a variety of instruments, including mechanical rotor-type, electro-magnetic, and acoustic current meters. The choice of which current meter to use depends on among other things the environment in which the measurements are made. In the surf zone, currents are typically measured by mounting electro-magnetic current meters to fixed poles jettied into the bottom. Divers deploy the current meters and the data are logged in real-time through a cable to shore. Shelf and slope currents are measured using current meters on a mooring with a weight on the bottom to anchor the mooring in place and a float at or near the surface to hold the mooring

vertical. Alternatively, current meters are placed on a tripod, a pyramid-shaped frame that sits on the bottom. Current meters deployed on the shelf and in deeper water are self-contained, requiring an underwater power supply and data acquisition system to store the data until the instrument is recovered. In the past decade acoustic current meters that provide a profile of currents through the water column, as opposed to at a single point, have been used extensively to provide detailed information on coastal currents.

Cross-References

- [Pressure Gradient Force](#)
- [Rip Currents](#)
- [Shelf Processes](#)
- [Surf Zone Processes](#)
- [Tides](#)
- [Vorticity](#)
- [Wave–Current Interaction](#)
- [Waves](#)

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Coastal Dynamics

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Definition

Coastal dynamics refers to physical processes (dynamic factors) like waves, wind, currents, and sediment supply that affect and shape the coastal morphology independently of changes in sea level or land level (Fig. 1).

Introduction

The shorelevel is the point where the land level and the sea level cross each other. The morphology of the shore depends on several factors.

The tidal cycle records the daily variations between low-tide and high-tide (Melchior 1978), which the shore morphology is adjusting to.

For the trained eye of a sea level specialist, it is usually possible to identify the mean sea level and high-tide level with an accuracy of a few centimeter (Mörner 2010).

The position and variability of the shorelevel is determined by several different dynamic, oceanic, and land level factors, as illustrated in Fig. 1. Changes in any of these variables will lead to a change in the position of the shorelevel in time and space.

The coastal dynamic factors affect, shape, and re-shape the shoreline, either regardless of what happens with the land and sea levels or in interaction with such changes (Fig. 1).

For meaningful analyses of sea level changes and coastal evolution, it is necessary to consider all the factors illustrated in Fig. 1. Ignoring to do so, may lead to unfounded statements and conclusions with respect to sea level changes and coastal protection.

The study of coastal dynamics has a vital mission to fulfil, and it has grown into a successful international organization (e.g., CD 2017).

Tides and Multi-decadal Cycles and Oscillations

Besides the daily, monthly, and annual tidal cycles and the 18.6 years lunar tidal cycle (Melchior 1978), there are also longer-term cycles like the 60-years cycle (Chambers et al. 2012; Parker 2013; Hansen 2015) and the oceanic multi-decadal “oscillations” (the Pacific Decadal Oscillation, PDO; the North Atlantic Oscillation, NAO; Atlantic Multi-decadal Oscillation, AMO; etc.) to consider.

The existence of multi-decadal cycles and oscillations implies that our analyses must extend over several cycles in order to be able to record true mean values.

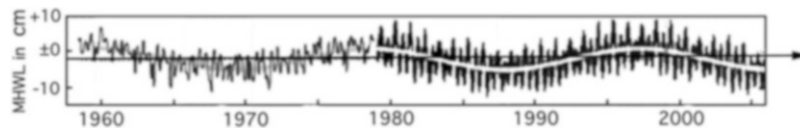
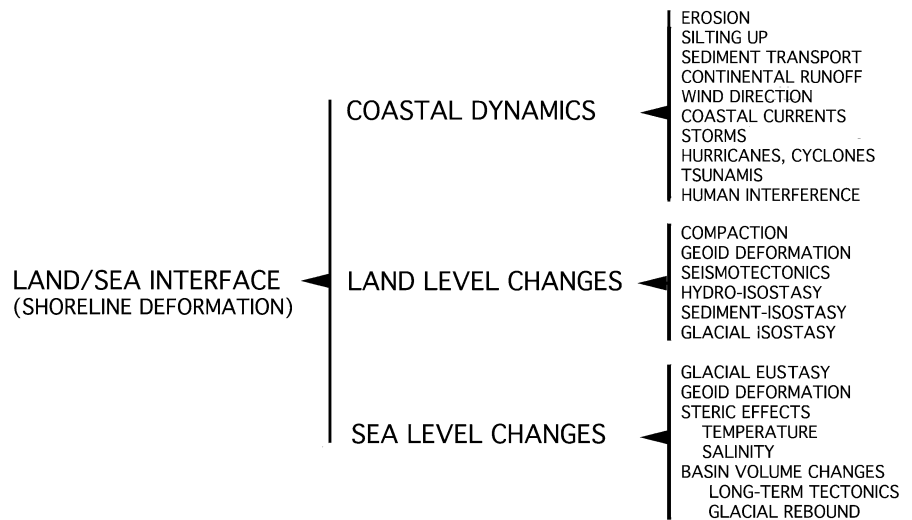
This is illustrated in Fig. 2 with respect to the 18.6 years cycle covering two full cycles (Mörner 2010). The trend over the last 9 years is a fall in sea level by about 6 cm, while the true mean sea level change over the entire 49-years period is zero (± 0.0 cm).

The Main Coastal Dynamic Variables

There are several different coastal processes, which are capable of changing the coastal conditions and shore morphology independently of changes in sea level or land level (Fig. 1). Sometimes, however, the dynamic changes occur in

Coastal Dynamics, Fig. 1

Many different variables affect the stability of the land/sea interface (the shoreline). They fall within three major groups of changes; the coastal dynamics, the level of the land surface, and the level of the sea (modified from Mörrer 2010). For proper sea level analyses, it is fundamental to understand and consider all variables involved in shore deformation in time and space



Coastal Dynamics, Fig. 2 Changes in annual mean-high-water level (MHWL) along the French Guiana–Surinam coasts from 1958 to 2006 (redrawn from Gratiot et al. 2008). The inter-annual changes in sea level

exhibit a clear 18.6 years' tidal cycle where the mean value has remained virtually stable throughout the entire 49-years period

mutual interaction with changes in ocean level or land level. The coastal dynamic factors deform the shorelevel or point of sea/land interface laterally (Fig. 3) in opposite to the change in sea level or land level, which deforms the shorelevel vertically upward (sea level rise, land level fall, or sediment compaction) or downward (sea level fall or land level rise).

Coastal Erosion

Erosion of coastal segments may be driven by several factors, and is, by no means, a straightforward indication of sea level rise. This is also the case at the deformation of the so-called “Bruun equilibrium profile” in shore morphology (Bruun 1962; Schwartz 1967).

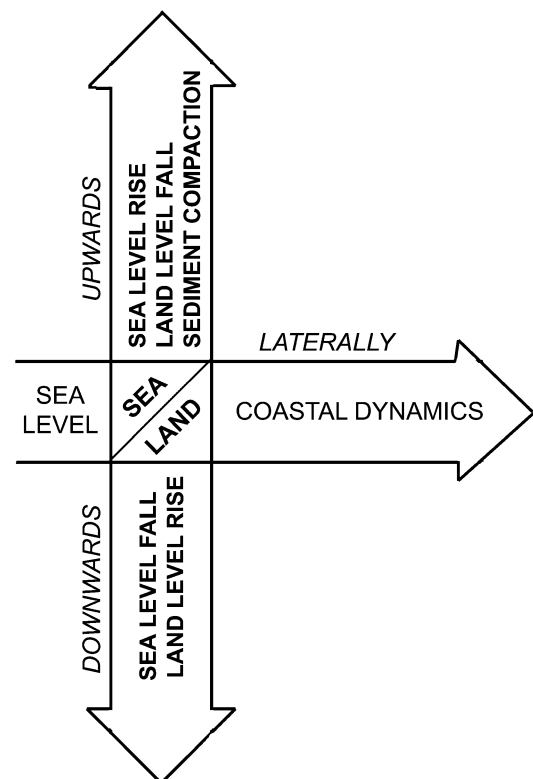
According to Science Daily (2017), “coastal erosion is common phrase referring to the loss of subaerial landmass into a sea or lake due to natural processes such as waves, winds and tides, or even due to human interference.”

Active coastal erosion implies a constant lateral re-adjustment of the position of local sea level.

Silting Up

When the sediment supply is strong, the sediments may grow vertically by silting up as well as horizontally by coastal progradation.

As a consequence, the land/sea interface may become displaced.



Coastal Dynamics, Fig. 3 Coastal dynamics deform the shore laterally (horizontally), while the changes in sea level or land level deforms the shorelevel vertically

In river deltas, there is a continual sediment accumulation forcing the shore to move seaward.

The motion of sediments is a key factor for coastal environment and its dynamical changes in time and space (e.g., Sheng 1986; Soulsby and Whitehouse 1997).

Sediment Supply

Some coastal segments are fed by a constant sediment supply from nearby rivers. Changing the sediment supply may lead to sediment starvation of a beach. Due to changes in runoff (below), the sediment supply may change and thereby generate sediment starvation.

If the shore morphology is changed, the position of mean sea level may become displaced laterally, despite sea level remaining constant.

Continental Runoff

Changes in the supply of water and sediments in a river system may generate strong effects in the adjacent coastal systems (e.g., Milliman and Ren 1995). Delta areas may starve in parts or as a whole. Beach fed by constant sediment supply may starve, and even disappear.

Building of large river dams or diverting water for irrigation affects the amount of water and sediments reaching the delta area.

In Bangladesh, for example, large dams decreased the water masses reaching the sea, and the salinity increased, which was misunderstood in terms of a sea level rise.

Natural changes in the runoff via rivers are controlled by alternations in the catchment area between wetter and dryer periods.

Satellite altimetry records a higher sea level over the eastern part of the Mediterranean Sea and the Black Sea. This is a function of increased runoff from the huge catchment area of the Black Sea and has nothing to do with any rise in sea level (Mörner 2005).

Changes in Wind Direction

Changes in wind direction, especially the prevailing wind direction, may significantly affect the coastal conditions, and lead to significant deformations of the shore morphology. Even the direction of long-shore drift may change. We have numerous examples of this all over the globe. As a result, the line denoting the sea/land interface may become displaced laterally (landward or seaward) without any real vertical change in sea level.

Coastal Currents

The coastal currents are essential for the local to regional shore stability (e.g., Gelfenbaum 2005). The coastal currents may change their course because of factors like wind direction, silting up, sediment supply, water runoff, and, not least, human interference with the coastal conditions. This may generate significant shoreline deformation in the horizontal dimension.

Storms

A strong or extremely strong storm may rapidly deform a shore. Some beach-ridges were formed by a single storm. The extreme waves may at the same time cause severe erosion. With reorganization of the shore morphology, the point of sea/land interface would become displaced in horizontal dimension.

Hurricanes and Cyclones

Hurricanes or cyclones are extreme storms that hit a coast with very strong wind and wave forces. The coastal destruction may be strong, even disastrous. The shore morphology may become seriously altered with a new land/sea configuration, despite the actual sea level remaining constant.

Tsunamis

Tsunami events are characterized by high and fast-moving waves. The big events may sometimes rise to really destructive wave dimensions (e.g., the 2011 event in Japan, the 2004 event in the Indian Ocean, the 1755 event at Lisbon). The coastal damage and reconfiguration of the land/sea distribution may become very severe. Tanaka et al. (2012) recorded extensive coastal deformation at the 2011 tsunami event in Japan, just to refer to one recent example.

Human Interference

Human activity may often lead to severe coastal damage and erosion. Just the very strongly increase in beach and shore utilization for pleasure and tourism put pressure on the coastal stability.

Clearance of mangrove vegetation often leads to severe erosion, and has become serious problem in recent decades.

Removing the protecting mangrove vegetation generates a strongly increased vulnerability to storm destruction.

Quarrying of shore sand, beach-rock, corals, and eolianites are all destructive activities for shore persistence.

Even the massive removal of sea cucumbers from the shallow waters off the beaches in Fiji led to removal of sand, which opened the shore for erosion.

Our regulation of rivers affects coastal runoff and sediment supply to the sea (above) affecting the coastal shore stability.

Building of protecting jetties and seawalls often lead to unforeseen effects including increased erosion.

Summary

The stability of a shore in time and space depends of several different factors. Besides changes in sea level and in land level, there are a number of dynamic factors operating. Those factors may generate significant changes in shore morphology and shore position without being linked to any changes in sea level (eustasy) or land level (compaction or tectonics).

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Coastal Erosion

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Synonyms

Beach erosion; Shore erosion; Coastal abrasion

Definition

Erosion (from Latin *erosio*, to wear, to wear away) refers to surface processes acting to remove material. In the coastal zone, the wearing away and transporting away of sediment particles and rock fragments are known as *coastal erosion*. *Coastal erosion* applies to the entire coastal zone (e.g., Jolliffe 1982; Pranzini and Williams 2013), i.e., the zone extending from the offshore environment to the coastal onshore area, but primarily refers to the inland displacement of the point of sea/land interface (the mean sea level). The loss of shore deposits (e.g., beach and dune deposits, marshlands) and landmasses (e.g., rocky headlands) into the sea is driven both by natural processes that include waves, wind, currents, and slides and by human interference with shore processes.

Introduction

The coastal shape or assemblage of landform configurations is conditioned by the extent geology, ground relief, and shoreline character (Owens 1982). The character or nature of the shore is developed in a wide spectrum of landforms from equatorial to polar coastal segments (Valentin 1952; McGill 1958; Carter and Woodroffe 1994). Many coastal landforms are ubiquitous (e.g., beaches, dunes, spits and bars, cliffs), but some specialized forms are latitudinally dependent as in the case of coral reefs, beach rock, mangroves (Spalding and Grenfell 1997), and boulder beaches on polar (glacial and paraglacial) shores (e.g., John and Sugden 1975; Carter and Woodroffe 1994). These kinds of prevailing coastal conditions set the coastal character that is in a more or less constant state of flux. Those conditions, which are likely to change over time for a number of reasons (below), may initiate coastal erosion or re-shaping of the shore. Study of erosion and sediment redistribution is called *coastal morphodynamics*.

The change from coastal stability to coastal erosion is caused by many different processes that may involve neotectonics, changes in prevailing wind direction, deviation or reinforcement of coastal currents, reestablishment of equilibrium profiles (e.g., Dean 1991; Dean and Dalrymple 2004), sea-level rise, sea-level fall, exceptional storms, hurricanes/cyclones/typhoons, and tsunami events (below).

Origin of Coastal Erosion

A suite of different physical processes are involved in the overall concept of coastal erosion, which is a catchall term for many different specific processes. *Coastal abrasion* refers to the mechanical erosion by particles in the moving waves and occurs when breaking waves containing rock

particles (sand, gravel, stones) erode the shore or headland. The correlative geomorphological process is also known as *corrasion*. When waves cause particles and pieces of rock by collision, grinding and chipping to become progressively smaller, smoother, and rounder, it is termed *attrition*. Bedrock surfaces may also become weathered and eroded by *solution* (especially in the case of sandstones and limestones) and *corrosion*. *Hydraulic action* and fracturing occur when waves (and wind) compress air or water into cracks. In combination with freezing and thawing, this may lead to the fracturing off of rock fragments.

All coastal segments are in principle susceptible to erosive forces that lead to the displacement of sediments and rocks. Of the several processes and factors that may generate coastal erosion and remodelling of the shoreline, most erosion is often accompanied by the interaction of multiple biophysical processes. While shore cliffs are in a more or less permanent stage of erosion, other coastal segments may move into (or out of) an erosional condition due to changes in ambient coastal conditions that may be short termed or relatively long lived. The processes involved in such changes (below) are of primary interest for meaningful interpretations of observed coastal erosion (e.g., Pranzini and Williams 2013; Mörner 2015).

Changes in Prevailing Wind Direction

Variations in prevailing wind directions may alter, or even revert, active shore-building conditions by affecting longshore drift. Erosion in updrift shore segments is often balanced by deposition in downdrift segments, but along some coasts, there is a decrease in sediment load of the longshore drift because sediments are lost offshore beyond the depth of closure by cross-shore currents (e.g., Finkl 2004). Because sediments in these cases are effectively lost from the coastal sediment budget, shorelines so affected are in retreat (retrograde). In the opposite situation where shorelines are prograding, the sequence may take the form of a complicated braid of shorelines where one system is cutting the previous one and where even the direction may revert as a function of changes in prevailing wind directions (e.g., Schröder 2015; Hansen et al. 2016). Under specialized conditions where there are areas of very high land uplift, as occurs in northern Sweden with an uplift rate of 10 mm/yr, erosion of an older beach-ridge system has been observed in progradational systems which at first seems counterintuitive.

Changes in the prevailing wind direction have been recorded at many coastal sites. The longest continual record of changes in prevailing wind direction covers 10,000 years and is established by alternations in longshore drift, with shoreline erosion updrift and downdrift building out of beach ridges in southwest Sweden (Mörner 1980).

Changes in Coastal Currents

The term *coastal currents* refers to water masses that are in a roughly shore-parallel motion between the shore and the shelf break (Gelfenbaum 2005). In this case, we only consider the currents affecting the shore, however. These currents, which constitute relatively uniform drifts in deeper water adjacent to the surf zone, may be tidal currents, transient, wind-driven currents, or currents associated with the distribution of mass in local waters. Currents that drive coastal morphodynamics and which affect shore processes are linked to winds, waves, tides, and offshore morphology. Longshore currents are driven by waves and winds, generating the longshore drift responsible for the building of beach ridges, spits, and spurs. Tidal currents can modify the coastline and offshore bottom topography. Cross-shore currents refer to currents that are directed perpendicularly (normal) to the shore.

The near-shore currents may change both in force and in direction over time, which may generate coastal erosion and/or reorganization of the shape and character of the beach. Currents, among other factors, largely determine the shape of configuration of the shore and along with waves determine the beach type (Short 2003).

Human interference with coastal (beach) processes due to the construction of seawalls, groins, and jetties along the shore often affect the coastal currents in an unwanted manner, generating increased problems instead of providing coastal protection (Pilkey and Wright 1988; Young et al. 2009; Mörner and Matlack-Klein 2017).

Changes in Runoff and Sediments Supply

Changes in runoff and sediment supply may lead to beach starvation and reorganization of the beach morphology. Lack of adequate sediment supply to maintain a healthy beach often occurs when human constructions decrease the amount of river water flowing to the sea or when climate changes to drier and warmer conditions (e.g., Leader et al. 1998; Walling and Fang 2003; Syvitski et al. 2005; Zhang et al. 2006). This situation has become an accelerating problem today because so many rivers have become utilized for irrigation and dammed for electrical power generation (e.g., Mehta et al. 2012). Major river deltas, deltaic environments, cheniers, and coastal marshlands are sometimes seriously eroded by the changes in river runoff (e.g., Milliman and Mei-e 1995; Wahid et al. 2007; Syvitski et al. 2009; Giosan 2014; Day et al. 2016).

Readjustment of Equilibrium Profile

The motions of the water due to waves, tides, and currents along a sandy coast tend to generate a more or less equilibrated balance between erosion and deposition that shapes the onshore-offshore relief (bottom topography or bathymetry) according to some sort of idealized *equilibrium profile*

(Brunn 1954, 1962; Schwartz 1967). In coastal engineering and geomorphology, this equilibrated shore profile is also known as *active shore profile* (e.g., Beech and Andrews 1993).

There is an extensive literature concerning coastal equilibrium profiles and what effects it may lead to (e.g., Brunn 1954, 1962; Strahler 1966; Schwartz 1967; Dean 1991; Pilkey et al. 1993; Cooper et al. 2000; Andrews et al. 2004; Mörner 2015). The problem is much more complicated, however, than just the reestablishment of an equilibrium profile deformed by sea-level rise. This concept is further discussed below and illustrated in Fig. 2.

According to Andrews et al. (2004): “The Bruun Rule has no power for predicting shoreline behaviour under rising sea-level and should be abandoned. It is a concept whose time has passed” (cf. Fig. 2). This opinion should invalidate previous claims (e.g., Titus 1986; Titus et al. 1991), but by no means is established in the literature, as there still appears to be conceptual value in the cognition of gross shore profile change due to erosion and sedimentation.

Instability of Cliff Sections

Coastal cliffs are steep escarpments at the coastline that form cliffed or abrasion coasts. In many instances these common erosional landforms are very scenic, if not majestic, shores as erosion of the cliff face exposes internal layering and other types of rock formations (e.g., Leary 2014). Coastal cliffs are usually formed in sedimentary bedrock, but they may also occur in thick till deposits in formerly glaciated areas, and in older marine beds, not to forget volcanic formations as well.

Cliffed coasts or *abrasion coasts* refer to coastal segments where wave action has led to the formation of steep, unstable cliffs where waves undercut the cliff at its base. Sudden cliff

collapses (slides and rock fall) are a more or less natural part of active sea cliffs that supply sediment to the littoral drift system. In addition to wave action undermining cliff faces, earthquakes may intensify erosive collapse as occurred, for example, during the M 5.7 earthquake in Christchurch, New Zealand, in 2016 (BBC 2016).

Coastal erosion is perhaps most dramatically expressed in coastal sections marked by impressive coastal cliffs (e.g., Leary 2014). Some examples of these erosional features include The Seven Sisters, Seaford Head, England; The Twelve Apostles, Great Ocean Drive, Australia; Stevns Klint (with the Cretaceous/Tertiary boundary section) in SE Zealand, Denmark; and Etretat, France, just to mention a few.

Bedrock cliffs and undercut notches are often linked to *wave-cut platforms* or *rock-cut platforms* (e.g., Sunamura 2015), in most cases denoting the local high-tide level (as shown in Fig. 1).

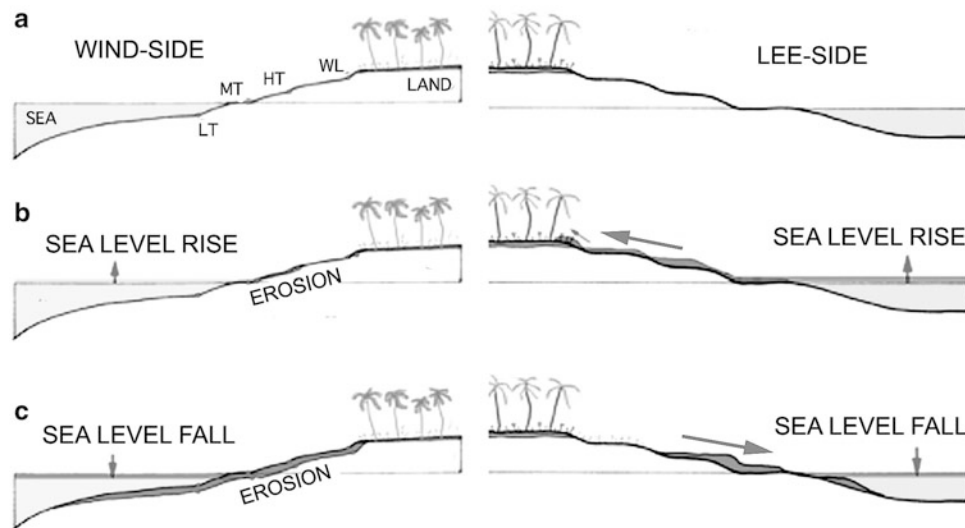
Sea-Level Changes

Coastal erosion is often interpreted as a sign of sea-level rise (Brunn 1954, 1962; Schwartz 1967; Titus 1986; Bird 1993; cf. point 4, above). This is not tenable, however (e.g., Andrews et al. 2004; Mörner 2015), because any change in the position of sea level, whether positive or negative, will affect the Bruun equilibrium profile with the potential of generating coastal erosion (Fig. 2). The relation between coastal erosion and sea-level change is far more complicated than suggested in the Bruun Rule, but there are some generalizations that can be made between sea-level stands and coastal erosion deposition. These generalizations occur outside the realm of human interference.

Coastal Erosion,

Fig. 1 Undercut notch in the bedrock with associated wave-cut platform (flat rock surface at the cliff base) grading over into the high-tide level of the sandy beach to the left in the background (Naviti Island, Fiji, photo; Mörner 2017)





Coastal Erosion, Fig. 2 Island wind- and lee-side morphology (a) and erosion/deposition in association with a sea-level rise (b) and a sea-level fall (c), respectively. Indicated on the exposed windward-side of the island (a) are the low-tide level (LT), mean-tide level (MT), high-tide level (HT), and swash limit (WL). When sea-level rises in (b), minor erosion takes place as the shore profile cuts inland. On the island's depositional lee-side, eroded materials are redeposited upward (arrow pointing inland) over previous land surfaces. When sea level falls in (c),

sediments on the shoreface are eroded as the shore profile equilibrates to the new morphodynamic conditions. On the lee-side, the eroded materials are redeposited seaward or down profile. As a result, the former beach is stranded above the new lower tide level and becomes a dry land surface that is exposed to biophysical weathering and vegetative overgrowth, i.e., as observed in the Maldives, for example (Mörner et al. 2004; Mörner 2015), and elsewhere as a general principle.

Because erosion in one segment of the coast implies deposition at another (e.g., Fairbridge 1961; Bird 1995; Mörner 2015), typically in an eroded updrift segment versus a depositional downdrift segment, one can use this morphodynamic balance to ascertain whether sea level is rising or falling.

A clue to modal changes in sea-level may be found in a depositional shoreline segment (e.g., on the lee-side of an island, as illustrated in Fig. 2) downcoast from the erosional site that was assumed to be caused by a change in sea level (e.g., Mörner et al. 2004; Mörner 2015).

Sea-level rise tends to deform the equilibrium shore profile, which may generate some degree of costal erosion (Fig. 2b) depending on the alongshore sediment supply. On the island's lee-side, the eroded sand is deposited in over previous land levels (as marked by an arrow in Fig. 2b). Stratigraphically, this implies that marine sand is deposited over a former land surface. This gives evidence of a sea-level rise (or, alternatively, the result of storm surge deposition).

A fall in sea level will inevitably lead to extensive erosion, by wave and current action, of the previous equilibrium offshore profile (Fig. 2c). Under a falling sea-level scenario, eroded sand will be redeposited at a lower level seaward (as marked by an arrow in Fig. 2c). Materials in the former beach, which had become stranded at and above the former sea-level position, become susceptible to subaerial weathering processes. This provides evidence of a fall in sea level (whether of local or regional dimensions).

Exceptional Storms

Most open-ocean costal segments are impacted by various types of high-energy conditions that feature meteohydrodynamic conditions with large waves, strong currents that often culminate in *storm surges* (e.g., Brennimeyer 1982; Aagaard and Masselink 1999) that can cause extensive erosion. When some storms are exceptionally vigorous and their intensity exceeds normal storminess, they are classified as once in a century storm or one in a thousand year storm. These exceptional storms cause exception coastal erosion. Sedimentary deposits left by these kinds of extreme storms are referred to as *tempestites* (Myrow and Southard 1996). In addition to the erosive impacts of these severe storms, they have potential to damage life and property (NOAA 2017).

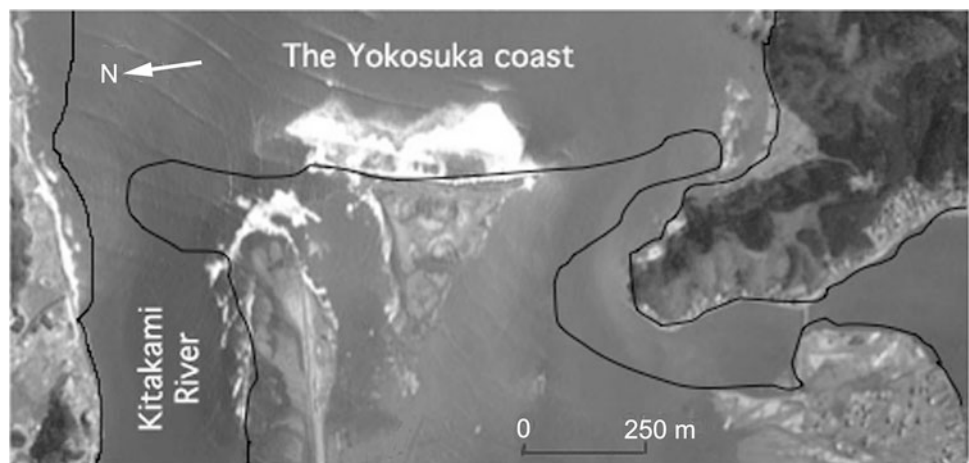
A particular coastal storm usually affects only parts or segments of a coast. Local storm impacts depend on a range of properties such as the nature of the shore materials (rock or sediments), degree of exposure to high-energy conditions, and coastal morphology. The same storm may cause severe destruction in one part of the coast, while in another part of the same coast, there may be only vague or insignificant impacts (Mörner 2015). Therefore, the degree of local erosion is not a straightforward measure of the severity of the storm as many factors are involved.

Ameliorative effects on coastal erosion are provided by mangrove vegetation that is quite effective in the reduction of *storm surge* (McIvor et al. 2012). Coastal erosion is

Coastal Erosion, Fig. 3 The Sidr Cyclone, which hit Bangladesh coasts in 2007, caused extensive coastal erosion, as evidenced by the exposed tree trunks in living position on the beach at low tide. The horizontal roots, previously covered with mud, indicate that about 80 cm of deltaic muds were removed by the cyclone and that no rise in sea level was involved (Photo at low-tide: Mörner 2009)



Coastal Erosion, Fig. 4 The shore of the Yokosuka coast in Japan before (black line) and after the Tohoku-Oki tsunami. The land/sea boundary configuration as of March 11 2011 was recorded by Tanaka et al. (2012) and here redrawn and modified by Mörner (2015). The coastal remodelling is very extensive.



exacerbated by deforestation of mangrove vegetation such as occurs in many tropical and subtropical coastal regions (e.g., Rahman et al. 2009). The loss of fringing mangrove forests, which protect the shore from erosion, results in a “coastal squeeze” that makes such shore more vulnerable to loss of mangrove sediments (e.g., Truong et al. 2017).

Hurricane-Cyclone Events

Hurricanes or tropical cyclones impact coasts with very strong forces that often modify the shoreline configuration via the processes of erosion deposition. Major coastline readjustments are thus often associated with these high-energy events. The nomenclature used – i.e., hurricanes, cyclones, and typhoons – differs for historical and geographical reasons, with a useful classification scheme to be found on Wikip (2017). The long-term activity trend of these large storms has been much debated. On a global base, tropical

cyclones have remained at about 87 ± 10 events per year. By modelling, Emanuel (2005) claimed: “the hurricane intensity has increased markedly since the mid-1970s.” By observational facts, however, Rojo-Garibaldi et al. (2016) demonstrate that the number of hurricanes from 1750 to 2010 in the Gulf of Mexico and the Caribbean Sea exhibits a decreasing trend toward the present, with anticorrelation to the number of sunspots. This observation has wide implications for understanding of coastal vulnerability to future erosion.

The Bangladesh region is regularly struck by big cyclones, often with tremendous erosional effects. The *Sidr Cyclone* hit the coasts in 2007 causing extensive erosion (Fig. 3) and exposed trees in living position along the beach at low tide. Removal of these coastal sediments was storm induced (Fig. 3) and had nothing to do with any purported rise in sea level (Mörner 2010).

Tsunami Events

Tsunami events usually lead to severe coastal erosion and remodelling of the shore (Fig. 4). The accumulation of redeposited material derived from a tsunami, termed *tsunamite*, is a special type of *tempestites* (Shiki and Yamazaki 2008). The tsunami events of Lisbon (1755), Messina (1908), Indonesia (2004), and Japan (2011) are all notorious for the erosion and remodelling effects onshore as well as offshore (e.g., Fonseca 2005; Tanaka et al. 2012). The Tohoku-Oki tsunami event in Japan (2011) had a maximum wave height of 19.5 m and a maximum run-up of 40 m extending 5.4 km inland (Lario et al. 2016). The erosion and remodelling of the coasts were extensive, as illustrated in Fig. 4 (based on Tanaka et al. 2012).

Human Interference

Human interference with the coast has a long history that was a major factor affecting coastal morphology and shoreline position due to induced erosion. This influence began with early agriculture in the late Stone Age and accelerated with intensified human activities and engineering constructions along coasts in historical time (e.g., Davis 1956). Coastal engineering interference with coastal processes often leads to negative implications for the sustainability of the beach environment (Pilkey and Wright 1988; Young et al. 2009; Finkl and Makowski 2010). On the whole, coastal protection is a complicated matter where consideration has to be given to a lot of different factors, some of which often remain unknown or uncertain. While it seems obvious that fishing utilizing explosives are likely to generate disastrous effects on coastal stability, it seems quite unforeseeable that even the harvesting of thousands of sea cucumbers may lead to severe coastal erosion (Mörner and Matlack-Klein 2017).

Human interference with the coastal runoff (Mehta et al. 2012), the sediment supply (Syvitski et al. 2005), land/sea configuration (Davis 1956), and installation of coastal protection constructions (Young et al. 2009) have a significant, sometimes even dominant, effect on coastal erosion and remodelling.

Summary

Coastal erosion is a key issue in Holocene coastal evolution (Carter and Woodroffe 1994; Davis 2009). However, coastal erosion can be generated by numerous different factors including human interference with coastal process (Davis 1956). Any changes in wind conditions, coastal currents, water runoff, sediment supply, cliff collapses, sea-level change, and episodes of severe storms and tsunamis have the potential of affecting coastal morphology and to affect the position (horizontally and/or vertically) of the shore by erosion or deposition.

Interpret of coastal erosion as a straightforward sign of sea-level rise can be misleading because sea-level fall (like many other natural dynamic processes) may also generate coastal erosion.

Cross-References

- Accretion and Erosion Waves on Beaches
- Barrier Island Formation and Development Modes
- Barrier Island Landforms
- Beach Erosion
- Bioengineered Shore Protection
- Bioerosion
- Changing Sea Levels
- Coastal Changes, Rapid
- Coastal dynamics
- Coastal Hazard Indicators
- Coastline Changes
- Cross-Shore Sediment Transport
- Depth of Closure on Sandy Coasts
- Dynamic Equilibrium of Beaches
- Erosion Processes
- Geomorphology and Sea Level
- Longshore Sediment Transport
- Meteorological Effects on Coasts
- Paleocoastlines
- Rock Coast Processes
- Sea-Level and Climate Change
- Shoreline Response to Littoral Drift Barriers
- Shore Protection Structures
- Swash Zone Dynamics
- Tsunami Deposits

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Coastal Erosion Management

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Definition

Coastal Erosion Management (CEM) is the dynamic, multidisciplinary, and interactive approach in responding to coastal erosion processes to prevent or mitigate economic and social losses. Ideally, such management should also seek to achieve preservation or conservation of natural habitats. CEM requires a detailed knowledge of coastal processes, sediment behavior, and interchange within the coastal zone (offshore, shore, and inland), and how changes, both natural and due to human activities, alter these controls of erosion (see “► Coastal Erosion” in this volume for more detail). The cycle of CEM begins with the identification of a perceived erosion problem, determining the cause(s) and understanding the erosion processes. This basis then guides planning, preparation, and decision-making as to the appropriate response, considering the time frame of expected results, including monitoring whether or not the solution is working. CEM is one aspect of overall coastal zone

management and usually follows the objective of an informed involvement and cooperation of all stakeholders to assess the societal goals in a given coastal eroding area, and to take actions to meet specific objectives. CEM should seek to balance environmental, economic, social, cultural, and recreational goals over the long-term, all within limits set by the natural dynamics of an area. In the past, erosion management responses often have not been successful for a variety of reasons including quick fixes sought in response to emergency events, or misdirected plans based on cost/benefit analysis, or merely placing sole emphasis on protecting buildings and infrastructure without concern for all other coastal aspects. Such short-term protective projects have increased or shifted the erosion problem, caused loss of habitat including economic drivers such as beaches and dunes, and resulted in more property and people in high-hazard zones, increasing vulnerability to future losses. Future CEM plans can be strengthened by having a broader basis for decision including:

- A long time frame point of view
- A stronger emphasis on the restoration and utilization of natural habitats for protection against erosion
- Planned removal of anthropogenic structures to restore sediment supply
- Factoring in both global and regional changes that will increase erosion problems (e.g., sea-level rise, tsunami potential, El Niño)

The spatial, temporal, environmental, physical, and social knowledge related to the area of interest forms the core of CEM to eliminate or at least minimize coastal erosion.

Introduction

Current projections indicate that coastal erosion processes will reach unmanageable proportions due to increasing urbanization during the on-going warming climate and sea level rise. Coastal erosion processes and their management become much more critical as coastlines constitute some of the most densely populated and developed land zones in the world. That concentration of population isn't likely to decline because the coastal zone is the ideal area for different productive activities (Gracia et al. 2018).

Coastal erosion is a normal process that has been active since land emerged from the sea. A suite of natural process and anthropogenic influences (most of the time interconnected) are involved in coastal erosion, which itself encompasses many different specific processes. Coastal erosion can occur as a slow pervasive process (i.e., due to ongoing subsidence) or rapid during extreme events (i.e., storm waves).

It is important to highlight that coastal erosion only becomes a problem when there is no room to accommodate the occurring changes (e.g., shoreline retreat). In this sense, coastal erosion is not an issue along areas where high, stable grounds back up the coast, but erosion is a severe problem for low elevated areas and uplands underlain by unstable materials, rural and urbanized. No matter the place, rates, and causes, coastal erosion is a process that currently raises a significant hazard to ecosystems and human activities and therefore must to be managed (Rangel-Buitrago et al. 2018).

World megacities (e.g., Jakarta, Manila) are located in coastal zones where the conjunction of geographic, economic, ecological, and historical conditions are attracting people, and stimulating different migration processes. Only during the last half century have world coastline populations recorded values close to 5% of the annual average for urban growth rate (UN-Habitat 2009). Under current and projected climate change scenarios, it has been estimated that along low coastlines, almost 35% of residences, if sited within 200 m from the sea, may be severely affected by erosion-related property losses over at least the next 50 years (UN-Habitat 2009).

All of the above make clear that the world needs sustainable development strategies accompanied by an optimal coastal erosion management. The coastal erosion management choice must be a function of a series of variables that include: erosion severity, causes, property rights, funding, legislation, and coastal use. In the same way, CEM must consist of all available knowledge, techniques, infrastructure, and institutional-political tools required to minimize or eliminate coastal erosion related impacts.

Coastal Erosion Management Approaches

The typology of coastal erosion management approaches can be summarized in five categories as follows.

Protection

The traditional approach to managing coastal erosion was to address the problem only at the point of its attack, and to out engineer Nature. And the objective was primarily to protect the fast-land property, be it buildings or farm land. Protection of property is still the driving force in CEM, and related issues such as disruption of sediment supply, beach loss, and habitat destruction are not addressed. This approach consists in the preservation of vulnerable areas (i.e., population centers, economic activities, and resources) by means of constructing hard structures, or alternatively soft protection measures which does include beach restoration. Both hard and soft approaches seek to eliminate or mitigate erosional damage to the coast by use of barriers to reduce wave impact and control hydraulic processes that influence sediment transport (Fig. 1).

The measures to be applied may be summarized in four specific structural solutions that can be developed stand-alone or in combination, depending on the particular conditions of the coast where they will be applied.

Hard Defenses

This approach has been used for CEM since ancient times. Structures are built of different kinds of materials (i.e., rocks, wood, concrete, steel), either parallel or perpendicular to the shore, to provide a barrier between sea and land that resists wave energy or to trap and accumulate sediment. Shore-hardening structures include:

Breakwaters: Structures placed onshore and offshore, generally constructed of hard materials, and designed to absorb wave energy before the waves reach the shoreline (Fig. 1a). These structures are necessary to protect harbor entrances but should have limited application for beach protection, because they disrupt lateral sand transport.

Groins: Structures that extend perpendicularly from the shoreline with the design purpose to trap and hold sand, thereby nourishing the beach compartments located between them. They cut the longshore transport of littoral drift.

Revetments and riprap: Shore-parallel, sloping features which break up or absorb the energy of waves but may let water and sediment pass through (Fig. 1b). Revetments are made of concrete or shaped blocks of stone laid on top of a layer of finer material. Riprap consists of layers of hardrock boulders with the largest, often weighing several tonnes, on the head. Riprap has the advantage of good permeability and looks more natural; however, these structures are in the sea-wall family with the same detrimental effects. Their permeability can contribute to piping which contributes to the wall's sinking, requiring new material to be added.

Bulkheads and seawalls: Shore-parallel walls designed to hold bluffs and banks by completely separating land from water. Bulkheads act as retaining walls, keeping the earth or sand behind them from crumbling or slumping. Bulkheads are sometimes called "sand retention systems." Seawalls are primarily used to resist wave action and fail for a variety of reasons including cutting off the sediment supply, steepening of the beach profile, and causing wave reflection and refraction that cause scour erosion in front of the wall, and increased erosion rates at the ends of the wall.

Gabions: These metal cages filled with rocks are stacked to form other structures (i.e., seawalls, revetments, gabions) and are one of the poorest choices for such construction on the coast.

Geotextiles: Permeable fabrics which can hold materials while water flows through. Geosynthetic tubes are large tubes consisting of a woven geotextile material filled with a slurry-mix (Fig. 1c). The mix usually includes dredged material (e.g., sand) from the nearby area but can also be a concrete mix



Coastal Erosion Management, Fig. 1 Examples of the coastal erosion management by the use of the protection approach. (a) Breakwaters in Colombia, (b) revetment in Italy, (c) geotextiles in Senegal, and (d) beach nourishment in Spain

or mortar. These tubes are placed in the same manner as seawalls or groins but may be buried as well.

Most of the time, hard defenses are employed because they give immediate and tangible protection, and also they generate a sensation of an apparent solution to the protected populations (*The placebo effect* according to Rangel-Buitrago et al. 2018). The long history of the use of such shore-hardening structures has proven that they are a poor choice for CEM in most cases, and there is an extensive literature on how and why such structures fail (see, for example, Paganelli et al. 2014 for a review of short comings). The greatest short coming is that such structures do little or nothing to eliminate the coastal erosion process. Instead, erosion is increased in front of the structures or is transferred to nearby areas because the sediment supply is blocked. The use of hard structures is based on the protection of economic benefits of a given coastal area. However, it is essential to highlight that economic benefits accruing from protection depend on the

values of the land to be protected. Some advantages of the use of hard protection structures include:

- Prevention of physical damage to property as a result of waves and flooding
- Prevention of loss of [economic] production and income
- Prevention of land loss through erosion
- Prevention of loss of infrastructure (mainly focused on recreational activities)

On the other hand, hard structures generate adverse effects, such as:

- Accelerated bottom erosion in front of the hard structures and downdrift scouring (beach loss)
- Disturbance of sediment supply and beach reduction (beach and habitat loss)
- Restricted public access

- Potential risks for bathers (e.g., generation of rip currents, boating hazards, bottom holes)
- Loss of aesthetic visual effects on the seaside landscape

Perhaps the principal problem with the use of hard structures is that once built, they have fixed the location of the coastline in its position at the time of construction. This is highly problematic because coasts are dynamic landforms which respond to several processes such as rising sea levels and wave climate. Management should allow the shoreline to migrate, both landward or seaward; once “held in place,” the system is in disequilibrium.

Soft Defenses

The term “soft” defenses implies designs that have a less severe impact than hard structures, and they usually consist of coastal features that have been created or reshaped using engineering. While hard protection attempts to stop natural forces such as wave energy and tides, soft defenses attempt to adapt the system and supplement natural processes. Soft techniques include:

Beach nourishment or artificial sand supply: The artificial addition of sediment of suitable quality to a beach area that has a sediment deficit (Fig. 1d). This may be achieved through a direct placement of sands on the beach, through pumping or trickle charging (placing sediments at a single point). Sediment placement can be on the emerged part of the foreshore (“beach nourishment”) or under the water line (“underwater nourishment”). Often construction of artificial dunes accompanies such beach nourishment projects. This approach is most common, but it is expensive, destructive of habitat, requires frequent maintenance, may reduce aesthetics, and can actually increase the rate of beach erosion.

Beach drainage: Beach drainage is achieved by a system of perforated pipes under the beach to pump out water, lowering the water table. This allows water to percolate into the beach during backwash, reducing the seaward movement of sediment. The few documented examples were generally not successful.

Beach scraping: Artificial reprofiling of the beach when sediment losses are not severe enough to warrant the importation of large volumes of sediment. Reprofiling is achieved using existing beach sediment. This is a very short-term approach to combating erosion as the sand is lost in the next big storm.

Sand by-passing: Reactivation of sediment transport processes by pumping sediments that accumulated up-drift of coastal infrastructure and injecting them down-drift to maintain beaches. Although successful, this approach is rarely adopted because of the high costs and needed maintenance of the pumping station system. A variant of sand

by-passing is to place sediment dredged from navigational channels onto adjacent beaches to reactivate sediment supply and transport.

Construction of stable bays: By increasing the length of the coastline, wave energy per unit length of the coast is diluted. While some coastline segments are protected, erosion continues between these hard points leading to the formation of embayments.

Cliff drainage: In order to reduce cliff failure by slumping and land sliding, the pore pressure in the rock or sediment is reduced by piping water out of the cliff or bluff, therefore preventing accumulation of water at weak rock boundaries.

Cliff profiling: Slumping and landsliding of cliff and bluff faces can be reduced by lowering the slope angle to increase cliff stability. The angle at which a cliff becomes stable is a function of rock type, geologic structure, and water content.

Rock pinning: Prevention of slippage in seaward dipping rocks can be achieved by bolting layers together to increase cohesion and stability. This does not prevent wave attack at the cliff base but reduces the threat of mass movement and thus reduces net erosion rates.

All of these soft protection techniques need ongoing and constant monitoring, maintenance, and engineering (Edge et al. 2003). Hence, high costs and capability needs should be considered when such approaches are considered. The implementation of soft defenses represents a significant shift from an action-reaction basis (erosion problem = structure) to the adoption of a more proactive approach.

Land Reclamation/Land Claim

The primary goal of this approach is simple: create new land from areas that were previously below high-tide level. This approach is considered an “aggressive” way to protect the coast to such an extent that it also may be called “advance the line,” “attack the line,” “land reclamation,” or “reclamation fill.” Land fill and reclamation is practiced with the sole objective to gain land area to be used for economic purposes, and its use is particularly prevalent in large urbanized coastal areas where land values are elevated (i.e., Hong Kong).

Most of the time, the land claim approach is employed in deltas or estuaries due to the shelter afforded to economic developments, such as ports, as well as the availability of large areas of flat land, that is accessible from both sea and land (French 1997). Generally, this approach is completed with the enclosing and protection of shore or nearshore areas (e.g., diking/polders), as well as sand filling practices, often using the same techniques used in beach nourishment.

The main advantage of the use of land claim approach is the gain of additional coastal land that can be used for diverse

purposes (i.e., urbanization) and its implementation requires the use of mature soil that is optimal for agricultural production. This approach provides an attractive and valuable source of land in places where increases in the coastal population are projected.

Although the direct gain of land is beneficial, the land claim can generate negative impacts. This approach requires either the enclosure of intertidal habitats by hard defenses or their elevation to be increased above sea level to prevent inundations, thus producing a direct loss of coastal ecosystems. In the same way, hard defenses used to claim land causes erosion and scour of the coast and can hinder and even prevent natural ecosystem adjustments in response to changing factors such as sea level rise. Another disadvantage is related to acidification and pollution of the coastal zone. Acidification is linked to the action of bacteria in new sediments which create sulfuric acid when exposed to air, while pollutants can be introduced through the use of dredged sediments for raising land elevation. The above can be a big issue if the claimed land is to be used for agriculture or when coastal waters are essential for development of fishing activities.

Surge Barriers and Closure Dams

The constructions of structures to prevent extreme water levels penetrating a specific area are large-scale coastal protection projects able to defend rivers, estuaries, and tidal inlets. Closure dams are fixed structures that permanently close off a delta or estuary while surge barriers are gates or movable/fixed obstacles that can be closed to prevent flooding when an extreme water level is expected. Both cases are commonly applied at narrow tidal inlets, where the length of the structure is not required to be so high, and where defenses behind the barrier can be reduced both in height or width.

Closure dams and surge barriers may be beneficial for economic purposes. After construction, both structures can be used to develop trade activities. In the same way, they can provide a flexible approach to maintain most of the natural dynamics while still offering reliable flood protection. In particular cases, these structures allow the additional benefit of enhancing a system's natural capacity to clean itself. This can be achieved by independently opening and closing selected barriers, depending on water movement. Closing barriers improve the ability to drive water out of the system, therefore increasing its mobility and dispersing pollutants.

The main disadvantage of implementation lies in its high costs. To implement such protection, high economic investment is required during the construction, and continuous maintenance needs are high. In the same way, surge barriers and closure dams can generate floods on the landward side of the barrier when water levels are high and, in the case of

movable barriers, if the defense remains closed for an extended period. When a movable barrier is used an extra investment in flood warning systems is required to provide information about closure times. This approach can change the chemical, physical, and biological properties of the coastal system altering the inflow and outflow of water. This may generate alterations to water salinity, temperature, suspended matter, and nutrients affecting ecosystems.

Accommodation

Accommodation includes all approaches for addressing coastal erosion by revising and reorganizing human activities in the coastal zone, or even at the river basin level. This approach allows acceptance of a certain degree of flooding and erosion by construction methods, changing land use, and improving preparedness. The above means a continued usage of land at risk without attempting to prevent the area from being damaged by natural events, and allowing conservation and migration of ecosystems (Fig. 2).

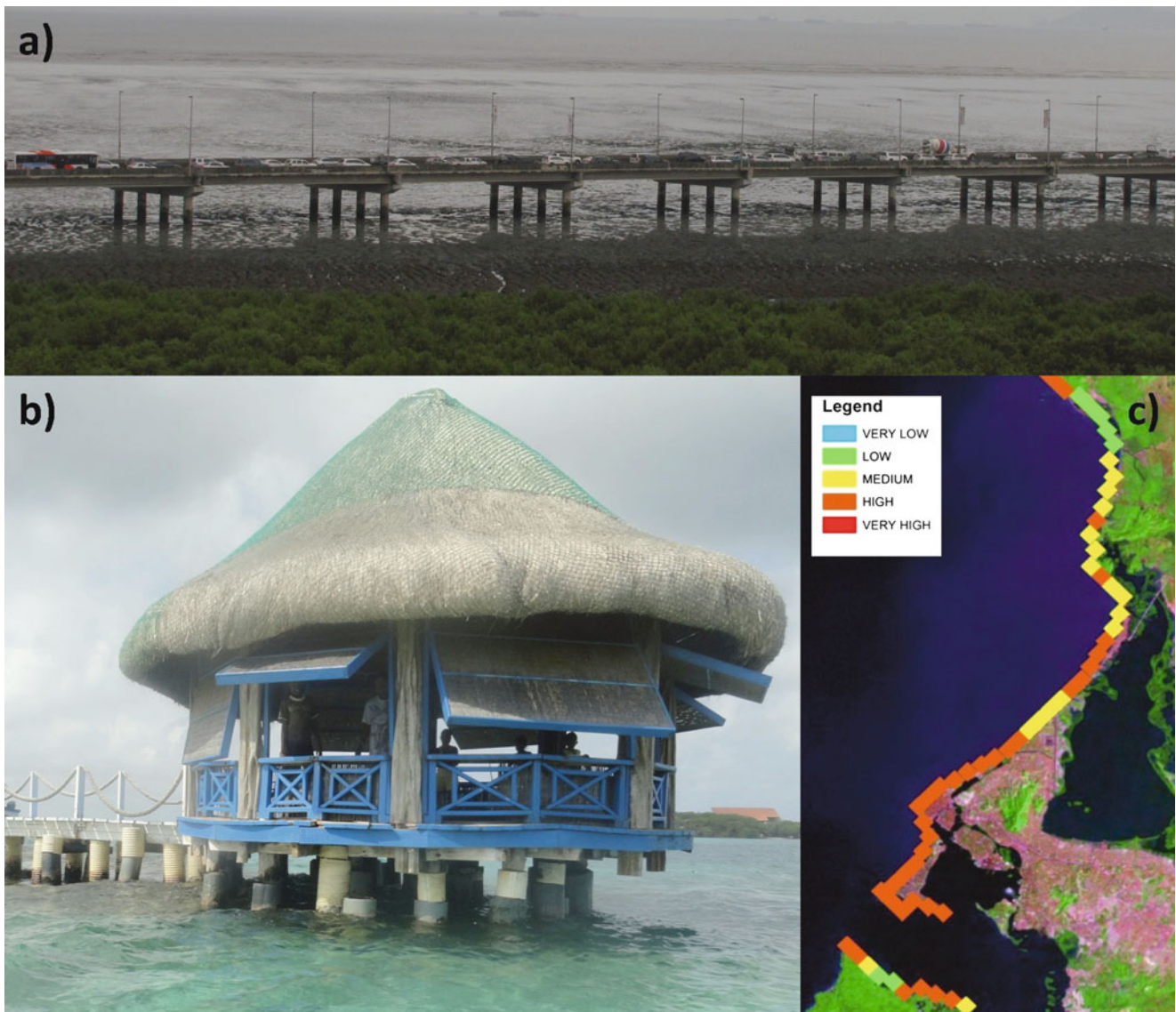
In general terms, the accommodation approaches can be divided into two fronts:

- Technologies that allow physical changes to accommodate increased coastal erosion such as, flood proofing, restoration, and floating agricultural systems.
- Information systems to enhance understanding of coastal erosion risks in order to develop appropriate responses to eliminate or at least, minimize impacts. These include early warning systems, indicators, and mapping/zoning.

Accommodation provides opportunities for inundated land to be used for new purposes and some compensatory economic benefits could be derived from this action. However, considerable costs may be involved in the planning and restructuring of land uses. The necessary expenditure may place significant stress on national budgets, especially in developing countries.

The social and cultural implications of the implementation of the accommodation approach can be significant. A shift in the economic activity of an area can change lifestyles of the coastal inhabitants. Also, this approach may lead to living conditions being less desirable, for example, if properties are subject to occasional flooding, or if problems with sewage disposal occur. Public safety and health may thus be adversely affected by this option.

By answering large-scale coastal erosion challenges through accommodation, the full range of coastal erosion collateral effects can be addressed within a longer-term perspective. This may also reduce the need for specific direct coastal protection measures. However, accommodation must be implemented proactively as it requires advanced planning and the acceptance that some coastal zone values could be lost (IPCC CZMS 1990).



Coastal Erosion Management, Fig. 2 Examples of the coastal erosion management by the use of the accommodation scheme. (a and b) Road and building raised (building code application) in Panama.

(c) Example of mapping/zoning in Colombia. The very low to very high designations are for coastal erosion potential

Building Codes

Building codes for the most part address mitigating the impacts on buildings by wind and flood waters, but a few of the regulations in coastal building codes are related to addressing erosion (Fig. 2a, b). Building code requirements specific to managing erosion (FEMA 2008) include:

- Adequate elevation of the building above the expected flood level plus the addition of wave height (i.e., building raised on stilts so flood water and erosive storm waves and/or surge can pass under the structure without inducing scour).
- Keep ground-level under building open or use breakaway walls to allow passage of flood waters without causing significant structural damage or erosional scour.

- Adequately anchor structures against flood flows so they are not swept away to become battering rams against other structures.
- Locate utilities and septic systems below ground at depths not likely to be reached if erosional scour does occur.

Floating Agricultural Systems

Floating agricultural systems are an adaptation approach to cope with increased or prolonged flooding and erosion. In a more specific way, such systems utilize waterlogged areas over long time periods to produce new soil, food, and at the same time avoid damage from flooding and erosion.

This approach uses a substrate of rotting vegetation that acts as compost for soil creation and crop growth in

waterlogged areas. The vegetation floats on the surface of the water, thus creating new spaces for agricultural purposes. Floating agricultural systems mitigate flooding erosion because plant roots create new soil and alter soil properties (i.e., aggregate stability, improve hydraulic function, and shear strength). In this way, cultivatable area can be increased, and communities can become more self-sufficient and at the same time control coastal erosion.

Floating agricultural systems are a relatively labor-intensive approach, so they can provide employment opportunities within communities. Such systems may also aid in food security by increasing the land output while supporting poor and landless people. The main disadvantages of this approach are there are very limited areas where such physical systems can be developed (e.g., shoal areas with low wave and current energy), and such systems are limited by climate conditions.

Forecasting and Early Warning Systems

Long-term weather forecasting provides some guidance in focusing on where coastal erosion events may occur in the future. This allows both managers and the public to be warned so that optimal decisions can be taken to reduce the adverse effects. As such, the primary goal of forecasting and early warning systems is to reduce the exposure.

Forecasting systems help anticipate potential events before they occur. They may determine the hazard and risk. The forecasts increase lead-time for the people to prepare and evacuate and help provide a short time for minimal responses to minimize potential erosion damage. The lead-time can range from hours up to a few days, depending on local hydrological and topographical conditions (susceptibility). The forecasting systems estimate expected erosion using data inputs from simulation tools and models (e.g., storm surge models).

In contrast to forecasting systems, the goal of early warning systems is to issue warnings when erosion is imminent or already occurring. Early warning systems need interrelated elements:

- Assessments and knowledge of coastal erosion-related process in the area
- Local hazard monitoring (forecasts) and warning service
- Erosion risk dissemination and communication service
- Community response capabilities

Forecasting and early warning systems go together. Once erosion is forecasted, early warning systems provide the crucial step to communicate the threat. Due to the current climate change scenarios, both are of vital importance and help to manage and minimize the catastrophic effects of coastal erosion. It is important to highlight that forecasting and early warning systems are not a stand-alone response to

minimization of the impacts of coastal erosion. Both must be coupled with emergency planning measures and should also contain an awareness raising element.

As stressed above, forecasting and early warning systems are not sufficient on their own to reduce coastal erosion risk. People's reactions to warnings – their attitude and the nature of their response – have a significant bearing upon the effectiveness of this approach. These systems are only useful when all inhabitants of a coastal area know what the forecasting and system of warning means, what the stages of warning are, and what to do when the warnings are given (Tompkins 2005). The accuracy is a vital point in the implementation of this approach. System inaccuracies might lead to complacency if previous forecasting and warnings were unfounded. Also worthy of note, in the USA, weather reports often state “beach erosion is expected” which usually means the back beach area will be eroded (i.e., bluff, dunes, buildings' footings) and storm surge and overwash are likely (the beach will still be there after the storm – the houses, maybe not).

Natural Indicators

The term indicator can be defined as the ability to measure and describe variables or as a parameter that indicates some characteristic or metric (Carapuço et al. 2016). An “indicator” may be interpreted as a quantitative/qualitative statement or measured/observed parameter that can be used to describe an existing situation and measure changes and trends over time. In the coastal erosion context, an indicator is a selected group of descriptors that express the predisposition of a coastal segment to be affected by coastal erosion, including evidence of past erosional events. In the coastal erosion management context, there are three types of indicators:

Geoindicators: Describe the physical/morphological features of coastal systems and can be grouped into geology (expressing processes and materials – their relative resistance to erosion) and landforms (expressing the coastal type).

Driver-based indicators: Used to assess coastal erosion vulnerability, namely, in large-scale applications involving semi-quantitative coastal indexes (e.g., scaled in classes from 1 to 5 or from very low to very high). These indicators include the parameters driving coastal erosion (i.e., wave characteristics, sea level, wind).

Process-based indicators: These reflect the interaction between coastal erosion driving mechanisms and the coastal morphology. These indicators are obtained through formulations or models that combine the driving variables (e.g., wave height, sea level) and the morphological parameters (e.g., cliff height, lithology). The resulting indicator has a physical meaning, directly related to the coastal erosion hazard.

Indicators are used to evaluate erosion susceptibility of coastal areas through a relatively fast and low-cost approach. And indicators are of practical use in coastal zone planning and management, because they allow a different kind of monitoring in coastal erosion management. However, indicators are not sufficient on their own to reduce coastal erosion but are tools and the selection of the appropriate indicator and associated thresholds is fundamental to obtain optimal coastal erosion hazard estimation.

Mapping/Zoning

Ideally, mapping should be the first step in CEM, because this is the broadest approach to define those coastal areas which are at risk of erosion under specific conditions (Fig. 2c). Coastal zone mapping usually consists of a range of map types from traditional (e.g., topography, land forms) to specific hazards (e.g., FEMA flood maps, storm surge maps, erosion rate maps). Computer modelling combined with other map data bases allows construction of hazard-specific maps (e.g., Puerto Rico's Tsunami Zone maps). With computers, combining various types of coastal hazard maps provide for hybrid maps to express general coastal vulnerability, usually expressed in terms of a graded coastal vulnerability index (CVI). Coastal erosion is always an aspect of this type of mapping, and all of these map types can be applied to CEM. FEMA flood maps have designated "wave velocity zones" which can be taken as erosion zones, especially for coasts of erodible material. Storm surge maps similarly may indicate the reach of erosive waves and overwash/backwash currents. Projected sea-level rise inundation maps show where new vulnerable shorelines will be. In the Pacific, modelling of El Niño flood areas can be mapped. And old maps and air photos provide data bases for mapping historic erosion rates and changes in those rates. The applied goal of all of these maps is risk reduction, but they also support CEM in working to reduce the impact of coastal erosion. Mapping shows the expected extent of the erosion process in a given location, based on various scenarios.

Mapping is usually the basis for zoning, the common management approach for appropriate land use planning, especially in erosion-prone areas. Coastal zoning is the division of areas into zones that can be assigned different purposes and user restrictions, based on the knowledge of the coastal erosion hazard and other restrictions. Coastal zoning allows multiple users to benefit from a coastal area under a broader sustainable management strategy. Coastal zoning schemes can constitute the regulatory and planning framework for CEM and other coastal management issues. As noted by Bapulu and Sinha (2005), mapping creates easily read, rapidly accessible data sets that facilitate the identification and zonation of areas at risk of erosion and also help prioritize mitigation and response efforts.

One of the drawbacks of past mapping is that the conclusions for some hazards, such as coastal erosion, flood levels, and storm surge levels, are based on past records (historic) or models which also use past information (e.g., wave data, storm frequency) and do not have mechanisms for including the changes that are known to be occurring (e.g., sea-level rise, subsidence rates, impacts of land-use changes). Some newer mapping techniques are now including these variables and generating maps that are forward looking, rather than expecting conditions to remain as they were in the past (e.g., AMBUR, Jackson et al. 2012).

Climate change must be a primary consideration in future mapping. The effects of climate change are far more reaching than just the sea-level rise, and many of the coming changes will create new interactions with the dynamic nature of the coast. For example, changes in extreme waves, occurring as a result of climate change, will generate changes in the areas susceptible to erosion. So traditional maps will require periodic updates to reflect the changing risk of erosion and also to update the existing zonation. Maps should have many final users including developers to determine areas at risk of erosion and by insurers to assess premiums in areas where property is at risk. Maps must be integrated with other procedures, such as emergency response planning and town planning before the full benefits can be realized.

Use of Ecosystems

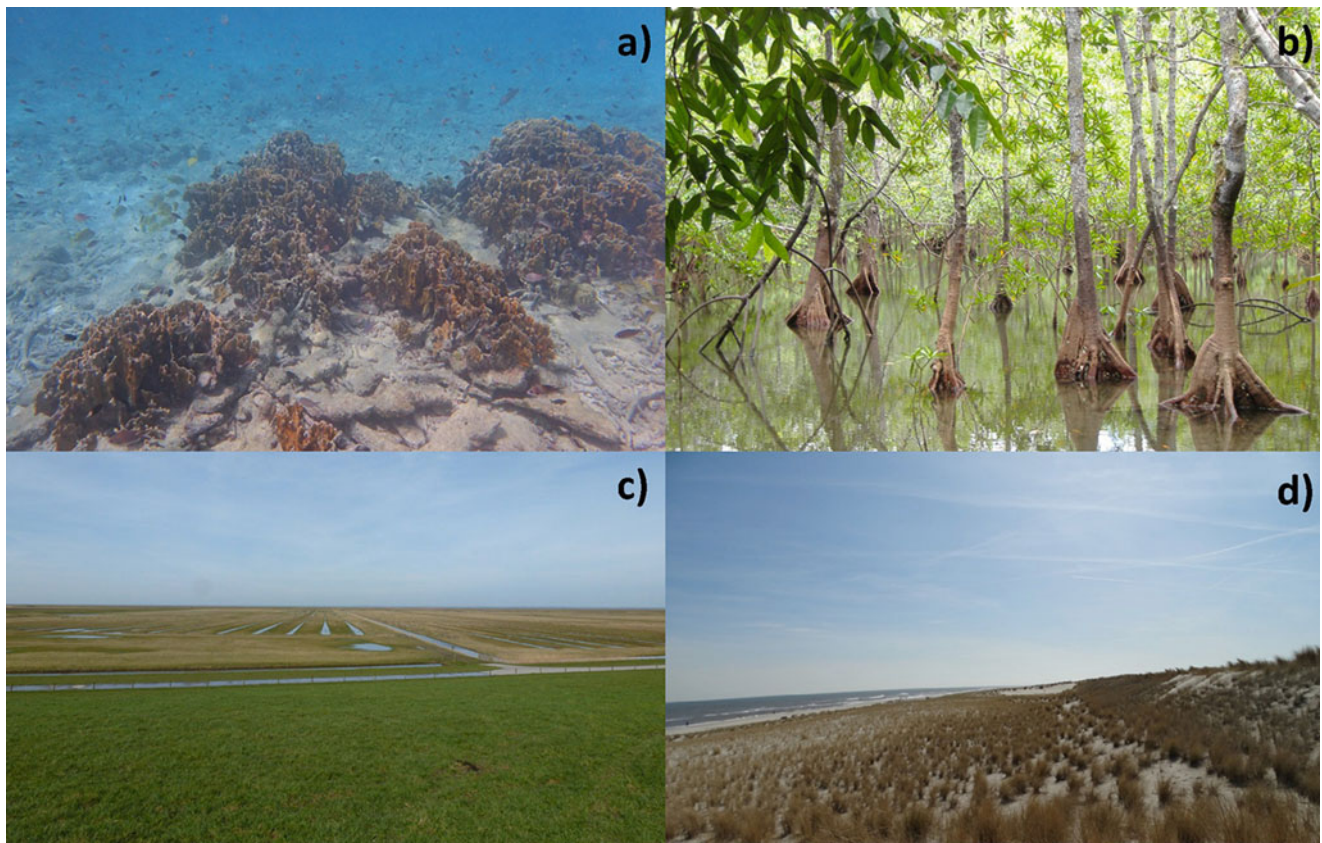
In recent years, an ecosystem-based CEM strategy has been brought into practice as an approach that is more cost-effective, and sustainable, than conventional schemes. This approach is based on the natural environmental influences over processes related to coastal erosion (e.g., sediment capture and energy attenuation) by restoring or creating ecosystems (Fig. 3).

The use of ecosystems for coastal erosion management is an integrated process to conserve and improve ecosystem health that sustains ecosystem services for human well-being. This approach combines two particular actions:

- Develop or maintain ecosystems quality
- Ensure the best way for delivery of different ecosystem services to human well-being

CEM based on habitat diversification can be applied worldwide, particularly in areas that have sufficient space between existing urbanization and the coastline to allow for ecosystem development (for example, in accommodation strategies). Because ecosystems have the natural ability to reduce extreme wave effects (Shepard et al. 2011), their growth can keep pace with sea-level rise through sediment accretion if available (Kirwan et al. 2010).

The use of this approach gives multiple benefits in comparison with common conventional methods of shore hardening.



Coastal Erosion Management, Fig. 3 Examples of coastal erosion management by the use of ecosystems. (a) Coral reefs in Bonaire, (b) mangroves in Ecuador, (c) salt marshes in Holland, and (d) dune vegetation in Holland

These benefits include natural habitat conservation, the creation of recreational spaces, carbon sequestration, improvement of water quality, production of fisheries, and wave attenuation. In fact, many existing coastal ecosystems already provide some degree of protection with no installation cost.

Despite the many advantages, it is essential to highlight limitations in the implementation of ecosystem-based coastal erosion management. Ecosystems demand space to flourish and sometimes require more space than conventional protection approaches. Flexibility in planning is required as ecosystems usually have a recovery capacity after losses (e.g., storm damage), but this recovery is not immediate. In some cases, recovery can be prolonged and can even become lost due to external factors. In the same way, ecosystem development and functionality will depend on the coastal setting, hydrodynamics, human factors, and habitat dimensions, together with the severity of coastal erosion. Some common protective ecosystems with respect to CEM are as follows.

Coral Reefs

These ecosystems have many functions in the coastal erosion management process, but the most important are

sediment generation and energy dissipation (Fig. 3a). The organisms that make up reefs have skeletons composed of calcium carbonate which allows them to build the rigid reef framework and also to produce sediment when broken up by wave activity or bioerosion. So reefs serve a double function in building the wave-resistant ecosystem and generating sediments. Fragmentation and erosion processes (including bioerosion) transform reef structures into biogenic, calcareous sands, able to feed the coast, and maintain its stability. Reef organisms can also contribute to sedimentation through allogenic processes related to mucous secretions. Coral secretions trap suspended particles in the water column, forming aggregates that sink rapidly to the bottom. Secretions by corals can make it the dominant form of suspended organic matter within and around coral reefs (Marshall 1968).

Their shape and structure allow coral reefs to act as barriers that dissipate wave energy, providing a natural submerged breakwater. Their geometry (porosity, surface, tortuosity, roughness, and the overall void matrix), water depth above the reef system (depth of flow), and length in the direction of wave propagation are vital points in the wave energy dissipation process (Gracia et al. 2018).

Corals collaterally generate environmental and socioeconomic benefits. These ecosystems create high-value biodiversity hotspots that also filter waters, positively affecting overall water quality. From a socioeconomic point of view, reefs provide an essential source of food and income for local livelihoods (i.e., fishing and tourist activities) and contain marine organisms with chemical compounds that have a high potential for medical use.

The main disadvantage lies in that protection activities can be complex technologically and politically – mainly when stress sources are found, and significant economic development activities demand to be foregone in order to protect the reefs. Unfortunately, global warming and acidification of sea water are drivers of reef degradation which cannot be managed and eliminated on a local level.

Wetlands (Mangroves and Salt Marshes)

These are land areas saturated with water, seasonally or permanently, such that they take on the characteristics of a distinct ecosystem (Fig. 3b, c). The prime factor that distinguishes wetlands is the characteristic vegetation of aquatic plants adapted to the unique hydric soil conditions. The major coastal wetlands, mangroves, and salt marshes, are termed “the genuine ecosystem engineers” because of the conjunction of their intrinsic properties (width, structure, tree size), their link with other ecosystems, their function as marine nursery grounds, and their ability to contribute to sediment production and reduction of coastal erosion.

Wetlands can significantly reduce the energy of any fluid that is moving through them. The energy power lost when wind and waves pass through roots and branches can range from 15% to 65%, minimizing seabed scour and further sediment movement (Spalding et al. 2014). At the same time, their structure can reduce winds across the surface of the water, reducing the reformation or propagation of waves. In the 2004 Indian Ocean tsunami, some areas fronted by mangrove stands suffered less damage than unvegetated shores. Vegetation structure is essential in determining possible protection during extreme erosive wave events and tsunamis. Of course, this protection capacity also depends on the size and forward speed of the storm, tsunami characteristics, and coast setting. However, it has been determined that wetlands are more efficient at reducing surge levels if the event passes over relatively quickly.

Mangrove and marsh-grass roots help generate and bind new sediments and soils. The above-ground roots slow down the water-flow process, stimulate sediment deposition, and reduce erosion. Sedimentation rates correlate with root density and between 70% and 80% of suspended sediment brought in from coastal waters may be trapped in wetlands (Furukawa et al. 1997).

Wetlands also provide habitat for a diverse range of plant and animal species (breeding and nurseries for fish, birds, shellfish, and mammals) and can release organic matter to the ocean (e.g., mangroves) providing support for marine life. From an economic point of view, they often serve as an essential source of food and resources for livelihoods (e.g., wood, fishing, tourism) for local communities.

Seagrass Meadows

Seagrasses can significantly influence the hydrodynamic environment by reducing flow velocity, dissipating wave energy, and stabilizing sediment. When a wave reaches the ecosystem, a negligible energy reflection and weak wave attenuation by friction are produced. This situation is opposite of what happens with most hard structures, where the same process gives rise to higher energy reflection and wave attenuation by breaking and friction with further loss of sediment. Additionally, meadows can stabilize and maintain sediments in shallow areas. Due to their ability to dampen waves and currents, seagrass canopies can increase sediment deposition, decrease resuspension, and even directly intercept suspended sediment. Also, substances secreted by epiphytes can bind sediment particles to seagrass leaves. In the same way, they can influence the original seafloor bathymetry through the accretion of rhizomes and roots in the sediments, thus exerting new forces over hydrodynamics and sedimentation.

The efficiency of this ecosystem is primarily and firmly based on the density, standing biomass, plant stiffness and incident energy flux (Gracia et al. 2018). Optimal conditions for enhancing coastal erosion defense provided by seagrasses can be reached at shallow waters and low wave energy environments. Like reefs, seagrass ecosystems can create new high-value biodiversity hotspots and also can filter waters, positively affecting overall water quality. Seagrass meadows support commercial fisheries (offering nursery functions for commercial species).

Artificial plantings of mangroves and seagrasses to reestablish existing ecosystems or encourage new growth should be a consideration for any CEM plan. Removal of either of these systems usually leads to an increase in the shoreline rate of erosion. A primary problem in implementing their use is that some coastlines have high levels of recreational or development activities that disturb these ecosystems.

Shellfish Beds, Banks, and Hardgrounds

A shellfish community or single shellfish species can create shell banks or beds on the sea floor that produce sediment as well as offering a wave-energy dissipating effect. Like reef skeletal material, most shells are calcium carbonate and produce sediment when fragmented. Some, like serpulid worms (polychaetes), are encrusters and bind sediment or may build small reef-like structures. If organisms have a cement

binding, they produce a hardground that is resistant to erosion. These ecosystems are topographically rough, with fractal complexity capable of reducing wave energy and erosion (Commuto and Rusignuolo 2000). Their structures can act as barriers that generate dams, to hold pools of water, and increase immersion time above the shoreward bank margin, facilitating sediment deposition. Extensive shellfish banks and beds can minimize the impacts of direct water flow, extreme waves, storm surges, and can stabilize the shoreline. Under low-to-moderate flow action, increased deposition on beds causes sediment to build up to form banks that can be higher than the ambient substrate.

Shellfish beds or banks implementation has several collateral positive effects:

- Can provide habitat for other species by creating a hard substrate with high surface complexity, acting as attachment sites for sessile organisms and refuges for mobile microorganisms, supporting high levels of species diversity.
- Can accrete dead shell material such that the bank grows in size and mass over time (except where restricted by tidal exposure or when harvested) with decay occurring at varying rates.
- Can provide food for other organisms, either when consumed directly or through the species assemblages they support.
- Can filter the water and reduce turbidity by extracting phytoplankton and organic and inorganic particles from the water column.
- Carbonate shells accumulate carbon in the calcium carbonate of their shells that help reduce the concentration of greenhouse gasses.

Dune Vegetation

Natural dunes offer both a protective barrier during storms and an obvious sand supply for natural nourishment of adjacent beaches (Fig. 3d). Some management schemes include bulldozing sand dikes or ridges under the guise of “dunes,” but true dunes are a complex ecosystem that relies heavily on vegetation and a soil biota. Dune vegetation stimulates dune growth by trapping and stabilizing wind-moving sand. Small plants located on the face of eroded dunes can enhance the natural development above the limit of direct wind or wave attack. These coastal ecosystems need conservation in any CEM plan. Dunes are extremely sensitive to foot traffic and should be protected by such means as constructed walk overs, or limited access paths.

Dune vegetation can be transplanted to encourage the growth of new foredunes along the toe of existing dunes, as long as these species are tolerant to occasional seawater inundation. Planting vegetation from seed can be undertaken but will not usually be successful in the dynamic foredune environment. Transplanted

grasses also need fertilization to encourage the associated soil biota (e.g., Vesicular-Arbuscular Mycorrhizal Fungi).

Vegetation sowing and transplanting per se will not construct new dunes or completely prevent erosion. Plants encourage natural recovery by creating a reservoir of sand within the dunes that better enable their ability to withstand erosion. Fencing, thatching, and beach recycling are often necessary to help in sand accretion. Additionally, these works can provide extra protection from waves and will reduce damage due to trampling. Once vegetation is well established, dunes may become self-sustaining, although any erosion damage will need to be quickly repaired.

The use of dune vegetation can decrease adverse effects on landscapes (even artificial dunes). In the same way, developing a dune system can restore a degree of the natural character to places that once had natural dune complexes before development and provide a valuable coastal habitat for animals. From an economic point of view, dunes help to reach multiple management objectives, such as public access to recreational resources and hazard mitigation.

Transplanting and management of appropriate dune vegetation will have no damaging impact on the natural environment of the receiving area but can be harmful to the borrow area, particularly when sand is bulldozed off of the beach. Over-harvesting of transplants from any region can give rise to increased erosion in specific points. This may be most significant for salt-tolerant sand-couch vegetation, as the borrow area will necessarily be a foredune susceptible to wave over-washing and wind erosion. Also, the use of dune vegetation in coastal erosion management typically requires dune fencing or thatching to achieve success. The construction of fences or thatching will disrupt public use of the coast, so provision must be made for controlled access.

Managed/Planned Retreat

Managed/planned retreat has the primary objective of reducing population, property, and infrastructure at risk through the planned withdrawal from coastal hazard zones, including erosion-prone areas (see “► [Managed Retreat](#)” in this volume). Specific to the erosion hazard, managed retreat implements mitigation tools designed to move existing and proposed development out of the path of both short- and long-term coastal erosion (Fig. 4). The approach may include managed realignment where shore-protection structures are removed selectively to allow natural coastal environments to be reestablished. This coastal erosion management approach is based on a simple philosophy: get out, avoid damage, and recognize that the coastal zone dynamics should dictate the type of management employed.

This approach works well in large areas where retreat strategies allow coastal ecosystems (e.g., wetlands) to adjust to increased levels of the sea through a slow landward migration. On the other hand, in small areas (e.g., islands), retreat

Coastal Erosion Management,
Fig. 4 Example of a new habitat construction (dunes) after moving the existing line of defense in Holland



does not offer a suitable alternative. There may be little or no land for resettlement and also many ecosystems, heritage, and cultural assets that will be affected or could be lost entirely.

Managed/planned retreat raises significant transboundary implications. The institutional capability to conduct a retreat on a temporary or permanent basis must be established. Currently, managed/planned retreat is supported under practices and legal tools that allow moving property and people out of coastal zones affected by erosion.

Management Realignment

Managed realignment is the process of moving back existing defenses in a coastline allowing the reestablishment or creation of a new intertidal habitat area through its natural regenerative capacity or reclaimed land (Fig. 4). This intertidal habitat will be highly effective at attenuating wave energy and also will reduce offshore sediment transport and therefore erosion. Intertidal habitats also can form dense root mats which increase the stability of intertidal sediments, helping to mitigate coastal erosion (USACE 1989).

This approach involves well thought out plans for the retreat of communities and ecosystems with a landward shift of the ecosystems. It is appropriate in areas where low-cost agricultural land holds sway, where there are few structures or directly affected stakeholders, and costs of compensating owners are minimized.

Managed realignment can be implemented following four methodologies:

Removal: Defenses are removed allowing the coastline to respond more naturally to hydrodynamic process.

Breaching: Selected existing defenses are removed to allow tides and waves into the previously protected land.

Realignment: Change in the position of the line of defense to favor the development of a more dynamic coastline.

Controlled tidal restoration: Tidal flow into the embanked area is controlled while defenses are maintained.

The main disadvantage of the managed realignment is the loss of private properties and commercial income. This approach is not politically and economically viable in urban areas or where agricultural land is of high quality.

Avoidance

Avoidance is ensuring that new infrastructure and assets are not permitted in areas likely to be affected by coastal erosion, without substantial planning controls being implemented. Avoidance may not seem like a managed retreat, but defining areas where no development is allowed because of coastal erosion, strategic ecosystems, or limited sediment sources should be part of any managed retreat plan. This approach can be developed in four specific ways:

Setbacks – hold the line: A coastal setback can be defined as the distance to keep structures out of coastal erosion or at least at a reasonable distance from a hazardous process. The idea of a coastal setback is simple: to allow room for the high water mark to naturally move inland throughout all the service life of the property. This approach enables erosion to continue along specific sections of the coast, while further development is not authorized or at least restricted. Setbacks can be divided into two types:

- **Fixed:** Limits growth for a set distance landward from a specific reference feature.
- **Rolling:** The regulatory line shifts landward according to topography or shoreline movement (dynamic or natural processes are used to lay out the setback line).

The adoption of arbitrary distances which do not indeed represent the threat from erosion can generate problems; in that sense, setbacks should be established based on real information (i.e., shoreline evolution or extreme water levels).

Rolling easements: This legally enforceable expectation is that the shore or human access along the shore can migrate inland instead of being squeezed between eroding coasts and a fixed property line or physical structure (Titus 2011). In a more specific way, it is a legal instrument that allows publically owned land, and land use restrictions, to migrate inland as shorelines retreat.

Rolling easements, like coastal setbacks, are measured against a coastal benchmark (e.g., high water line, dune crest or vegetated area). Because these benchmarks are expected to move dynamically with changing coastal conditions, the easement is said to “roll.”

A rolling easement can work in two ways:

- As a regulation that prohibits shore protection.
- As a property right to ensure that an area moves inland with the natural retreat of the shore.

Although the regulatory approach is the more common way to prevent shore protection, the nonregulatory approach may sometimes work better. Private land trusts, government agencies, and (for some plans) even private citizens can buy (or secure donations of) rolling easements from property owners. An owner who has voluntarily engaged in the creation of the rolling easement is more likely to perceive the arrangement as fair than a landowner subjected to government regulation.

No-build/No dwelling zones: These are areas, defined under laws and policies, that are not recommended, or are forbidden for human habitation and infrastructure development because of a high degree of coastal erosion susceptibility. The basis of this approach lies in the conservation of an eroding shoreline which is more than just a buffer zone and usually is a state policy to uphold the people’s constitutional right to life and property.

Construction disincentives: Legal strategies to encourage people not to build in high-risk zones or areas of critical habitat. This approach is based on the premise that construction over regions with high erosion rates cannot be covered for insurance or any federal assistance (i.e., small business loans and funding to rebuild infrastructure).

Acquisition

Land acquisition is the process wherein coastal lands are acquired by the government to protect them from future

coastal erosion related problems. This acquisition can be made through negotiated agreements where owners voluntarily sell the property, or, in extreme cases using compulsory purchase.

In some nations, the acquisition approach is developed through a legal figure called “coastal public reserves.” It consists in the designation of a strip of land measured from the high water mark and running along the shoreline as land owned by the state. This approach provides benefits concerning conservation, providing public access to the shore, contributing to recreational and tourism needs, preserving aesthetics, and protecting habitat.

Relocation/Resettlement

This is the process of one person or a community leaving an area affected by coastal erosion. The approach can be developed in three different ways:

Active: Undertaken by moving a building back either before it is threatened, or, if threatened, before it is damaged.

Passive: Rebuilding a destroyed structure in another area, away from the shore, and out of the coastal hazard zone.

Long-term: Implies a broader strategy through community zoning or land use plans that identify a frontal zone of buildings likely to be affected by known erosion rates.

Although relocation may represent the most effective long-term adaptation strategy for some coastal communities, this option is still considered outside the range of acceptable possibilities due to political, institutional, sociocultural, and economic considerations. Also, the “where” and “when” to relocate has not always been clear or adequately enforced.

Sacrifice/Abandonment

The sacrifice approach allows property loss when other suggested protection measures are not viable, or the accommodation and retreat option does not exist (Fig. 5). Sacrifice may be viewed as an “extreme” strategy, because it is frequently recommended after extreme-erosion events leave buildings battered and broken (Fig. 5). Yet abandonment is usually the safest and most economic, because it avoids reconstruction in the hazard zone and the repeated losses that follow. This planning strategy can be developed in four ways:

- As a “do nothing” approach in which buildings are regarded as having a fixed lifespan, and when their time comes to fall into the sea, no attempt is made to protect them.
- As an emergency program that encourages relocation of buildings before a hazard event to avoid economic loss.
- As a law that bans post-erosion reconstruction or by requiring relocation landward of the revised post-erosion setback control line.



Coastal Erosion Management, Fig. 5 Examples of sacrifice/abandonment approach in (a) Colombia and (b) Brazil

- As denial of insurance and subsidy programs in order to discourage rebuilding actions after erosion.

According to (Williams et al. 2018), the minimum needs to develop a sacrifice practice in a coastal area are:

- Identify the rationale behind the decision.
- Use demonstration sites to present the feasibility of accommodating retreat.
- Assess the geomorphological/ecological changes and indicate the advantages of allowing them to occur.

Summary

In the twenty-first century, the wise implementation of coastal erosion management is facing a challenging set of global changes: an increasing population in the coastal zone, the parallel increase in property development, the on-going sea-level rise, increasing the reach of erosional processes, potential increasing storminess (wave energy, storm, surge, rainfall runoff), and the unsustainability of traditional antierosion projects (e.g., increasing costs for hard and soft approaches, and decreasing sand supplies for nourishment). Past coastal erosion management has not been particularly successful, primarily because social-political systems are more driven by the profit motive than by utilizing the knowledge base

and tools to avoid and/or mitigate erosion. The time has arrived to do more than just implementing one or a combination of the previously presented approaches. Improving CEM requires policies and implementation processes involving more in-depth knowledge of the processes implicated in coastal erosion and optimal strategies and operative management frameworks. The basis should be based on historical and scientific knowledge in order to come to solutions that fit CEM frameworks and should be carried out following the best available techniques. Clearly, some traditional strategies must be abandoned and current underutilized approaches given more consideration in addressing erosion problems (e.g., use of ecosystems, managed retreat).

CEM is not about just solving a problem by the use of one of the five approaches presented in this entry but rather about future livability with the coastal environment. CEM must be the critical point, together with the ongoing process, that demands a constant identification of risks and opportunities, implementation of coastal erosion reduction measures, and especially the review of strategy effectiveness. The optimal performance of any strategy must be carefully monitored and assessed and results feedback through the entire cycle to improve use and future interventions.

Strategic planning for CEM relies on understanding the processes responsible for shaping coastal morphology. This process not only needs to include the local or regional procedures, but also the inclusion of the processes that lead to the formation (even the physical and chemical weathering) and the physical sediment supply that are of importance for the coast. Likewise, such strategies must be developed within policy frameworks that set clear objectives, and in an institutional environment where stakeholders have different defined roles.

CEM can be a slow process, so expected results are likely to be met in the medium- to long-term time frame. In this sense being proactive is the crucial issue. Flexible strategies instead of reactive measures should be adopted to strengthen CEM and thus improve the coastal environmental quality. A considerable level of cooperation is required from all stakeholders to take actions using CEM. This will help foster and close existing links between CEM, disaster risk reduction, and climate change adaptation, as well as between science and policy.

Cross-References

- [Beach Management Tools](#)
- [Coastal Erosion](#)
- [Coastline Changes](#)
- [Managed Retreat](#)
- [Modes and Patterns of Shoreline Change](#)
- [Natural Hazards](#)

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Coastal Evolution in Microtidal Seas in Holocene

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Introduction: Origin of Coasts and Their Main Evolution Factors

A general layout of the shoreline depends on land and bottom surface and the sea level. Both the Earth's crust and water volume undergo continuous and slow balance changes which affects the coastal zone landscape. Coast evolution rate depends on several factors as described by Cooper (1994), including geological, geomorphological and tectonic settings, paleotopography, climate, sea level changes, wave energy (wind), volume and direction of sediment supplied by river load, and intensity of coastal processes.

It would difficult to determine the impact of all the factors everywhere on the Earth. Therefore, the southern coast of the Baltic Sea was selected as a specific laboratory, where the magnitude and intensity of certain factors influencing the development of the coast can be limited. Such approach facilitates understanding of coast evolution processes.

An area extending from Scandinavia to the Baltic Sea basin was covered by an ice sheet in the past. The glacier had transported and then deposited moraine material (till, sand, xenoliths, boulders) its retreat creating a specific surface topography in the Southern Baltic area.

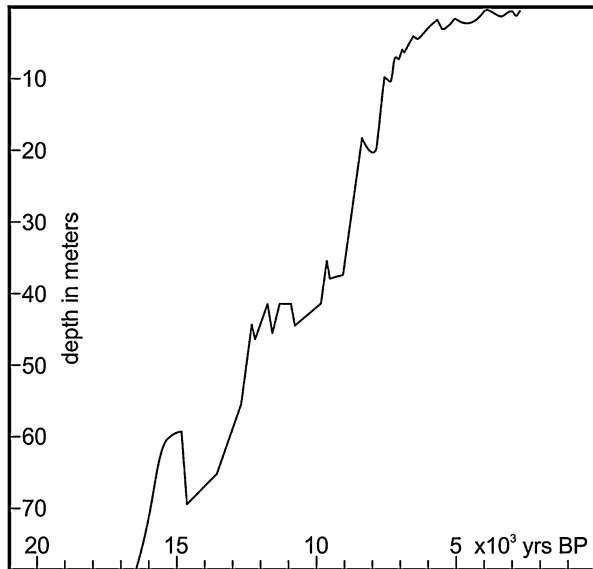
Moraine uplands of mainly undulating relief, that may be locally planar or hilly, are predominant. In turn, plain relief is attributable to remains of former ice-dammed lakes, bottoms of ice-marginal valleys that used to transport the meltwater and areas of marine accumulation plains.

The coastal evolution in the southern Baltic Sea in the Holocene was proceeding in the past and is proceeding now under the influence of: eustatic sea level change and glacio-isostatic adjustment (GiA). (Harff et al. 2017).

Ca. 8 ky BP sea level of the Baltic Sea was approximately 20 m lower than at present (Fig. 1). Where the coast and the land meet, the so-called coastal processes take place and modify the coast. The evolution rate of the coast depends mainly on its geological structure. The southern Baltic, coast is composed of sand and till originated during the Holocene deglaciation (sedimentation) in the temperate climate zone and in the nontidal sea environment.

Several factors and their intensities have been affecting the development of the Baltic Sea coast since the Holocene, among which the sea level rise and its fluctuations (Fig. 1) along with wave action have been playing key roles.

Sea level fluctuations trigger a displacement of the main zones of wave energy dissipation and the hydrodynamic zones related to them. Eustatic sea level fluctuations, which are associated with climatic changes, and neotectonics within the coast (uplift or subsidence of the basement), which induces the so-called “relative sea level fluctuations with growing trend,” are crucial in a particular timeframe (Harff et al. 2017).



Coastal Evolution in Microtidal Seas in Holocene, Fig. 1 Relative sea level changes of the southern Baltic in Late Pleistocene and Holocene. (Modified after Rotnicki 1995)

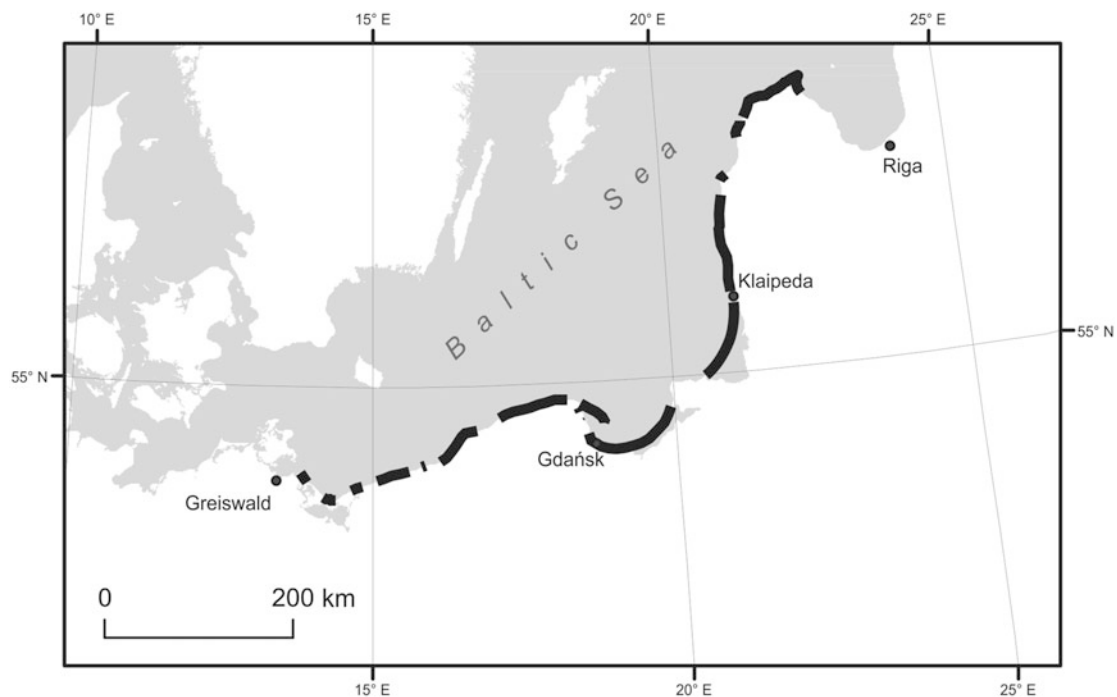
Within the southern Baltic, “pulsing and oscillating” sea level fluctuations occurred during the Holocene transgression. Apart from the continuous sea level rise, the transgression was slowed down a few times and even short-term regressions were demonstrated to have happened (Rotnicki 1995). These fluctuations took several hundred years and ranged within a few meters (Andrén et al. 2011; Uścińowicz 2014).

In the Late Holocene, during the Subboreal (6–5 ky BP) the marine transgression started to grow weaker (because the ice shield’s cooling effect changed) and the mean sea level rise rate dropped to ca. 2 mm/year. The southern Baltic shoreline got close to its a present configuration, while the sea level was 2.5–3 m lower than now. During the Subboreal and Subatlantic (i.e., within the last 5 ky), the shoreline was getting slowly closer to its present configuration.

The sea level rise rate has been a very important factor in shaping the coast profile. Long-term sea level fluctuations, i.e., the sea level rise and decline, result in shore erosion or accumulation, respectively.

Origin and Migration of the Southern Baltic Sea Barriers

Ca. 80% of the southern Baltic coast is composed of barrier and/or spit-type shores which include extensive dune systems a few up to 40 m high (Fig. 2). They developed on low-lying



Coastal Evolution in Microtidal Seas in Holocene, Fig. 2 Some peninsula, spits, and barriers of the southern Baltic Sea coast (Furmańczyk and Musielak 2015, supplemented)

sections of the coast, mainly in places where ice-marginal valleys and coastal lowlands reached the sea.

The dune-covered sections of the coast feature different dune morphologies. They are mostly represented by one or a few berms running parallel to the shore. These forms separate the sea from: ice-marginal valleys, wetlands, and coastal lakes. Dunes cover postglacial moraine deposits (sand and till) along some section of the coast.

As a result of long-term impact of the sea on the shore, its outline is generally levelled out and feature dominant barrier landforms that are specific for the fluctuating nature of the sea level rise in the Holocene (Furmańczyk and Musielak 2015).

During earlier phases of the southern Baltic Holocene transgression (8 ky BP), shore destruction was very intense, whereas during later phases (since 6 ky BP), when the shore erosion was much weaker, the transgressing sea water moved inland to shape the coastal zone landscape (Mörner 1980; Rosa 1984; Rotnicki 1995; Uścińowicz 2003). Storm-driven waves undercut the moraine uplands causing their retreat and leading to the formation of cliffs. Depending on wind direction and sea level changes, sandy sediments produced by erosion were supplying the beaches (in front of cliffs and on their both edges) and the submerged part of the coastal zone. Subsequently, those sediments were carried away along the shore or into deeper parts of the nearshore zone.

Simultaneously, waves, that were affecting gentle slopes of valleys, produced wide beaches and underwater bars parallel to the shore. As the sea level was rising, these were gradually moving inland. During periodical episodes of sea level decline the bars emerged above the sea surface, the shore featuring a wide beach. The beach became a source of wind-transported sediments building the dunes. At that time barrier dunes, the height of which depended on the volume of sand material available, were formed. The underwater bars, mentioned above emerged and formed barrier islands.

Ca. 4.5 ky BP, there was a longer period of evident sea level stabilization, and even a decline (Fig. 1). Throughout that period, aeolian processes were intense within the area of wide beaches and emergent bars to produce rows of sandy barriers in the form of dune berms parallel to the shoreline. The next sea level rise, that was not as rapid, resulted in a gradual migration of these barriers toward the land (Figs. 3 and 4). Similar phenomenon was observed on the Atlantic coast of US (Leatherman 1988).

Sea level fluctuations in the coastal zone are of different origin, amplitude, and duration, while their impact on the coastal zone differs also spatially. The sea level rise rate has an essential impact on a way the spit/barrier coast is reconstructed. Tillmann (2015) developed a similar rollover model for the German North Sea coast.

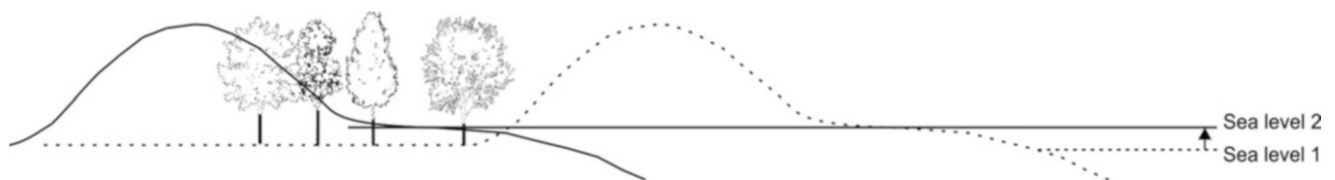
Simultaneously, the former land areas transformed into present-day spits, experienced a rise of a groundwater table due to the sea level rise; increased water content of coastal lowlands was conducive to the development of peatbogs, as well as for the growth of bushes, trees and even forests. Coastal lakes emerged every time the groundwater table rise was faster than the peat-forming process. Some of those lakes were being filled up by dune sands from barriers that were migrating due to the sea level rise (Fig. 5).

Those barriers that were moving further inland might have cut off parts of sea bays. When the sea level rise rate was faster than that of the dune berm rise produced by aeolian processes, the barriers were disrupted and inlets linking coastal lakes with the sea or flood deltas were formed as a result (Rosa 1984).

The functioning of moving barrier systems should be associated with the origin and gradual development of all the large accumulation landforms of the southern Baltic coasts. It was the sea level fluctuations modified by the vertical Earth crust movement (uplifting and subsiding) that were deciding on the development of both cliff and dune coasts (Furmańczyk and Musielak 2015).

The moving barrier effect mentioned above occurred in the Łeba Barrier area. A dune berm was formed there during the sea level fall that lasted several hundred years. During another episode of the sea level rise, the barrier moved inland landwards across a several hundred year-old forest. Nowadays, tree trunks are periodically unearthed on the beach and in the shallow nearshore bottom after storms (Fig. 4).

In many places along the coastal barriers where peat had been forming, there were separated swamps or shallow lakes. As the sea level was rising and the land was subsiding, a further transgression of barriers resulted in their gradual encroaching the peatbogs or coastal lakes. At the same time, erosion affecting the seafront side of the dune berm (rollover and barrier translation), the beach, and the seabed revealed peat outcrops at different depths within the nearshore zone (Fig. 5). This effect can be seen on aerial photographs of



Coastal Evolution in Microtidal Seas in Holocene, Fig. 3 A sketch of dune barrier migration along with the sea level rise

the coastal zone off the Dziwnów Spit near Międzywodzie (Fig. 6). Peat deposit outcrops are visible as dark patches, the bright areas being sands covering the peat. Peat outcrops occur at a depth of 1.5–2.0 m.

The barriers, at both ends, contact hills of moraine uplands or front moraines, the sand, and till slopes of which are being eroded by the sea. The material eroded this way migrates along the shore and nourishes the adjacent barriers. It is transported by waves and currents that depend on wind force and direction. The barrier size and thickness at sites of dune formation depend on the amount of sand material released by the erosion of the nearshore seabed or adjacent cliffs.

Present Responses of the Coast to Sea Level Fluctuations

The barrier coasts and peninsulas at the southern Baltic coast were evolving throughout the Holocene and the evolution has not terminated yet. The evolution produced all the sandy spits of the southern Baltic Sea (Fig. 2), including the largest. Sediment transport perpendicular to the shore, controlled by



Coastal Evolution in Microtidal Seas in Holocene, Fig. 4 Tree trunks revealed after the storm event on the beach of Łeba Barrier. Some trunks stand in situ, while the rest migrate along the shore (Mielczarski et al. 1990). (Taken from www.polskaniezwykla.pl) Photo: Jacek Meller

sea level fluctuations was a dominant factor in their formation, while the longshore transport resulted in their evolution.

The wind-generated waves markedly affect the seabed and the shore, transmitting a high amount of energy within certain spatial ranges and periods of time. As shown by field experiments, both wave parameters and water level changes at the shore, are crucial for morphodynamic effects of energy dissipation (erosion, accumulation). This concerns the entire spectrum of sea level fluctuations occurring in the nearshore zone, including storm surges, baric waves, infragravity waves, seiches, etc.

When the sea level is low, waves of higher parameters rebuild the seabed and drive the sediment transport perpendicular to the shore in the so-called “dynamic seabed layer” (Deng et al. 2014).

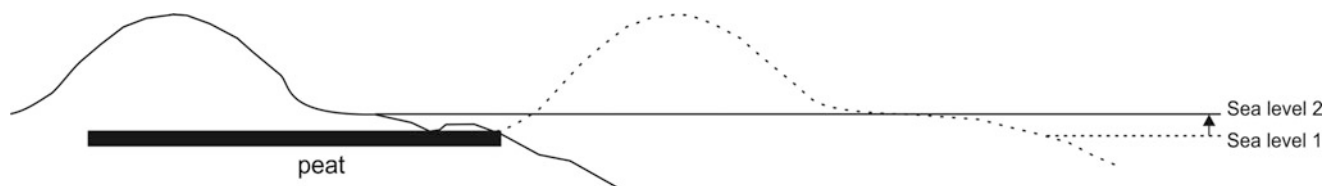
The rising sea level drives the wave energy dissipation zone inland, while the swash may reach the cliff foot or the dune base, triggering its erosion.

Erosion of a coast and the seabed affected by river inflow is the main source of sediment in the coastal zone. The material delivered this way moves along the coast or perpendicular to the shoreline depending on a resultant direction of the transporting factor (wind, waves, currents) and the underwater slope. As shown by results of long-term experiments and theoretical studies, the processes directed perpendicular to the shoreline are more diverse than those directed parallel to it. This is because the water velocity and sediment transport intensity are affected by complex and pronounced temporal and spatial changes in the short and long term. The following coastal forms are formed or transformed as a result of cross-shore transport: underwater bars, shore berms, beaches and, with an input of aeolian transport, coastal dune berms. If the direction of the transporting factor is oblique to the shore, a longshore sediment transport is triggered, its direction depending on the angle of approaching waves.

In the southern Baltic, sea level rise records have been maintained since 1811. The record series is shown in Fig. 7.

The process is modified by the North Atlantic Oscillation, which controls North Sea inflows to the Baltic.

The inflows are responsible for local periodical sea level rise in the southern Baltic. The sea level may rise by up to 1 m, the elevated level remaining for a couple of days or even weeks. It was also possible to identify periods of time (A) with the sea level fall of a few centimeters which usually takes



Coastal Evolution in Microtidal Seas in Holocene, Fig. 5 A sketch of peat outcrop origin

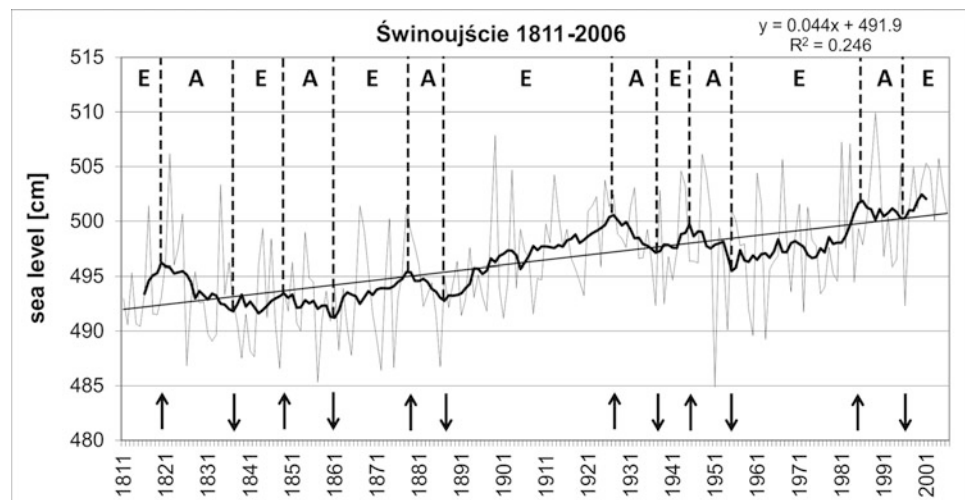
Coastal Evolution in Microtidal Seas in Holocene, Fig. 6

Aerial photograph of Miedzywodzie coastal zone (Dziwnów Spit) revealing peat outcrops and sand sediments. (Photo taken by GIS-Pro Szczecin company)



Coastal Evolution in Microtidal Seas in Holocene, Fig. 7

Graph of mean annual, 11 years moving average and linear trend, sea level fluctuation at Świnoujście between 1811 and 2006, divided on potential phases of erosion (E) and accumulation (A). (Musielak et al. 2017). Arrows mark the maxima and minima 11 years moving average within phases of erosion and accumulation



about 10 years. Even such slight falls resulted in a greater shoreline stability and even widening of the beach, sand accumulation within the foredune (also in front of the cliff), and in raising of the dune berm (Fig. 8).

Periods of time (E) which are basically longer than periods (A), are characterized by a rise of the mean sea level, which produces mainly narrowing of the beach and even its complete disappearance after storms. The foredunes, dune berm, and cliff foot are also damaged then. The periods (A) mean sea level stabilization or even a fall as well as an increase in the barrier dune volume or dune accumulation at the cliff foot.

Shore exposure to main directions of incoming waves (including the shadow cast by islands) and wave fetch length are of significant importance for the evolution of the coast. Other important factors include: locally differing rates of the

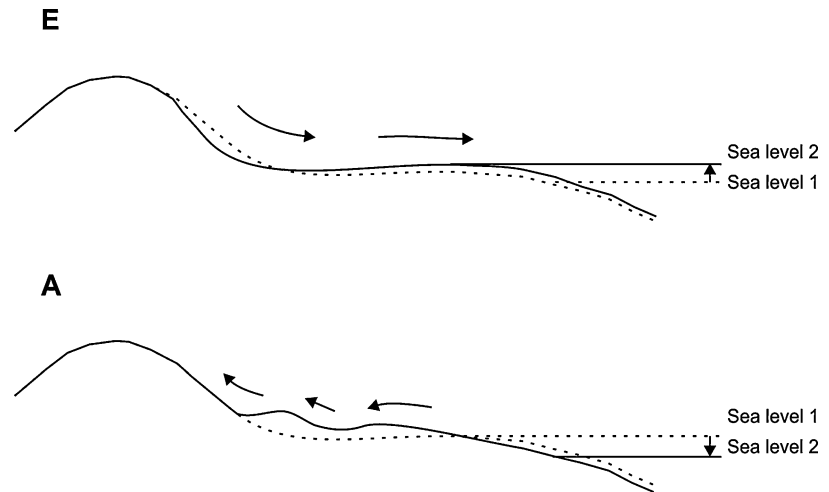
Earth crust vertical movement and the vulnerability to storm surge of a shore due to its specific location in a bay or a basin protected by a group of islands, as observed in the southwestern Baltic Sea. In addition, direction and passage velocity of cyclonic systems (low-pressure systems) which produce water swell (especially in bays) and control the direction of wind and waves affect the coast as well.

On a time scale of hundreds and thousands of years, all those factors were changing in line with climate changes. Hence, the coast evolution trends present in the past were changing too. Therefore, the magnitude and direction of changes observed along the southern Baltic coast cannot be treated as uniform (Musielak et al. 2017).

Adverse hydrological and meteorological conditions (strong waves, storm surge) coupled with the sea level rise,

Coastal Evolution in Microtidal Seas in Holocene, Fig. 8

A sketch of morphological results decades sea level changes: rising up – erosion (E) and riding down – accumulation (A). Without impact of vertical Earth crust movement



may result in extreme effects leading to catastrophic changes in the coastal zone, including dune breaking and lowland flooding (storm flood); they may also trigger cliff erosion, thus posing a hazard to the infrastructure located in the vicinity. On a time scale of tens of years, they occur irregularly.

Conclusions

Factors and processes that decide on evolution of shores made of sand or sand and till are very complex. The most important of those factors and processes are:

- Geological structure, tectonics, and vertical movement of the Earth's crust
- The progressing sea level rise and its periodical fluctuations
- Location-specific conditions such as sheltering by islands, shoreline azimuth, situation in an embayment, etc.
- Hydrological and meteorological conditions such as the dominant parameters of wind, waves, etc.

All these factors and processes influence the evolution of the southern Baltic coast. In tidal seas the factors will be slightly different, with a much higher of wave energy, and the presence of permanent longshore currents.

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Coastal Flood Hazard Mapping

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Synonyms

Coastal flood hazard mapping; Coastal flood hazard plotting; Coastal inundation threat plotting; Flood hazard analysis by mapping; Mapping and floodplain boundaries; Mapping of coastal floods

Definition

Coastal flood hazard mapping (CoFHaM) is a tool used to determine flood zone limits inland and in other areas exposed to coastal floods due to different hazards such as storm, surge waves, sea level rise caused by climate change, inland storm surge, heavy rainfall, among others. Its areas are identified through maps made by one or more researchers using the Geographic Information System (GIS) software. Generally, its maps display particular levels of flood hazards and risks, with different parameters defined as low, medium, and high.

Introduction

Floods are among Earth's most common and most destructive natural hazards (Nadeem et al. 2014; Gigović et al. 2017; Udin et al. 2018 p. 47). Most of these large floods occur in territories near the coast. The coastline represents an attractive zone for the establishment of various economic activities (IPCC 2007; Halpern et al. 2008). This singular space, over centuries, has been valued and has received high anthropic activity as well. The everyday presence of densely populated urban settlements located on the beachfront, raises concern in the scientific community, due to the different vulnerabilities

and risks these communities endure. Coastal communities are increasingly experiencing climate change – induced coastal disasters and chronic flooding (Lipiec et al. 2018). These relevant coastal problems have led to the development of sophisticated coastal flood hazard mapping (CoFHaM) techniques in recent years.

Coastal flood hazard mapping (CoFHaM) has evolved over time. The first maps only presented static models of historic flood extensions as a guide for planners (Coller et al. 2018). In recent decades, coastal flood hazard mapping has been perfected with the introduction of computer modeling and mapping. These contemporary techniques offer a much greater power of cartographic representation, and they permit experts to evaluate a variety of possible events of floods in numerous places (Coller et al. 2018). The development achieved by the digital mapping of coastal flooding and the wide variety of models available today, which are generated through the use of computer systems, has also replaced the traditional use of paper maps. CoFHaM tools have also been perfected and diversified through the use and facilities provided by web mapping. Some of these websites are accessible and allow their users to share information about flood hazards (Coller et al. 2018). Examples of this service are provided by the environmental agency for the flood map service of England (Environment Agency 2013) and the National Flood Insurance Program (NFIP) in the U.S. Consumers can visit www.FloodSmart.gov to learn how to prepare for a flood, how to purchase a National Flood Insurance Policy, and how to access the benefits of protecting your home and property from flooding.

Coastal flood hazard mapping (CoFHaM) also originates as an instrument for territorial planning which, in a preventive way, guides urban planners and government officials in general. Furthermore, it helps them in the decision-making process when dictating insurance premiums which ideally reflect the way this risk is overseen. CoFHaM makes it possible to prepare answers in advance when facing emergency situations in areas where engineering protection is not practical. CoFHaM has begun to play an important role in raising public awareness of flood hazards (Coller et al. 2018).

Coastal flood hazard mapping tools are essential in flood risk assessment, as well as land use planning, supporting engineering decisions concerning flood control measures, and coordinating emergency responses (Coller et al. 2018). Modeling inland storm surge flooding is also a fundamental step in flood hazard mapping. As the primary tool for communicating flood exposure, flood maps quantify the spatial variation of flood effects (Hatzikyriakou and Ning 2017).

CoFHaM is used for a variety of potential situations such as hazard mapping impacts, vulnerabilities, and risks by means of land, air, and water, e.g., for structural vulnerability estimation (Hatzikyriakou and Lin 2017), flood risk analysis (FEMA 2017), and for offshore wind resource forecast. They

are also used to broaden the understanding of diurnal coastal flows, to improve models for offshore wind power (Colle et al. 2016), and to assess flooding potentialities in rivers located in large areas (Speckhann et al. 2018). Furthermore, CoFHaM is used to simulate storm surge waves, coastal flooding (AMA 2014 a, b; FEMA 2003a) and tsunami hazards (Priest et al. 2018); to map habitats and develop baselines in offshore marine reserves (Lawrence et al. 2015); and to undertake the mapping of the areas protected by Levee Systems (FEMA 2003b), as well as the urban coastal area (Gigović et al. 2017) and the coastal radiological hazards (Martin et al. 2018) among others.

Some researchers have undertaken some important theoretical discussions about flood hazard mapping in coastal zones. There are authors and institutions which analyze different scenarios and coastal limits where there is an incidence or direct impact of the sea towards the land surface, e.g., storm surge waves or sea level rise (FEMA 2003a; Hatzikyriakou and Ning 2017). Other scientists analyze the flood hazard mapping considering the river basin as a determining part of the coastal environment (Udin et al. 2018; Speckhann et al. 2018; Milanés et al. 2017, FEMA 2009). This second position is considered a holistic one, because it integrates two flood hazard scenarios: (1) *the danger produced inland due to floods of rivers during their outfall into the sea and the combination of this scenario and the impact on its coastal limits after sea-flooding* (see Fig. 1) and (2) *the danger coastal floods produce on land due to sea-flooding or storm surge, as a result of a tropical storm, a hurricane of a different category or the gradual rise in sea level due to climate change* (see Fig. 2).

In Figs. 1 and 2, by using CoFHaM, a digital flood hazard database was created by compiling the best available data on flood-hazard areas and coastal risks; a systematic method to integrate coastal and riverine flood hazard areas was considered. In this case, coastal flood hazard boundary in land use planning was incorporated into the digital flood hazard database. Also, a combination of housing census data alongside the digital flood hazard database, using the geographic information system, satellite imagery, and digital photographs, was also employed.

The definition of coastal boundaries by nature of hazard is significant in the preparation of flood maps (Milanés 2018 a, b). Generally, each floodplain or flood hazard area is divided into flood insurance rates which are based on the floodplain boundaries determined on a work map (FEMA 2015, p. 1). Floodplain boundaries are delineated on the best available topographic mapping using the water-surface elevations determined at cross sections. Between cross sections, water surface elevations in riverine analyses are interpolated (FEMA 2015, p. 4).

In this entry, a critical review concerning different methods for coastal flood hazard mapping designed by relevant authors were shown. The materials, methods, and scales used are also described. Finally, two case studies using several methods and areas exposed to different coastal flood hazards in south-eastern Cuba are presented. In both case studies, coastal populations are exposed to different vulnerabilities and risks.

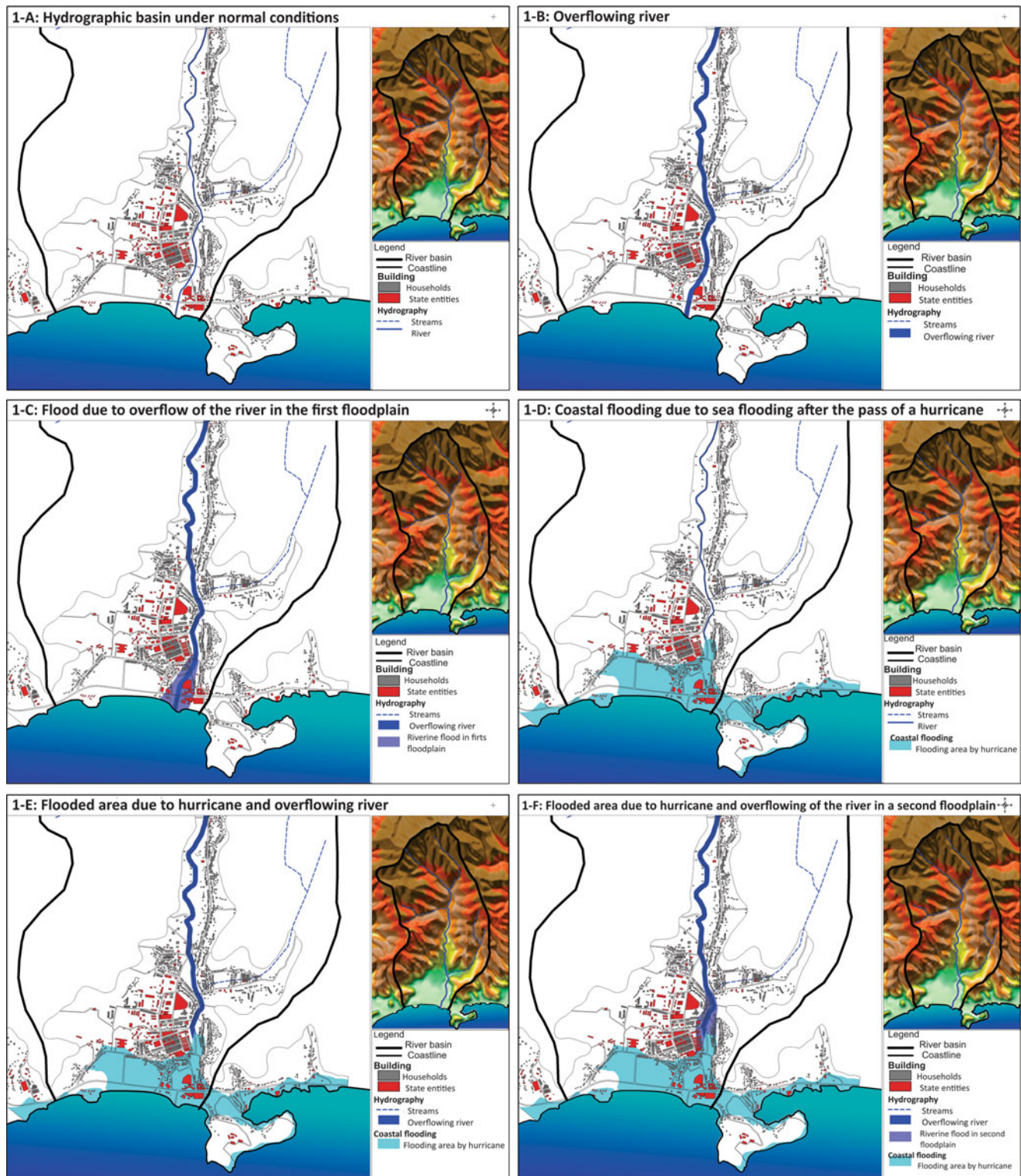
Methods and Tools Used for Coastal Flood Hazard Mapping (CoFHaM)

Each country determines the institutions which are responsible for centralizing risk maps. In the USA, flood mapping dictating insurance premiums which ideally reflect this risk is governed by the Federal Emergency Management Agency (FEMA) through its National Flood Insurance Program (NFIP) (Hatzikyriakou and Ning 2017; Crowell et al. 2007). FEMA prepares flood hazard mapping to create broad-based awareness of flood risk, provide data necessary for mitigation programs, and rate flood insurance for specific properties (FEMA 2015, p. 1). In Cuba, several institutions work in CoFHaM development, but the so-called “*Defensa Civil*” (The Cuban equivalent to FEMA) is the governing body which centralizes these studies and makes vital decisions.

Flood hazard mapping can be used for hydrodynamic models (Hatzikyriakou and Ning 2017; Dietrich et al. 2011). Other flood hazard mapping methodologies which combine flow frequency analysis with the height above the nearest drainage (HAND) model are applied in river basins. Its flood hazard mapping methodology, combines flow rate analysis of annual maxima with the HAND model (Speckhann et al. 2018), (see Fig. 3).

There are other tools of coastal flood hazard mapping which incorporate statistical analysis and meteorological modeling. There are others which use frequency curves over a range of potential rainfall and precipitation events (Coller et al. 2018). At times, the frequency analysis is also combined with the spatial modeling of the impact of floods on the curves of basic benefit costs (Morita 2008).

Figure 3 displays six methodological systems used by different authors for coastal flood hazard mapping. These tools were used in rural areas, river basins, coastal municipalities, and urban areas. A great difference is observed in the sequence of steps taken for these methodologies, which vary from three to eight stages according to their depth levels. Depending on the nature of the danger, Cuban methods differ in their ways of mapping floods; e.g., floods due to heavy rain (AMA 2014a) or due to sea floods (AMA 2014b). Other



Coastal Flood Hazard Mapping, Fig. 1 Schematic showing the forms of river flooding from its mouth to the sea. (a) Hydrographic basin under normal conditions; (b) overflowing river; (c) flooding due to river overflow in the section of the floodplain closest to the sea;

(d) coastal flooding due to entrance of seawater after the effect of a hurricane, without river overflow; (e) flooded area due to hurricane and river overflow; and (f) flooded area due to hurricane combined with flooding due to river overflow in a section of the floodplain further inland



Coastal Flood Hazard Mapping, Fig. 2 (a) Model of coastal flood scenarios limits when facing category, I, III, and V hurricanes; (b) model of coastal flood scenarios by sea level rise due to climate change for the years 2050 and 2100; (c) the Peninsula of “Los Galeones” is projected to be most affected area by the floods, which, according to the combined hazard models for the year 2100, will convert it into a key

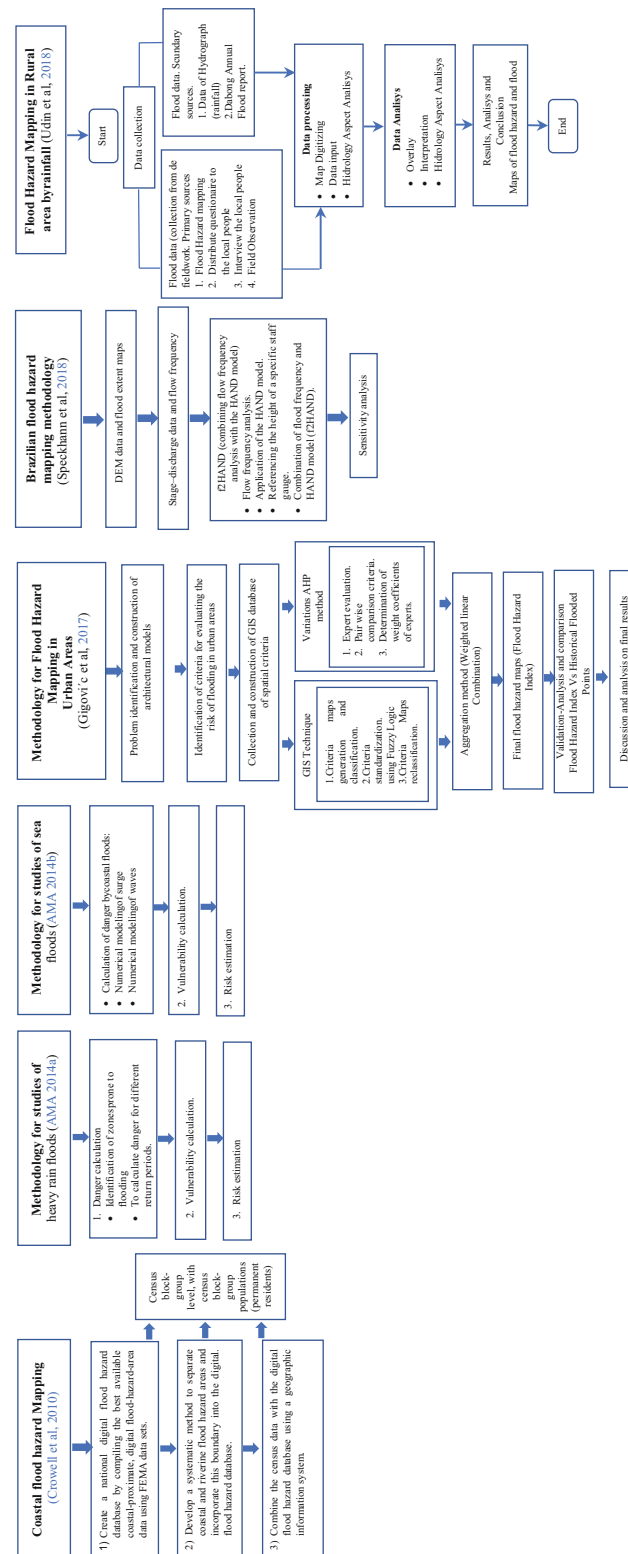
tools are used to separate the coastal and riverine zones and incorporate this new boundary line into the digital flood hazard database (Crowell et al. 2010).

Gigović et al. (2017) considers six factors which are relevant to the hazard of flooding in urban areas: *height, slope, distance to the sewage network, distance from the water surface, the water table, and land use*. Expert evaluations take into account the nature and severity of the observed criteria; and these are tested using three scenarios: (1) the modalities of the analytical hierarchy process (AHP); (2) the use of the fuzzy technique for the exploitation of uncertainty with the AHP method, and (3) contemplate the use of the traditional AHP method (See Fig. 3).

In order to communicate flood exposure, FEMA developed flood insurance rate maps (FIRMs) which have the purpose of classifying coastal areas into flood zones based on two principal criteria. The first is the area's base flood elevation (BFE), which represents the elevation of the estimated 100-year (i.e., 1% chance annually) flood event. The second criterion is the expected wave conditions associated with this event, specified as a range (e.g., “greater than 3 feet”). Another approach to evaluating FEMA's flood hazard mapping for a location is to compare it with simulated inland flooding from a historical event (Hatzikyriakou and Ning 2017, p. 954).

Material, methods, and scales used for coastal flood hazard mapping are diverse. For the application of coastal flood hazard mapping (CoFHaM) Geographical Information Systems (GIS), historical data on floods, satellite imagery, base flood elevations, wave and tide gage data quadrangle maps showing the location, designation, and elevation of all high-water marks in hard copy and as a GIS spatial data coverage are used. Some hydrologic models, contour mapping or digital elevation models, topographic data, multicriteria decision analysis, assessment of experts, nature and severity of observed criteria can also be considered. The latter are tested using different scenarios or analytic hierarchy process technique as well as satellite photos or digital photographs of each high-water mark, of each coastal infrastructure (such as roads, bridges, culverts, or canals). Other flooding, risk, and vulnerabilities factors are also mapped.

Other methods use aerial photography and on-ground reports from NOAA and/or the media (namely, media reports from source such as the Internet, newspapers, radio, mobile, television, smart phone), information provided by local and state emergency managers, and other federal agencies (i.e., governmental situation reports, mathematical models intended for riverine flood analysis, diagnosis and preliminary damage assessments, experts criteria, preliminary damage assessments for hurricane events, telephone logs and brainstorm among specialists). The final results can be successfully used for spatial city development. The work scales can oscillate from 1:2000 to 1: 25,000 or higher depending on detailed



Coastal Flood Hazard Mapping, Fig. 3 Different methodological schemes for coastal flood hazard mapping

studies. The smaller the scale the higher the levels of detail provided by the map.

On the other hand, the impact of models may change in accordance with and when compared to different models and countries. In the USA, areas within the one-percent-annual-chance (100-year) floodplain boundary are typically termed “Special Flood Hazard Areas,” and areas between the 1% and 0.2% annual chance (500-year) flood plain boundaries are termed areas of “Moderate Flood Hazard” (FEMA 2015) while in Cuba, areas that will be exposed to flooding by 2050 are considered moderate hazards, and areas already exposed since 2010 are in considerable danger.

Cuban Cases Studies

This section illustrates two case studies prepared in accordance with Cuban methodologies proposed by the Environment Agency of Cuba (AMA 2014 a, b). In the first case, flood scenarios are analyzed in the face of coastal hydrometeorological hazards combined with heavy rains, storm surge by category I, III, and V hurricanes, and sea level rise due to climate change from the 2050s up to the 2100 s. For the determination of assets and infrastructure exposed in the Primary Environmental Coastal Units for Integrated Management (PECUIM-Chivirico), located in the province of Santiago de Cuba, a multithreat and vulnerability analysis at a scale of 1:2000 was carried out. The models contemplate the vertical movements of the earth’s crust (*Neotectonics of the Cuban archipelago*).

The modeling of the flood scenario due to heavy rains was carried out following the calculations made by the provincial multidisciplinary group of hazard, vulnerability, and risk studies (CITMA 2013). The MapInfo version 12.0 Geographic Information System was used to obtain the flood areas for category I, III, and V hurricanes to also access the areas of Sea Level Rise due to Climate Change (SLRCC) and to generate all cartographic processing. For the raster format analysis, the Vertical Mapper module was used. Flooding areas were modeled considering the original values calculated by the Environmental Agency, which were validated after the passage of Hurricane Sandy through Santiago. The variables shown in Table 1 were incorporated into the studies.

Extreme floods through the years 2050 and 2100 s were calculated by the sum of the values of the SLRCC as a permanent flood scenario, and those obtained with the possible flood which may occur before a category V hurricane, (see Table 2).

For the analysis of the impact of sea level rise due to climate change in the area under study, the southeastern block of Cuba rise of 1 cm per year is taken into account (IGP 2010) while there are sectors lowering 0.1 mm per

Coastal Flood Hazard Mapping, Table 1 Calculation of the value of sea level rise (SLR)

Year	Estimated for CUBA (CITMA 2013)	Equinox high tide	Neotectonic	Final value of the SLR
2050	27.0 cm	50.0 cm	0.9 cm	77.9 cm
2100	85.0 cm	50.0 cm	0.9 cm	135.9 cm

Coastal Flood Hazard Mapping, Table 2 Calculation of the extreme SLRCC value at the PECUIM – Chivirico

Extreme situation with a category V Hurricane			
Valor del SLR (m)			
Year	SLR value due to climate change (CC h)	Category V Hurricane (HCT – V)	SLR value due to CC + HCT – V
2050	0.779	5.35	6.13
2100	1359	5.35	6.71

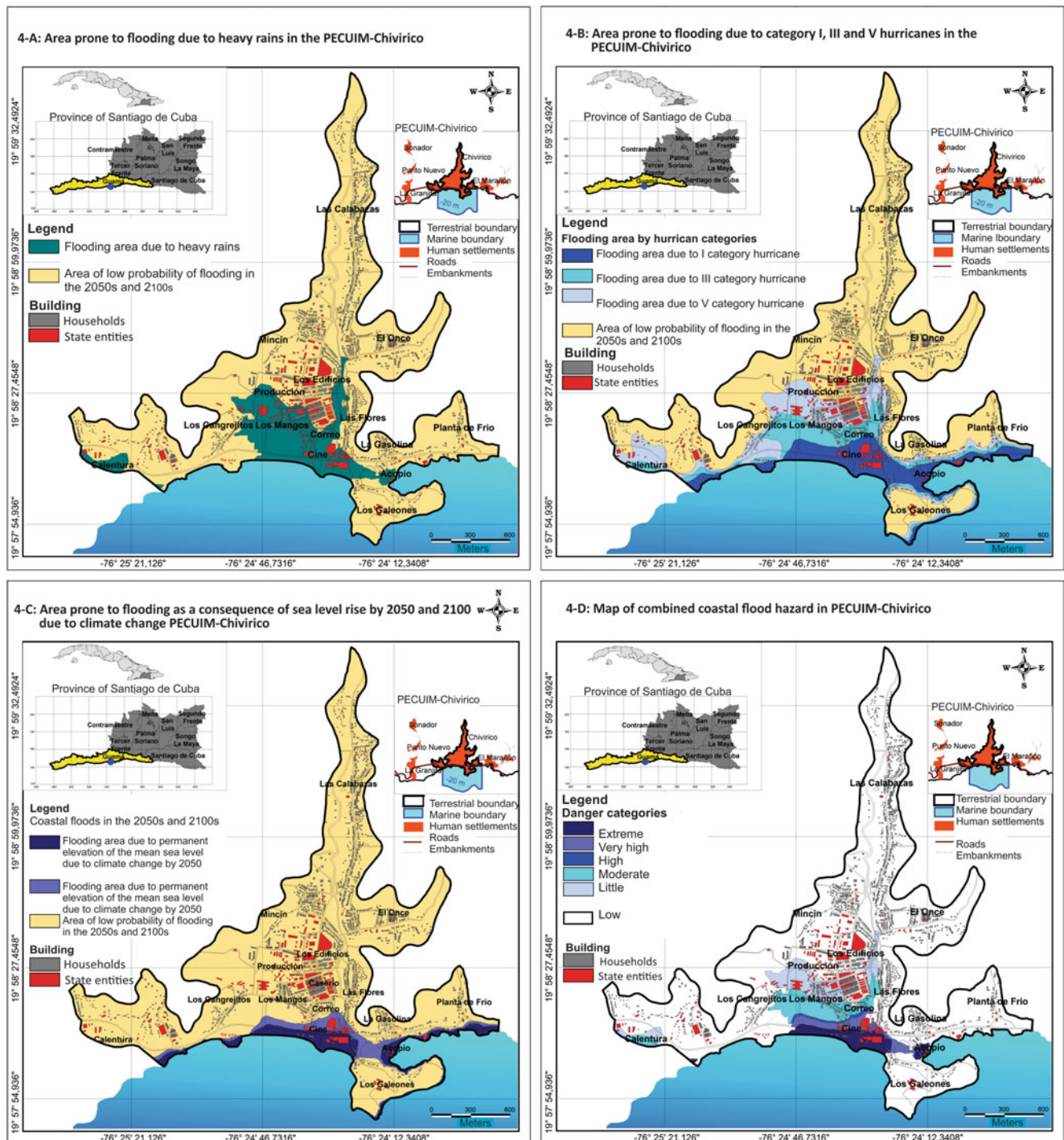
year, as is the case of the coastal settlement of the PECUIM Chivirico (Arango 2014). For this research both results are taken into consideration; and therefore, the resulting rise of 0.9 mm per year is the value that is used. As a final result of the first case study, four flood maps are provided taking different hazards into consideration (see Fig. 4).

The second case study corresponds to the flood scenario due to heavy rains for the province of Santiago de Cuba, which is located in the southeastern region of Cuba. In this case, the analysis is made 1:25,000 scale in the 11 municipalities of the province, where only two of them are coastal (Santiago de Cuba and Guama).

For hazard mapping, with the aid of the GIS, the layers of susceptibility factors or the scenario of occurrence of the hazard are combined with the trigger factor layer, which in this case corresponds to the isoyetic maps made for the three periods of return evaluated (5%, 10%, and 50% of probability). These periods consider the maximum daily rainfall that occurs in 24 h. The classification of the rainfall factor intervals is shown in Table 3.

For each return period, a rain map is obtained. This map is linked to the scenario of occurrence of the hazard. The multiplication of the layers is carried out considering ranges and weights established in the development of the research, with those obtained in the classification of the intervals of the rain factor in Table 3. The results of this combination provide the intervals for the classification of the hazard in Table 4. Finally, from this classification, different flood hazard maps for the different return periods are drawn up (see Fig. 5).

The province of Santiago de Cuba has an area vulnerable to flooding due to heavy rains of 401.7 km², which constitutes 6.5% of its total. Floods due to heavy rains occur mostly in



Coastal Flood Hazard Mapping, Fig. 4 (a) Area prone to flooding due to heavy rains in the PECUIM Chivirico; (b) area prone to flooding due to category I, III, and V hurricanes; (c) area prone to flooding as a consequence of sea level rise by 2050 and 2100 due to climate change;

(d) map of coastal flood hazard areas due to storm surge in categories I, II, and V hurricanes, combined with heavy rains and sea level rise due to climate change by the year 2100

municipalities with broad plains, low terrain and low permeable soils; among them, the two most affected and most extensive areas are the non-coastal municipalities of Mella and Contramaestre (Brito et al. 2013).

The floods in the coastal municipalities of Santiago de Cuba and Guama are smaller. These are produced mainly in the low areas and cones of dejection of the main rivers. In this area, the rivers have short courses and small basins. The inclination of its

Coastal Flood Hazard Mapping, Table 3 Classification of the rain-fall factor

Rain intervals in 24 h (mm)	Assigned value	Qualifier	Code on the map. Id
Greater than 401	4	High	3
Between 201 and 400	3	Medium	2
Less than 200	2	Low	1

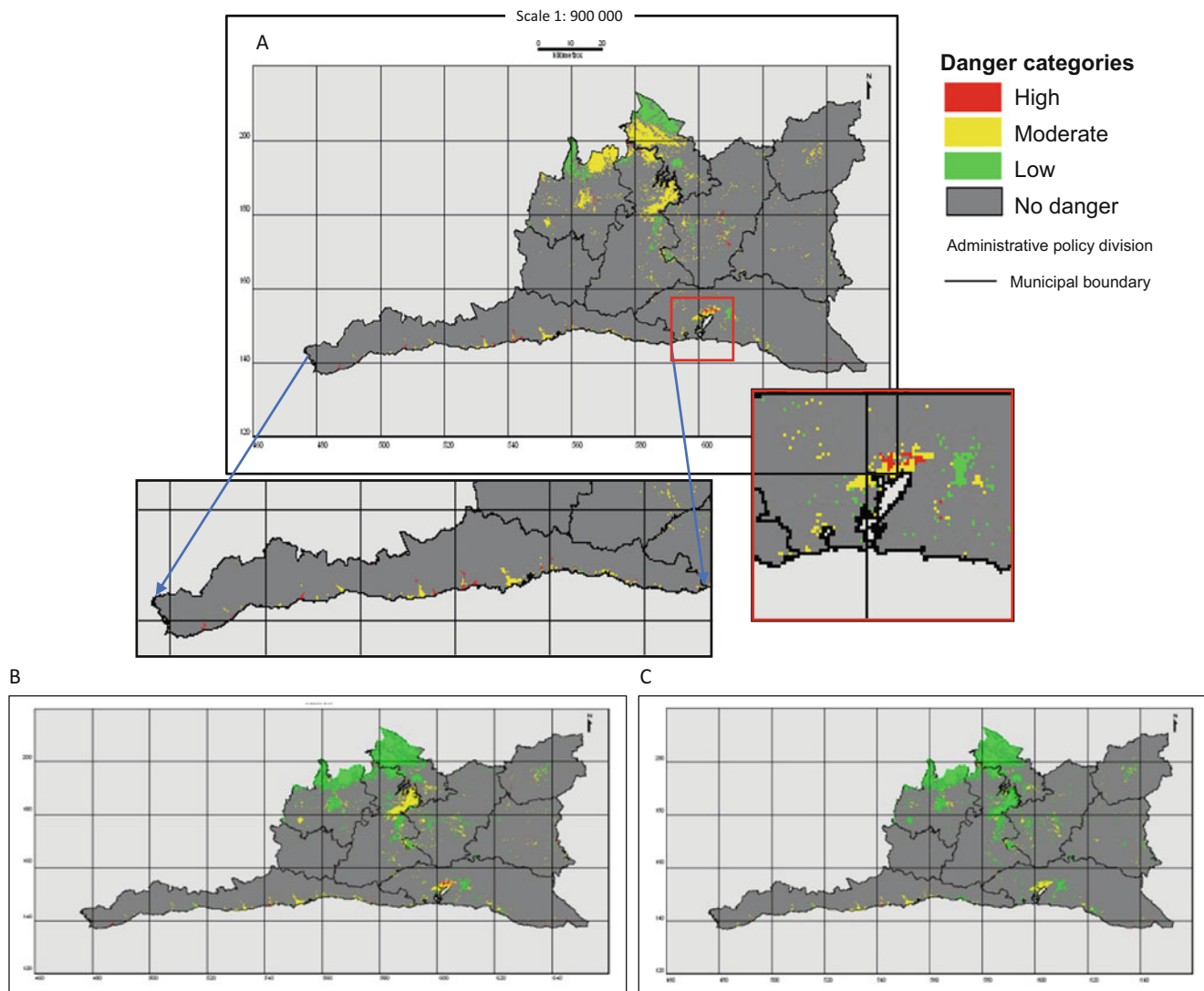
Coastal Flood Hazard Mapping, Table 4 Classification of the areas according level of danger

Ranges	Qualifier	Color on the map
0–6.6	Low	Green
6.7–13.2	Medium	Yellow
13.4–20	High	Red

slopes accelerates the runoff caused by the large accumulation of water in the low areas and, consequently, leaves large amounts of soil and materials of various origins behind, which interrupt road traffic and isolate these territories.

For the return period of 2% (Fig. 5a), the dangers associated with heavy rains for most of the territory of the province of Santiago de Cuba display a behavior that oscillates between medium and low values. The northwestern zone of the Santiago de Cuba bay has a high risk of flooding due to the presence of low areas and the confluence of four rivers of great importance for the municipality.

For a return period of 10%, the hazards associated with heavy rains have a more homogeneous behavior (Fig. 5b). This allows a large part of the territory to have low danger values. The areas located in the northern portion of the province are the most extensive. The areas

**Coastal Flood Hazard Mapping, Fig. 5** (a) Flood hazard for a return period of 2 years; (b) Flood hazard for a return period of 10 years; (c) Flood hazard for a return period of 20 years (Source: Modified from Brito et al. 2013)

characterized as moderately dangerous are distributed punctually in different places of the province; although, in the central part, the greatest extension for this range of danger is observed.

Those areas with a high hazard level are reduced to smaller sectors and are highlighted mainly on coastal areas. In this sense, the dejection cones of the six rivers – namely, *La Magdalena*, *La Plata*, *Uvero*, *Guama*, *Sigua*, *Baconao* – stand out.

For a return period of 20%, the dangers associated with heavy rains have a more balanced behavior, which is characterized by low risk values in a large part of the territory (Fig. 5c). Those areas characterized with moderate danger are distributed only in small scattered sectors of the territory and mostly in the areas of the bay of Santiago de Cuba, (see Fig. 5d). The areas with a high danger are reduced to only three coastal sectors: *Guamá*, *Sigua*, and *Baconao*.

Conclusions

Coastal flood hazard mapping helps in analyzing different hazard scenarios in the coastal zone, coastal urban areas, and river basins. In the case studies integrated in the river basin, two different types of coastal floods are mapped which often occur in a combined way: (1) from the sea to land (when the river flows into the sea) and (2) from the land to the sea (when the sea makes its way into the mouth of the river).

Mapping methods and CoFHaM tools are available to help urban planners and decision-makers not only to make effective use of the data but also to provide new knowledge to minimize risks. Flood maps play an important role in raising awareness among the population and urban planners concerning the different types of flood hazards in order to lay the groundwork for preparing responses to emergency situations.

The two case studies analyzed were developed in the same coastal province of the Cuban archipelago. Results achieved allow us to conclude that the CoFHaM tool is effective to determine different types of hazards through the use of varied scales and levels of action (e.g., from province to municipality or locality) with scales from 1:2000 to 1:25,000.

The first case study analyzed in Cuba ratifies the vulnerability of the archipelago to the forecasts of the rise of the mean sea level due to climate change, where maximum values of ascent of 0.59 m are proposed, for which adaptation and mitigation measures are needed for its confrontation. Furthermore, in the first case study analyzed, the hazards associated with coastal flooding due to combined hazards are classified into six parameters: extreme, very high, high, moderate, little, and low; while the results of the second case

study of coastal flooding due to heavy rainfall in the different periods of return (5, 10, and 50 years) are based on four parameters: high, moderate, low, and no danger.

Cross-References

- Coastal Boundaries
- Coastal Flood Hazard Mapping
- Coastal Risk
- Coastal Scenery

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Coastal Hazard Indicators

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Definition

A coastal hazard indicator is a measure of the state of coastal systems in relation to a given hazard, synthesizing complex information in a simple and resumed format. Coastal hazard indicators are commonly used to prioritize coastal vulnerability or assess the potential (or effective) impact of storms and should be used to facilitate coastal management actions. These indicators are often based on a parameter or set of parameters that characterize the coastal area, the driving mechanisms acting at the coast, or the interactive processes between coastal morphology and the underlying driving mechanisms. The coastal hazard indicators can, therefore, be classified into geoindicators, driver-based indicators, and process-based indicators.

Introduction

The understanding, analysis, and characterization of storm-induced hazards, such as coastal erosion or flooding, are extremely relevant for increasing the safety and reducing impacts of exposed population in coastal areas. Coastal hazards, namely, erosion and flood, are among the ones that generate more concern when analyzing the potential impacts of marine climate change, as expressed by a survey of 10,000 European citizens (Buckley et al. 2017). In a context of climate change, with sea level rise and changes in storminess, associated to the increasing occupation of the coastal areas, risks are also expected to increase in the future (van Dongeren et al. *in press*). The characterization of risks in a simple yet scientifically sound way can be performed using indicators of both hazard and exposure

(occupation). Coastal hazard indicators have been widely used for a long time (see Bush et al. 1999), based on multiple strategies and without a generic agreement or rule on the indicators to be employed. A large set of coastal hazard indicators can be found at the scientific literature, including morphological characteristics, wave and sea level parameters, and coastal evolution, among many others. The most commonly used parameters describe either the mechanisms acting over the coastal zone (driving mechanisms) or the coastal morphology/geology. The use of integrated parameters, which represent the processes associated with coastal hazards acting at the shore, is still in its infancy (Ferreira et al. 2017). A coastal hazard indicator should also integrate an associated time-scale (or return period), allowing a perception of the hazard potential evolution/change over time or within a specific time frame. Davidson et al. (2007) stated that an indicator must facilitate a management process, and, if that essential condition is not met, the term variable should be used instead. Thus, the main objective of coastal indicators is to support coastal management actions by synthesizing variables, processes, and information. In fact, coastal indicators have been used as tools to aid coastal management in the past, aiming to promote the interaction between coastal researchers and managers (Carapuço et al. 2016).

Storm-related coastal hazard indicators have been the subject of recent literature reviews by Carapuço et al. (2016), Nguyen et al. (2016), and Ferreira et al. (2017). Carapuço et al. (2016) identified and summarized the use of geoindicators. Nguyen et al. (2016) analyzed the use of all types of coastal vulnerability indicators, including geoindicators, hydrodynamic indicators, and coastal evolution indicators. Ferreira et al. (2017) reviewed the most relevant process-based indicators for the assessment of flood and erosion hazards in coastal areas.

Geoindicators

Coastal geoindicators refer and describe the physical/morphological features of coastal systems. For large-scale approaches, as the pioneer study of Gornitz et al. (1991), the geoindicators can be grouped into geology (expressing the relative resistance to erosion) and landforms (expressing the coastal type). A large number of geoindicators have been used, but several inconsistencies on their employment difficult the establishment of a coherent and common framework for their use. Carapuço et al. (2016), on their review, defined a framework for coastal geoindicators at sandy shores, providing details and insights on their use. These authors reported 43 indicators, targeting 16 coastal features, and were able to define a single and consensual criterion for the

majority of the sandy coastal features (for additional information on their use, see also Table 1):

Sandy environments in general: shoreline/coastline position, shoreline/coastline evolution, and sediment size/composition

Beaches: elevation, volume, width, and slope

Coastal dunes: elevation, volume, and width

Coastal barriers: elevation, volume, and width

Bush et al. (1999) included other geoindicators such as dune (alongshore) configuration, bluff configuration, or position relative to inlet location. These authors presented, however, a broader vision of the geoindicator concept with the inclusion of parameters such as erosion rate (a process-based parameter), the role of vegetation (an ecological-based parameter), or the existence of engineering structures (a result of human intervention). On rocky coasts, geoindicators also integrate the geological resistance (lithology and fracturing), the presence/absence of shore platforms, or the cliff height.

Driver-Based Indicators

Driver-based indicators are also commonly used to assess coastal vulnerability, namely, at large-scale applications involving semiquantitative coastal indexes (e.g., scaled in classes from 1 to 5 or from very low to very high). These indicators include the driving parameters related to wave characteristics, sea level, or wind. A compilation of the most commonly used driver-based indicators (namely, the hydrodynamic ones) and examples of their employment are provided at Nguyen et al. (2016). Driver-based indicators include (for details see Table 1):

Wave characteristics: height (e.g., average, significant and maximum), period, direction (exposure), storm energy, and storm duration

Sea level: relative sea level change, storm surge height, and tidal range

Wind: intensity/speed and direction

In contrast to the geoindicators, which exhibit alongshore shifts at a small spatial scale (meters), the variability of driver-based indicators (e.g., offshore wave characteristics, tidal range, wind characteristics) occurs at a larger alongshore spatial scale (tens to hundreds of kilometers). However, exceptions can occur at coastlines with a marked alongshore variability, such as the alternation of embayments and headlands, which often show a great variability on the exposure to waves and winds.

Coastal Hazard Indicators, Table 1 Comparison between computation type, return period calculation, short- (meters) to large-scale (tens to hundreds of kilometers) longshore variability, and inland expression of the hazard for selected geoindicators, driver-based indicators, and process-based coastal hazard indicators (Adapted and extended from Ferreira et al. 2017)

Indicator type	Indicator name	Computation type	Return period	Longshore variability	Inland hazard extent
Geoindicators	Shoreline position	DM/CE	No	Short scale	No
	Grain size		No	Short scale	No
	Barrier/beach elevation and width	DM/CE	No	Short scale	No
	Beach/coastal slope	DM/CE	No	Short scale	No
	Erosion rate	CE/F/M	No	Short scale	Yes
Driver-based indicators	Wave height	I/H	Yes	Large scale Short scale ^a	No
	Storm energy	F	Yes	Large scale Short scale ^a	No
	Sea level changes	F/M	Yes	Large scale	No/yes ^b
	Tidal range	I/H	No	Large scale	No
	Surge height	I/H	Yes	Large scale	No/yes ^b
	Wind speed	I/H	Yes	Large scale	No
Process-based indicators Overwash	Overwash depth	F/M	Yes	Short scale	Yes
	Overwash potential	F/M	Yes	Short scale	No
	Overwash extent	F/M	Yes	Short scale	Yes
Process-based indicators Flood	Flood depth	F/M	Yes	Short scale	Yes
	Flood extent	F/M	Yes	Short scale	Yes
Process-based indicators Erosion	Shoreline/berm retreat	F/M	Yes	Short scale	Yes
	Dune foot retreat	F/M	Yes	Short scale	Yes
	Vertical erosion	F/M	Yes	Short scale	Yes

CE cartographic extraction, DM direct measurement, F formulation, H hindcast, I instrumental, M model

^aRequires wave propagation models to include detailed longshore variability near the coastline

^bUsing a bathtub approach (or other flood extension approach) in association with the expected sea level change

Process-Based Indicators

Process-based indicators represent, in a simple way, the interaction between driving mechanisms and the coastal morphology and can be applied from large-scale to detailed approaches. These indicators are obtained through formulations or models that combine the driving variables (e.g., wave height, sea level) and the morphological parameters (e.g., berm height, beach volume). The resulting indicator has a physical meaning, directly related to the hazard, for example, the flood depth for coastal flooding or the shoreline retreat for storm-induced erosion. Process-based indicators are also able to discriminate the inland extent of the hazard, allowing the spatial (along- and cross-shore) representation of the coastal hazard. Process-based indicators, as most driving-based indicators, can be subject to a probabilistic analysis allowing the expression of their variability as a function of different return periods. Ferreira et al. (2017) suggested a set of process-based indicators for the assessment of hazards at sandy shores (for details on their use, see Table 1):

Overwash: depth, potential, and extent (Fig. 1)

Inundation: depth, extent, and discharge volume

Erosion (storm induced): shoreline/berm retreat, dune foot retreat, and vertical erosion (Fig. 2)

The authors also included recommendations on the use of these indicators and suggested formulas and models for their computation.

Process-based indicators can also be established for rocky coasts, as it is the case of the maximum width (inland extent) of cliff mass movements (see Teixeira 2006). This parameter can be used as an indicator that integrates all driving processes acting at the cliff and its geological resistance, being possible to establish return periods based on statistical analysis.

Summary and Final Considerations

Coastal hazard indicators have been often applied around the world to assess coastal vulnerability. The most commonly used indicators can be categorized into:

Geoindicators: related with the physical/morphological characteristics of the coastal zone

Coastal Hazard Indicators,

Fig. 1 Visualization of process-based indicators expressing overwash. Maximum overwash/overtopping depth over a seawall (*dashed black line*) and overwash extent (*solid black line*) at Fortaleza (Brazil) during a storm. Both indicators present a strong longshore variability, particularly the observed overwash depth



Coastal Hazard Indicators, Fig. 2 Visualization of process-based indicators expressing storm-induced erosion. Dune toe retreat (*solid black line*) and vertical erosion (*dashed black line*) at Fuseta Beach (Algarve, Portugal) after a major storm in March 2010. The dune toe

retreat includes an inland extent of the hazard, for the given storm (or associated return period). The vertical erosion shows a strong longshore variability as a function of changes on the natural dune height and effects of dune lowering by existing occupation

Driver-based indicators: an expression of the properties of the forcing elements acting at coastal areas

Process-based indicators: integrate driving mechanisms and coastal morphology representing the impact of the processes over the coastal forms

Most geo- and driver-based indicators are easy to obtain, since they can be extracted from existing cartography

(maps, digital terrain models, etc.) or instrumental measurements (or hindcasts), respectively (Table 1). These indicators can be translated into semiquantitative values (e.g., 1–10 scale), and their alongshore spatial distribution can be mapped, being a simple and often used proxy to express the alongshore changes in coastal vulnerability. They do not include the processes acting and transforming the coastal morphology, and, thus, they do not

represent the interaction between the morphology and the driving mechanisms. As such, geo- and driver-based indicators cannot be used as direct indicators of most hazards acting at coastal areas, such as overwash, inundation, or storm-induced erosion.

Process-based indicators use geo- and driver-based information and formulas to obtain a final representation of a given hazard, in association to a specific return period (Table 1). The use of process-based indicators implies a higher degree of expertise and, therefore, can reduce their applicability. Yet, these indicators allow an improved characterization of the hazard alongshore variability and often also include the cross-shore (inland) expression of the hazard (Table 1).

There is currently no universal application of coastal indicators or widely accepted intervals to classify each indicator according to the potential hazard (Carapuço et al. 2016; Ferreira et al. 2017). The lack of standardization of concepts and assumptions restricts the compared use of indicators among coastal areas (Nguyen et al. 2016). The selection of the appropriate indicator and associated thresholds is, however, fundamental to obtain proper hazard estimations. Thus, future studies addressing the classification, standardization, and definition of uses of coastal hazard indicators are required. Coastal hazard indicators should remain simple, unambiguous, and reproducible over different coastal types and regions. Indicators should be based on consistent methodologies, enabling the comparison across sites and expanding their usefulness for supporting coastal management actions.

Cross-References

- [Beach Erosion](#)
- [Beach Processes](#)
- [Coastal Management Practices](#)
- [Coasts, Coastlines, Shores, and Shorelines](#)
- [Setbacks](#)

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Coastal Hoodoos

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Introduction

Hoodoo is a descriptive and collective term used by Thornbury (1969) to describe such topographic forms as columns, pillars, and toadstool rocks of bizarre shapes. He recognized that differential weathering helped to develop and modify such outstanding topographic forms. In Bryan's study (1925), he also concluded that those he studied in the southwestern United States were chiefly the work of rain-wash plus effects of mechanical and chemical weathering. Basically, the term bears no genetic meaning, but was used as a descriptive term to describe those small landform features of odd shapes. Hoodoos occur in various environments and are associated with many different types of rocks. Previous studies of such hoodoos indicated that differential weathering almost always played an important role in sculpturing their forms. Hoodoos often became a scenic attraction rather than a geologic subject to be appreciated by visitors. Near Tuscon, USA, on the way to Mt. Lemmon, from Windy Point to Geology Vista, highly weathered granitic rocks

display hoodoos by the roadside. The place became a vista point and the hoodoos became an attraction for visitors to stop and take pictures. Actually, wherever hoodoos were found, whether they occurred in deserts, on coastal platforms, at the edge of highly dissected terrain or at riverbanks, they often became a site for visitors' appreciation. They are almost always recognized as earth's miracles and definitely as earth's heritage sites. In this regard, hoodoos are more of an artistic and aesthetic subject than a geological phenomenon.

Coastal hoodoos are relatively well studied. Trenhaile (1987) and Ollier (1984) described in detail the processes that caused the formation of hoodoos on coastal platforms and other coastal environments. Trenhaile pointed out that salt weathering which includes hydration, thermal expansion, and crystallization of various salts are important to form such features as tafoni and honeycombs.

He also pointed out that honeycombs are best developed in the zone above the level of normal wave action, but that periodically experiences heavy spray and splash during stormy weather. Bartrum and Turner (1928) first drew attention to the lowering, smoothening, and leveling of inter-tidal rock platforms by weathering processes apparently associated with pools of standing water. Water layering weathering includes alternate wetting and drying (slaking), salt crystallization, and many other chemical weathering processes. Collectively, they serve as a delicate process to sculpture hoodoos. Case hardening which often accompanies water layer leveling on platform surface often causes the formation of boxwork or frame-like patterns and weathering pits, which have raised rims. Some pits resemble miniature volcanoes with water-filled craters (Ollier 1984). He described elevated pits from Cape Patterson, Victoria as follows: "Small pits are filled by the tide or by rainwater and this keeps the immediately surrounding rock wet, while the platform around is reduced in height by wetting and drying. Eventually the pits comes to be like a miniature volcano with the crater filled with water and so kept wet, rising above a miniature plain which occasionally dries out." Other papers which described water layer weathering and salt weathering include Evans (1970), Gill *et al.* (1981), Goudi (1993), Mustoe (1982), Wellman and Wilson (1965). Pedestal rock is another type of hoodoo often seen in the coastal environment. Pedestal rocks are mushroom-like rocks, which are projections of solid bed-rock formed by flaking and hydration in a desert environment (Ollier 1974). Ollier also described pedestals from Cape Patterson, Victoria, and indicated that they were formed by blocks of arkose resting on a coastal mud-stone platform. Hills (1971) described similar features formed by a fallen block lying on a platform and

others formed by large calcareous concretions that now stand on pedestals above a water-layer lowered platform surface.

Hoodoos in Yehliu, Taiwan

Such hoodoos, including elevated pits, pedestals, tafoni, honeycombs, and others are well developed in the Yehliu Scenic Area in Taiwan. These hoodoos were studied by Hsu (1964), Lu (1978), Wang and Lee (1984, 1994), Hsu (1988), Hong and Huang (2001).

The most abundant hoodoos in Yehliu are pedestals, which are locally called mushroom rocks. One of the pedestals looks like a western queen's head and was called the Queen's Head (Fig. 1). Next to the pedestals is a layer with many Ginger Rocks, which are stand-up calcareous concretions originally occurring in a thin bed (Fig. 2). It is quite significant for so many concretions, widely distributed, but limited to a bed of less than 1 m.



Coastal Hoodoos, Fig. 1 Queen's Head, a pedestal rock

Coastal Hoodoos, Fig. 2

Ginger rocks, which are concretions stand out after lowering of adjacent ground

**Coastal Hoodoos, Fig. 3**

Candle rocks, which are elevated pits as described by Ollier (1984). At the very center is a concretion



This thin bed is the key bed that can be used for tracing and correlation of sedimentary strata. Candle Rocks are lower in the sequence stratigraphically (Fig. 3). They belong to another sandstone bed with randomly distributed concretions. Concretions exposed often led to the formation of elevated pits after uplifting and erosion. Furrows developed around the concretions, which were filled with water frequently. Surrounding the furrows is a zone of hardening which became relatively raised. After further uplifting, the Candle Rock becomes higher and an apron-like fringe area developed gradually. As a whole they are elevated pits as described by Ollier (1984). The

size of the Candle Rocks found in Yehliu may reach a height of 1.5–2.0 m, and a diameter of 2 m.

Geology of the Pedestal Rocks in Yehliu, Taiwan

The Yehliu coastal scenic area is located approximately 15 km to the northwest of Keelung City, Taiwan. The area is a promontory, approximately 1.5 km long and only about 400 m wide. Part of the promontory is so narrow that it is actually penetrated by sea caves or narrow openings along joints.

Strata exposed in Yehliu area are mainly bedded sandstone of Miocene age. These tilted strata strike N40E and dip at an angle of 18°–22° to the SE. The strata can be subdivided into eight units based on their lithological and morphological characteristics. Many sandstone beds are calcareous and rich in concretions. Pedestal rocks are mushroom-shaped, with a resistant head rock supported by a neck rock made of an underlying, less resistant rock. In general, a pedestal rock is 1.5–3 m high and composed of a head, a neck, and a base. The spheroidal head is 1.5–2 m in diameter and honeycombed. The neck, instead, is smooth-surfaced and thinner than the head. The distribution of the pedestals is clearly lithologically and stratigraphically controlled. They form a parallel and linear array regularly arranged on the tilted shore platform. The thin, calcareous sandstone bed in which the pedestals occur is tilted and protected from direct wave attack by another set of sandstone beds, which form small ridges. From top to bottom the upper rows of pedestals are about 2–4 m above sea level, and the lower pedestals are just above the sea level. The same sandstone bed exhibits no signs of pedestal development under the sea surface. A closer look at the pedestals clearly reveals that these pedestals start to develop once they left the sea surface. Gradually they became separated blocks to form pedestals with no neck, then with a thick neck, and then with a thin neck. Finally the neck broke, the head fell and the pedestal's life ended. The plane which connects the top of the pedestals inclines at an angle of 17°–20°, whereas the surface (actually the shore platform) which connects the bottom of the pedestals has an inclination of about 11°–13°. This is another indication that pedestals grew taller and higher as the shore platform was weathered lower. Those parts that rose above the sea earlier were weathered more. The pedestal rocks, which are higher in elevation, are taller and usually have a thinner neck. Those that stand lower are short and have a massive neck. The height of the pedestals are 1.5–3 m. The tallest is about 3.5 m. The diameter of the head may reach 1.6 m. The spaces between the pedestals average to a distance of 2–3 m. The total number of pedestals is about 180.

Owing to weathering of the calcareous sandstone, the surface crust of the head is dark colored compared to other parts of the pedestal and to those parts of the same bed below the sea level. Gauri (1978) gave a good explanation of the weathering process which formed such a dark-colored crust on a lime-rich rock.

Physical properties of the head and the neck of the pedestals are studied both in the field and in the laboratory. As a result of various means of mechanical tests, such as the Schmid Hammer Test, Slaking Durability Test, and Point Load Test, it is known that there is a difference in the physical strength of the different parts of the pedestals. The head of the pedestals is of much greater strength than the neck.

Water absorption by the head is also slightly lower than the neck, which indicates that the neck rock may have more open crevices that allow water to penetrate.

Microscopic study of thin sections of rock specimens from different parts of the pedestal also reveal important information. The head rock is composed of well-sorted quartz grains cemented mainly by calcite. The neck rock is composed of less well-sorted quartz grains but cemented by calcite and iron minerals. Thin-section study of the surface crust of the head rock, which is darker in color and honeycombed, indicates an enrichment of calcite and clay minerals as cemented materials (Wang and Lee 1994; Hong and Huang 2001). The result of microscopic study again supported the interpretation by Gauri (1978).

Wang and Lee (1984, 1994) in their study, concluded that differential weathering, including alternate wetting and drying, salt crystallization and other chemical processes accompanied by uplifting and erosion are responsible for the formation of the pedestals. Based on coral reef dating, estimated uplift rate of the area (about 2 mm/year), elevating of the red oil-painted safety markers on the platform (which is about 1–4 cm above the adjacent ground, and the first time the park placed the paint is about 50 years ago), and other evidences such as the height and the spacing between the pedestals, Wang and Lee (1984, 1994) estimated that the age of the pedestals is not more than 4000 years since their first appearance above the sea level.

Cross-References

- [Honeycomb Weathering](#)
- [Rock Coast Processes](#)
- [Shore Platforms](#)
- [Tafone](#)
- [Weathering in the Coastal Zone](#)

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Coastal Lake Systems

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Definition

To fulfill the definition being a “coastal lake” *sensu lato*, at least one property of a lake should have a relation to “coasts” or “oceans,” e.g., (part of the) form, sediments, characteristic of the water, hydrodynamics (e.g., tides), or organisms. These lakes often exhibit bottom and shore topographies solely deriving from the basin-forming process. A coastal lake *sensu stricto* should be in a long period of its history a part of the sea, either in its early time, or in its younger existence and into the future. The latter generally are termed “lagoons,” and many of them still have a (natural) connection to the open sea. Lagoons in most cases have an ocean-side barrier (barrier bars, barrier islands, spits, beaches, beach ridges, dune belts) formed by marine/littoral sediments and processes. These began to grow with slowing down of the postglacial sea level rise around 7,000 years BP (Anima 1990; Antikueira et al. 2005; Aritzegui and Wildi 2003; Carvalho do Amaral

et al. 2012; Dias and Fernande 2006; Dillenburg and Hesp 2009; Giannini et al. 2007; González-Villanueva et al. 2015; Hein et al. 2013; Kaiser et al. 2012; Longhitano et al. 2016; Otvos 2005; Peek et al. 2014; Perillo et al. 1999; Scheffers et al. 2012; Scheffers and Kelletat 2016; Tagliapietra et al. 2009; Toldo et al. 2000).

Forms and Types of Genesis

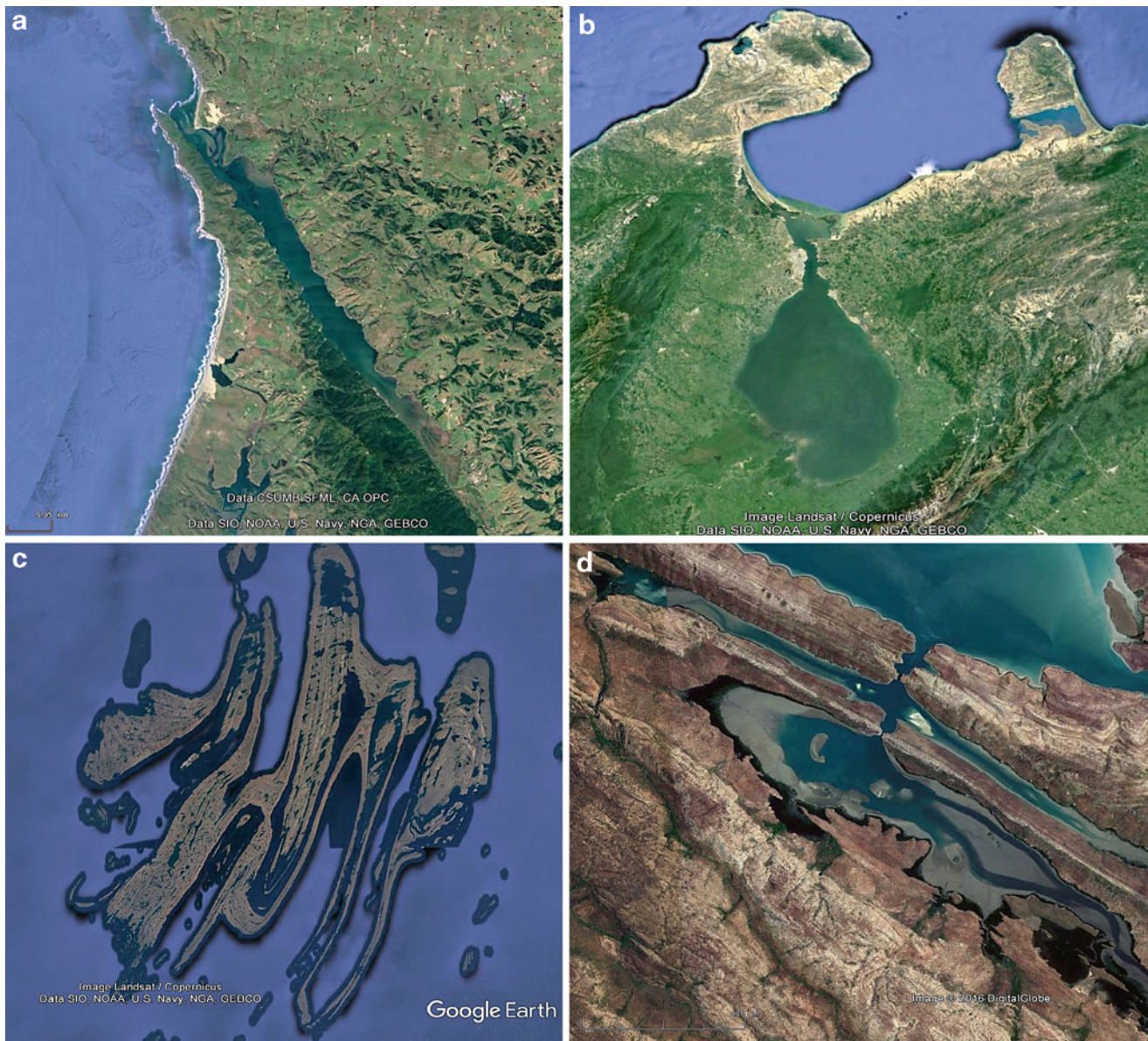
Depressions housing coastal lakes may have in their genesis no relation to marine or coastal processes being of volcanic/structural, glacial, periglacial or fluvial origin, getting their relation to the sea by tectonic uplift or subsidence, glacioisostatic emergence, or general (eustatic) sea level variations (Figs. 1, 2, and 3).

Coastal lakes with spatial and genetic relation to coastlines may occur in all latitudes and worldwide. As a lake (or pond) needs a depression filled with water, all exogenic processes actively forming depressions (like wind with deflation, glaciers with erosion, karst with solution) may leave basins. Along (formerly) glaciated coastlines, they are abundant in rocky landscapes emerged and emerging by glacioisostatic uplift. Other processes do not initially form depression but may leave basins by blocking originally open depressions like valleys by beaches, beach ridges, or coastal dunes. A high number of coastal lakes exist along sedimentary and low coastal relief, and their size may reach thousands of km².

Coastal lake basins may be closed by coral reef construction, eolian deposition, beach ridge systems like barriers and spits, as open spaces between ridges during delta progradation, or blocked by sea ice pressure on beaches (Figs. 4, 5, 6, and 7). The closure of lagoons in most cases are built by marine (and littoral) processes using (fine) sediments from the upper shelf in a direction to a former coastal indentation, or block openings by longshore drift.

Water Movements and Segmentation

Many coastal lakes and lagoons are isolated from the open sea. Nevertheless, their proximity to the ocean, a low relief, and in particular a large size of open water (or the longest lake axe in direction of strong winds) allow waves to initiate longshore drift processes along with their shorelines so far as these are not occupied by wetlands. Spits develop often in a rhythmic pattern (Figs. 8, 9, and 10), and a segmentation of the lagoons may be the result. Winter ice cover moved by strong winds also plays a role in this process, and some of the individual water bodies originate from thermokarst features in high latitudes.



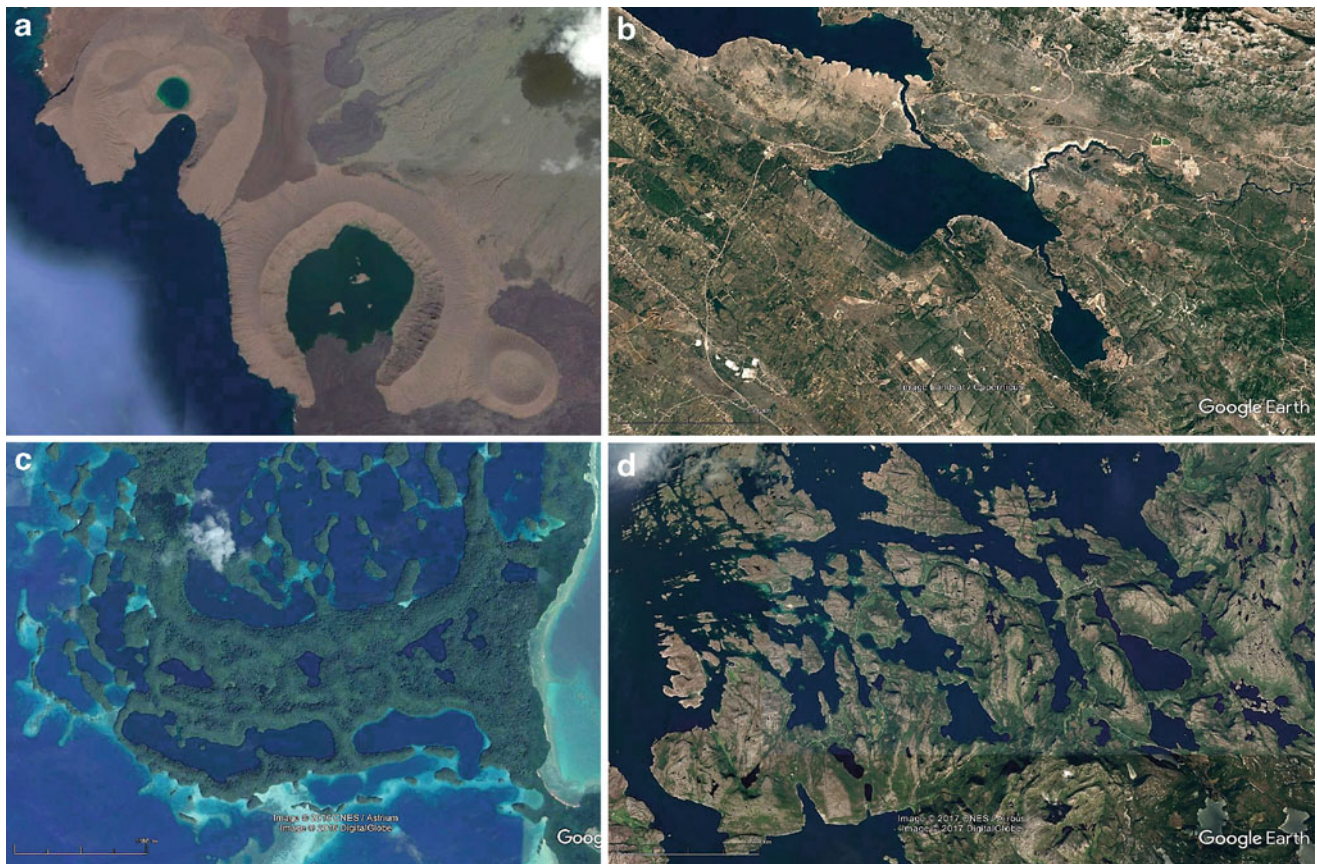
Coastal Lake Systems, Fig. 1 Coastal lakes deriving from endogenic (structural) control. (a) 22 km long lake on the San Andreas Fault, Point Reyes, California (USA), closed in the north by beaches and dunes. (b) Lake Maracaibo, a structural depression between folded mountains in NW Venezuela, >13,000 km² large, with a connection to the Caribbean via an old valley but blocked also by young Holocene beach ridge systems. The lake belongs to the oldest in the World (>20 Mio years). Max. depth is 60 m. Semidiurnal tides enter the lake and bring salt water to the northern parts, while the southern sections are fed by freshwater streams. Strong oil drilling results in a significant subsidence of the

eastern coastal section of up to 0.2 m/year. Scene is >400 km wide. (c) Belcher Islands in the SE of Canada's Hudson Bay are wide foldings cut and polished by Ice Age glaciers. They contain many lakes of different forms and openings to the sea along less resistant rock layers. The whole scenery (120 km wide) is rising from the sea since thousands of years. (d) Along the Kimberley Coast of NW Australia, long-term down warping of the land and Late Holocene sea-level rise have flooded structural depressions. Their entrance for seawater is two "horizontal waterfalls" as tidal range here is very high. Scene is 10 km wide (Credits: Google Earth)

Suspension, Soluble Matter, and Sediments

Lagoons fill depressions in the coastal landscape, and as thus, they have a function as sediment traps (Adomat and Gischler 2015; Carreira et al. 2011; Carvalho do Amaral et al. 2012; Murray and Mohanti 2006). These sediments may come from the terrestrial catchment area via rivers and hill-wash (the

latter in particular with soil sediments after forest clearance and land use), or as organic sedimentation from flora and fauna in the lagoons. Eolian deposition (deriving from beaches along the barriers, sometimes as shifting dune belts) also contributes to sediment infill. Sediments from the sea, either via openings in barriers by tidal currents, or by storm wave (and tsunami) wash-over processes may mark specific



Coastal Lake Systems, Fig. 2 Coastal lakes originating from non-coastal processes. (a) Two coastal lakes of volcanic origin on the Galapagos archipelago (Ecuador): the small one dammed by a crater ring of tephra, permeable for sea water; the larger one – once open to the ocean – blocked by a younger lava also permeable for sea water. Scene is 7 km wide. (b) Karst depression (small poljes) at the Croatian coastline, drowned by postglacial sea-level rise via river-like connections to the sea.

Scene is 30 km wide. (c) In the cockpit karst landscape of the Rock Islands of Palau (Micronesia), many lakes are connected to the ocean by karstified joint patterns or even caves, and marine organisms (e.g., millions of jelly fish) may live in these lakes now. 6 km wide scene. (d) In a glacier-sculpted landscape of northern Norway, lakes emerged from the sea by late- and postglacial glacioisostatic uplift. Their level now is between 1 and 44 m asl (scene 18 km wide) (Credits: Google Earth)

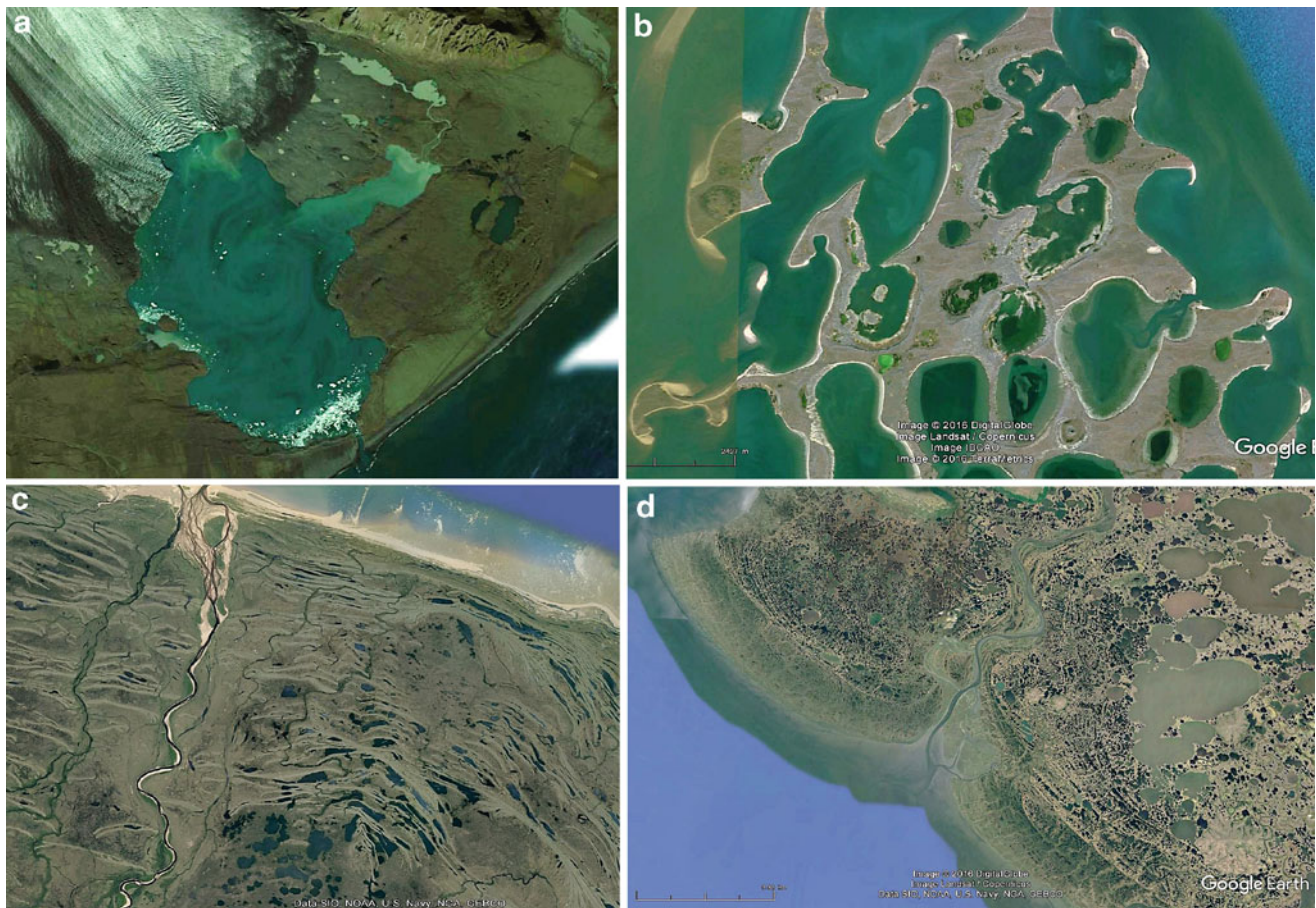
event layers in lagoonal geoarchives (Atwater and Moore 1992; Bondevik et al. 1997; Chaumillon et al. 2017; Fitzgerald and Miner 2013). The physical isolation from the ocean prevent lagoon sediments from disturbance by waves, but their proximity to the ocean also allows to record specific and extraordinary events. Coastal processes including sea-level variations, therefore, are stored in lagoons often in detailed records, and in the case of seasonal climates, even with wave-like layers in the resolution of months, supported by palynology using precipitation of pollen. Paerl et al. (2006) studied the influence of hurricane impacts of the last decades in Pamlico Sound (east coast of USA), and Grand Pre et al. (2011) identified warm periods with sea-level rise impulses from the content of benthic foraminifera in Pamlico Sound and planktonic foraminifera from the Gulf Stream in sediment cores. Maddox et al. (2008) found an 11,600-year long history of sea level, salinity, sediment, and nutrient changes in the Matagorda and Lavaca estuary complex of Texas (USA).

As long as in suspension, the color of lagoonal waters may indicate the origin of the sediment input (Figs. 11a, b), while vivid coloration may point to soluble matters (Figs. 11c, d). Chemical deposits even occur at the surface by evaporation in semi-arid and arid climates as in salt pans (Fig. 12).

Water Budget, Evaporation, and Salt pans

The water budget of a lagoon depends from at least five different sources:

- Run-off from the terrestrial relief including river discharge.
- Fresh ground water (identifiable by its radon content, compare Sadat-Noori et al. 2016).
- In-flow from the sea by open channels.
- Percolation of seawater through barriers.



Coastal Lake Systems, Fig. 3 Coastal lakes from glacial, periglacial, and glacioisostatic processes. (a) Glacier lagoon (Jökulsárlón) at the site of a former Vatnajökull outlet glacier at the southern coast of Iceland (scene 12 km wide). Suspension particles in the water show a fine gyre. (b) Thermokarst lakes on permafrost in northern Canada partly opened to the sea by thermoabrasion (scene 10 km wide). (c) Lakes dammed by a

series of glacioisostatic uplifted beach ridge systems between sea level and 40 m of elevation in this 13 km wide scene along SW of Hudson Bay (Canada). (d) A combination of glacioisostatic uplift of beach ridges and thermokarst on permafrost along the coast of western Alaska (USA) left thousands of small lakes. Scene 100 km wide (Credits: Google Earth)

- Lastly wash-over from extraordinary events in the ocean like winter storms, tropical cyclones, or even tsunamis.

“Expenditures” of the budget are in the first order evaporation (depending on the relation of lake size and depth/water volume as well as climate) and – during high lake levels – possible percolation through barriers into the sea. Variations of lagoon levels also depend on tides and irregular sea-level oscillations by differences in air pressure or during storm surges.

Even if a lagoon is blocked from the open sea, salt spray and splash from storms rise salinity as does the fresh water from land during evaporation. The chemical composition allows to identify the source of saline deposits (evaporates) in lagoons with negative water budget (Fig. 12). Saltpans, often colorful by bacteria, are found in high numbers along all subtropical coastlines of the world.

Anthropogenic Changes

During potentially thousands of years of existence, lagoons may pass alterations from climate and sea level variations, but especially from anthropogenic impacts in much more intensity and shorter intervals (Anthony et al. 2009; De Wit 2011; Harris 2006; Pérez-Ruzafa and Marcos 2005, 2008; Pérez-Ruzafa et al. 2011; Schallenberg and Schallenberg 2012; Terrell 2002; Thompson and Flower 2009; Wazniak et al. 2007). These may have different causes, some of them indirectly (e.g., interference of river run off by reservoir construction, forest cleaning, agricultural measures in the catchment areas, acidification), many of them directly (e.g., artificial closures or openings, discharge of pollutants and eutrophication, economic exploitation for chemical industries, or aquafarming (compare Fig. 13). Tourism industry, boating, and fishing are other negative impacts on lagoonal



Coastal Lake Systems, Fig. 4 (a) SE coast of Egypt at the Red Sea with a shallow depression blocked by active reef growth (scene 7 km wide). (b) A deep basin blocked by a fringing coral reef at the Red Sea coast of NE Sudan (scene 30 km wide). (c) Flooded lower section of

valleys (limans) south of the Amazon delta in Brazil, blocked by active sand drift (scene 25 km wide). (d) Partly active coastal dunes block a lower river course to form a ria-like lagoon (a liman) at the east coast of China. Scene is 22 km wide (Credits: Google Earth)

ecosystems, as is the danger by invasive species. Flocks et al. (2009) present their results from an investigation of large Lake Pontchartrain in southeastern Louisiana: most contaminants come from roadways and agricultural sources, leading to degradation and decline of benthic organisms as shrimps and clams. They found a significant increase of trace metals (Cu, Pb, Zn, Ni) in a mixing zone of the upper 1 m of lake sediments, a depth to which bioturbation and water turbulence combine. Similar conditions are found in the Gippsland Lakes in southeastern Australia. These are freshwater lakes sometimes with salt intrusion when rainfall causes lake levels to rise and water breaks through the dune belt. They suffer strongly from land use and forest cleaning in

catchments as well as from wastewater discharges by urban development with pollution and increased nutrient loadings resulting in algal bloom. Habitat loss and destruction of freshwater wetlands by salt ingressions are additional impacts on lake ecosystems.

Main Factors in Ecosystems of Coastal Lakes

Coastal lake systems exhibit the same factors as all other lentic systems, more or less modified by marine influences (Chagas and Suzuki 2005; Coutinho and Seeliger 1984; Ferrarin et al. 2014; Gönenc and Wolflin 2004; Hennemann et al. 2011; Liu



Coastal Lake Systems, Fig. 5 Lagoons as young coastal lakes dammed by sediment deposition. (a) In a river delta in SE Turkey, a lagoon is fringed by beach ridges (scene 35 km wide). (b) On St. Vincent Island in NW Florida (USA), lakes occur in basins within

beach ridge sequences (scene 16 km wide). (c) and (d) in northern and eastern Siberia (Russia), lagoons are closed by ridges formed from sea ice pressure on beach gravel (c: 10 km wide scene (Detail: 2 km long), d: 2 km wide scene) (Credits: Google Earth)



Coastal Lake Systems, Fig. 6 A rare type of lagoons have long axes more or less perpendicular to the coastline. These are limans (from the Greek word for harbor). They fill the lower section of valleys carved during sea-level low-stands. The barriers indicate rather shallow water conditions, and the missing infill from fluvial sediments point to only

small transport competence of these rivers. The left example shows a 65 km wide landscape in eastern Russia; the right one documents a fine set of limans at the south coast of Martha's Vineyard on morainic deposits (Massachusetts, USA), in a 6 km wide scene (Credits: Google Earth)

et al. 2013; Lory et al. 2010; Paerl et al. 2006; Selig et al. 2007; Smol and Douglas 2007; Thomaz et al. 2008). Because of their shallow nature, they may have no profundal aphotic region but a littoral, benthic, and open water (limnetic) zone.

The physical parameters, as well as chemical factors, play all an important role, triggering – among others – trophic relationships. Hydrodynamics as well as water budget, however, may be strongly influenced by marine intrusions (of salt



Coastal Lake Systems, Fig. 7 (a) 15 km long spit at 60° N and 166° E in eastern Russia, protecting a young delta from waves and current. (b) 18 km long liman with spits growing from both sides in western Alaska. (c) Barrier islands with wash-over fans into a lagoon on Prince

Edward Island, eastern Canada. (d) An unusual seaward curving barrier at the northern shores of the Sinai Peninsula (Egypt) point to very shallow water conditions by sedimentation from the Nile delta. Scene is 85 km wide (Credits: Google Earth)

water, organisms, tides) and, because of its neighborhood to the open ocean, by strong winds, sea ice, and even tsunamis. Other factors, in particular temperature regimes, community patterns, and diversity, depend on latitude and types of climate.

Generally, the width and depth of an opening influence the amount of seawater, chemistry, stratification, and mixing in a lagoon, as well as tides and currents in the basin (Kirk and Lauder 2000). Intermittent openings and closures, depending on higher tidal range, are described by Liu et al. (2013) for SE Australian lagoons. Significant changes for the ecosystems are also found by Cheng (1981) for New South Wales lagoons (Australia), where high floods from the land may sweep away sandy barriers, followed by their constant rebuilding from high waves. This leads to salinity variations from freshwater to hypersaline conditions. All these factors influence mixing and nutrient balance of coastal lakes.

Lakes, in general, are classified according to the intensity and frequency of intermixing into at least eight different types:

Amictic are those without any significant turbulence because of (nearly) permanent freezing as in Antarctica, the high Arctic or in extreme mountain elevations.

Oligomictic lakes mostly have no full circulation and occur in tropical lowlands.

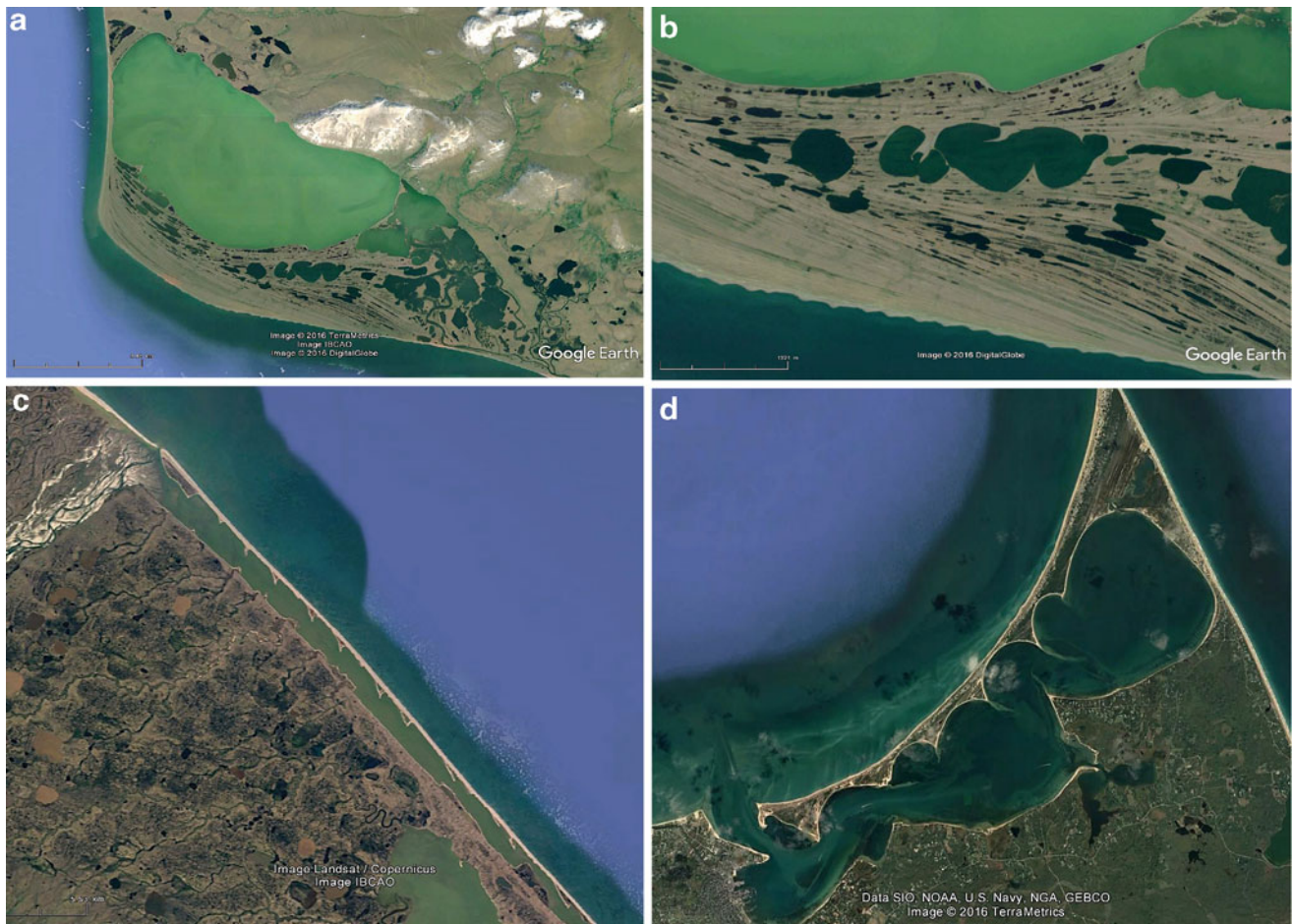
Meromictic lakes are only mixed in the surface water strata but not in deeper parts.

Holomictic lakes experience at least one full circulation per year.

Cold monomictic lakes as in polar and subpolar latitudes may have one full circulation in summer but no stagnation because of the brief period of the warm season.

Warm monomictic lakes occur in warm middle latitudes and the subtropics and have a significant stagnation time in summer. During cooling down to the winter period, homothermy happens and winds may easily mix the whole water column.

Dimictic lakes have two significant full circulation periods (spring and autumn) and two stagnation periods (summer and winter) and are found in the northern parts of North America and in eastern Europe.



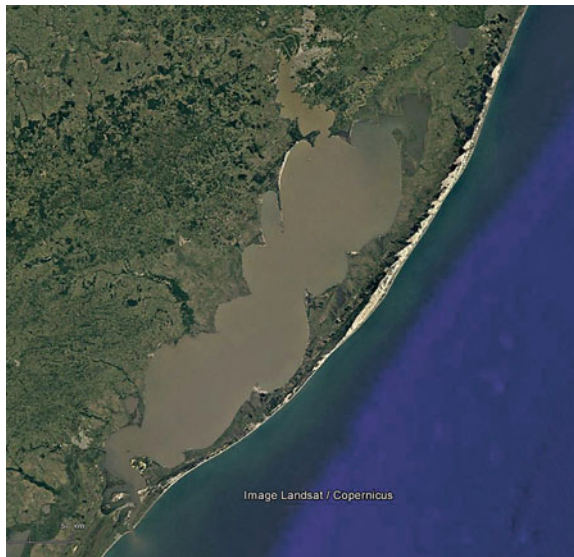
Coastal Lake Systems, Fig. 8 Lagoon landscapes with internal segmentation of the water bodies by spits. (a) A 22 km wide scene from western Alaska. In the 5 km long southern section (b), thermokarst evidently has initialized lake building. (c) A narrow but long lagoon is segmented by rhythmic spits forming along the inner fringe of the

barrier, probably by lake ice movement and pressure processes. This scene in northeastern Siberia (Russia) is 25 km wide. (d) Significant segmentation in a 10 km long lagoon at the NE spit of Nantucket Island, Massachusetts, USA (Credits: Google Earth)



Coastal Lake Systems, Fig. 9 (a) This 30 km long lagoon in northeastern Siberia (Russia) exhibits a final segmentation by similar shaped individual lakes. Smaller lagoons along seaward barrier sections show similar forms in a smaller scale. Thermokarst on permafrost at 70°N and pressure by lake ice certainly have contributed to these patterns.

(b) Lagoa de Araruama in Brazil, 45 km long, is also segmented similar to arctic examples, but hereby spit building from longshore drift along the lakeside flank of the barrier, initiated by easterly winds (trade winds) (Credits: Google Earth)

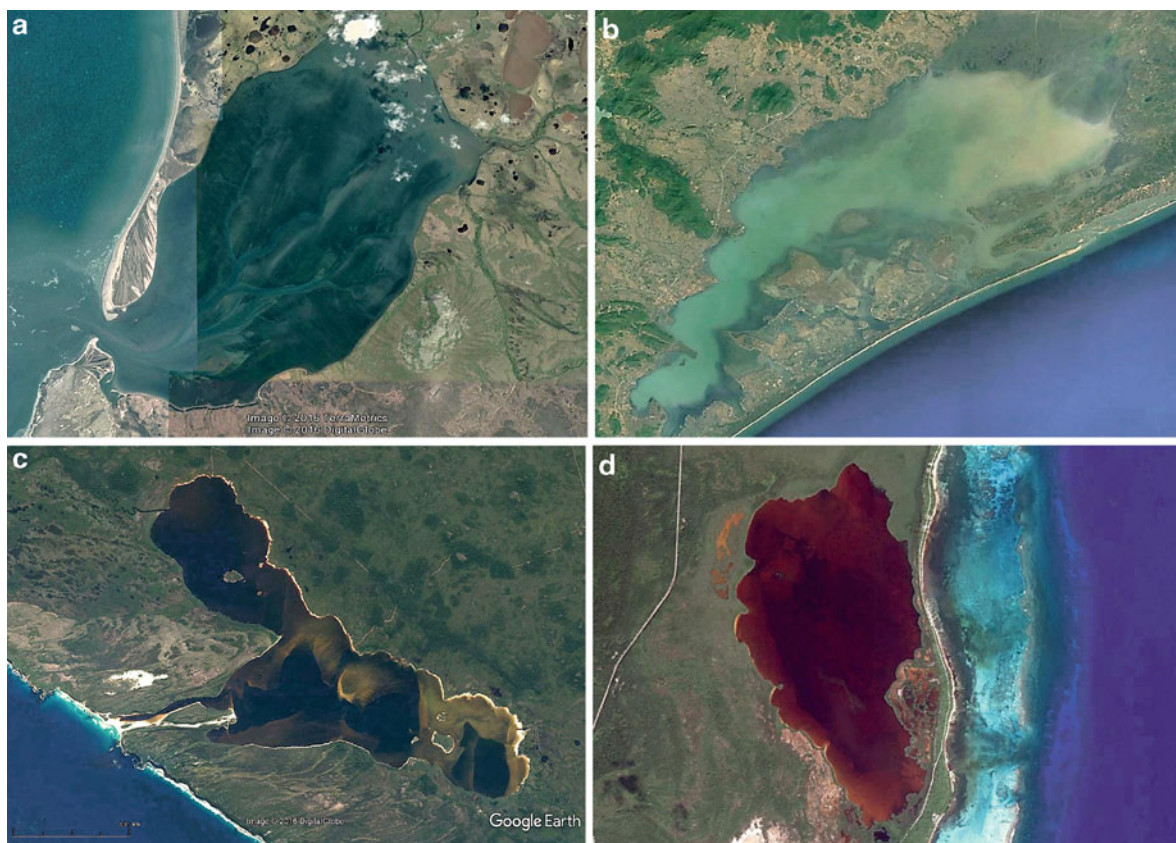


Coastal Lake Systems, Fig. 10 One of the largest lagoons on Earth is 265 km long Lagoa dos Patos in southeastern Brazil. Forelands and spits occur along sedimentary lagoon shorelines, while the northern and southern section exhibit large areas of wetlands. Visible suspension of fine sediments give a lake color with diminishing intensity to the south, because its source is a river mouth in a northern estuary (Credit: Google Earth)

Polymictic lake types are fully mixed day by bay because of daytime climate as in tropical high mountain regions.

This classification in the first order is related to temperature regimes in the lakes. With the exception of *polymictic* lakes, all other types occur as coastal lakes. Their mixing processes, however, are hampered if brackish or saline waters are present. Because of their higher density, these tend to remain stable in the lower water levels.

The nutrient regime of a lake is generally classified into three categories: one concerning water budget, the second concerning sediment budget, and the third concerning nutrient turnover. Water budget implies the main material flows between the hydrosphere and the atmosphere, like input by precipitation and inflow, and output by evaporation and outflow. Sediment budget implies organic as well as inorganic sedimentation with the input of dissolved and particulate materials and their accumulation in the benthos with adsorption and desorption, as well as the exchange of dissolved materials between the water and the sediment by chemical and organic nutrient cycling and turnover. Bioactivity of



Coastal Lake Systems, Fig. 11 (a) Sediment distribution and bottom forms from tidal currents in a lagoon of western Alaska (scene 10 km wide). (b) 65 km long Chilika Lake lagoon in northeastern India during monsoonal river run-off from the NE. (c) Broke Inlet lagoon in SW

Australia, 15 km long, shows light-brown colors from humus-bearing waters of surrounding fresh-water peatlands. (d) Intensive coloring of Laguna Negra in SE Yukatan (Mexico) from mangrove peatlands. Scene is 3.5 km wide (Credits: Google Earth)



Coastal Lake Systems, Fig. 12 Hypersaline lagoons and salt pans: (a) A 3 km wide scene at 47° S in Patagonia (Argentina), (b) 36 km wide scene and 20 km long Lake Bonney (c) from southeastern Victoria (Australia). (d) Three parallel belts of hypersaline lagoons and (mostly)

inactive dune ridges in SW Australia. They represent three phases of Late-Pleistocene and Holocene sea-level high stands (Credits: Google Earth)

organisms in coastal lakes is distinguished in particular by material and energy cycling during production, consumption, and destruction.

An extended ecoregion including a variety of aquatic habitats can be found along the southeastern coastline of Brazil (Dillenburg and Hesp 2009), including Lagoa Mirim and Lagoa dos Patos as the largest coastal barrier-lagoon system in South America. These habitats comprise brackish-salt lagoons, freshwater bodies, swamps, marshes, seasonal freshwater ponds, and estuarine lower river sections. Further to the north, Lagoa de Araruama is hypersaline with salinity up to 5.6‰ (Kjerfve 1994) and also suffers from large amounts of anthropogenic nutrient inputs. Beauclair

et al. (2016) used mollusks in shell middens (*sambaquis*) around this lagoon to identify potential changes in environmental conditions over long time. They were triggered by climatic variations and sea-level movements in older periods, and by strong anthropogenic influence in the younger past. Principally, all factors (physical, chemical, abiotic, biotic) are strongly interrelated in natural environments and control the kind and health of lagoonal ecosystems. As artificial modifications in discharge areas are much stronger since the time of urban growth, industrialization, and agroindustries, they accelerate intensity and frequency of changes to the disadvantage of coastal lake systems.



Coastal Lake Systems, Fig. 13 (a) Chemical industry on the base of evaporates in Ria Lagartos, a 22 km long lagoon at the northern coast of Yucatan (Mexico). (b) Modern aquafarming on shrimps and mussels in

an 8 km wide scene of China at 39° N. (c) Traditional aquafarming on fish and shrimps in a 3 km long lagoon at the south coast of Sumatra, Indonesia (Credits: Google Earth)

Important physical factors are light penetration (which depends on transparency influenced by suspension load and planktonic micro-organisms), temperature, and stratification depending on salinity, radiation, and mixing by wind and currents which potentially differ in the lake zones. An example for nearly perennial salt stratification is Lake Maracaibo (Venezuela), a rather deep (about 60 m) and one of the largest coastal lakes of the world. It is only disturbed in some way during the dry season with stronger winds and relatively weak river inflow (Laval et al. 2005).

Chemical factors are in the first order oxygen content (depending on depth, mixing, and life cycles) and phosphorus as the main determinant for primary production as well as for photosynthesis, plus carbon. Evaporates found in dry pans are the result of enrichment of solutes (mostly inorganic salts).

Biotic factors depend on physical and chemical ones as well as the interrelation of organisms (bacteria, algae, aquatic plants, zooplankton, invertebrates, fish, and other vertebrates) and their trophic relationship and food webs. Algal productivity generally is higher in waters with some salinity. This, however, may change in short periods as do accumulation of

nutrients deriving from soils and weathered rocks in lagoon watersheds.

Summary

The genesis and history of coastal lakes strongly depend on postglacial sea-level variations, either by flooding existing basins, damming coastal indentations by barriers built from sediments transported landwards from the upper shelf during a young transgression, or by longshore drift from nearby cliffs or fluvial sediments. Their history is archived in bottom sediments containing signatures of saltwater intrusions, which influence and limit biodiversity. These lakes (mostly called lagoons) can be classified according to their form and origin, age and relation to sea level, kind of water exchange with the open sea and salinity, or – like all lentic systems – to their water budget, stratification, circulation, or trophic character. Coastal lakes, in general, represent a young and shallow class of water bodies and often and intensively suffer from anthropogenic impacts like pollution and eutrophication from land use and urban areas in

their catchment area. Because of their worldwide distribution and extreme high number, research on coastal lake systems is still in its infancy.

Cross-References

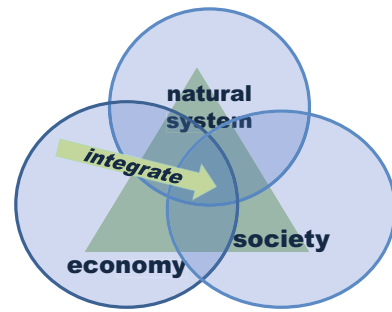
- Aquaculture
- Barrier
- Beach Ridges
- Eustasy and Sea Level
- Estuaries
- Longshore Sediment Transport
- Spits
- Washover Effects
- Wetlands

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Coastal Management Practices, Fig. 1 Essence of ICM: integration of sectors

Coastal Management Practices

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Rationale

Coastal management practices embrace a complex and broad subject, hard to cover in a short paper. This entry starts with general aspects. It continues with a number of management practice examples that give the reader a sense for the scope of coastal management in the twenty-first century. Our examples deal with coastal-plain sedimentary environments. These are and will be amongst the most vulnerable in the future. The main focus selected is not on traditional civil engineering designs, but rather on nature-based designs. In our opinion, they deserve wider attention as well as opportunities.

General Aspects

Coastal Management (CM)

Our coasts are of great ecologic, economic, and sociologic importance (e.g., Martinez et al. 2007). The essence of CM lies in the integration of these sectors (Fig. 1). Coastal management (often also called Integrated Coastal (Zone) Management, ICM or ICZM) is a set of human activities and techniques applied to direct (or keep) the coast, and its adjacent waters and land, in a desired direction or condition. CM integrates across the many sectors that have interests in the coast. From this perspective, it is a form of spatial planning that integrates planning, decision-making, and execution of concrete projects, based on an issue driven but integrated approach (e.g., Misdorp 2011).

Which kind of management practices is relevant for a coastal zone depends on the physical setting, the economic structure, and on the cultural heritage of the coastal

populations. Of special importance is how a particular governance system has evolved and how it works.

CM is important because coasts fulfil a wide range of functions for society. Some of these functions are conflicting with each other. For instance, nature conservation and socio-economic development often are conflicting. Moreover, coasts are under increasing pressure. It is foreseen that within decades, a much larger share of the world's population will be living in coastal areas where there is work, living space, and leisure. The process of CM starts with (future) requirements the local society has. This, in its turn, requires an integrated plan, directed at special issues as a general framework for measures.

Such a plan aims to prevent the development of various individual or site-specific separate plans, that are not related. The plan is made for a particular defined area to be managed. That area is characterized by distinct physical and ecologic properties and forms, (proposed) human activities, the acting governance system, and the way it developed. The CM plan is directed towards issues within this territory. An issue can be safety against flooding, or water pollution, or protection/development of valuable ecosystems (like coral reefs, mangroves, dunes, and beaches), or a combination of issues. CM is a cyclic process (Olsen 2003a). Loops may be followed more than once. Characteristic steps in a loop are: (i) issue identification and assessment, (ii) program preparation, (iii) formal adoption and funding, (iv) implementation, and (v) evaluation and feedback. During the cycle, an adaptive process takes place; there is growing interaction between the natural and the socioeconomic system. That is the reason why CM takes time, considerably more than separate plans and phases of government. This explains why the implementation of CM often is difficult (Olsen 2003b).

Management Practices

The world's coasts present a large diversity. Not only in natural habitats but also in cultural and governance settings, along with a wide variety of management practices. The aim of CM is always to work towards a sustainable development of coasts and their

populations. CM should be site specific and tailored towards the (cultural) history of the site. It means that CM cannot produce a general solution, as if there is just one recipe for all situations, even if the environmental conditions and issues are similar. In fact, the transfer of practices from one country to another (often done with the best intentions) needs great care, in order to be truly appreciated and accepted by the local applicants.

The impacts of climate change are becoming more and more understood. Processes of drastic coastal change (both environment and people) are upcoming. The traditional civil engineering approaches to coastal erosion and shoreline evolution were formulated in a past time of climatic stability. Facing a future time with major sea level rise, climatic instability, and more violent (rain) storms poses great challenges for humans and their management activities. Currently, ca. 60% of the world's population flock along coastlines, and this percentage is expected to rise further over the coming decades, prominently in Asian and African countries, putting large and increasing numbers of coastal inhabitants at risk of river and/or marine flooding (Neumann et al. 2015).

Coasts always are narrow fringes, along large land areas. Thus they can be easily disrupted, both from the sea side as well as from the land side. Because of the many pressures from both sides, the room for development along the coast becomes increasingly confined. The term “coastal squeeze” is coined for this process. It is a slow process, only noticeable after decennia of coastal development (Fig. 2). But while this goes on, characteristic coastal dynamics and biodiversity decrease and the natural “buffer” against flooding and erosion gradually disappears. Giving (back) room to the coast (and sometimes room for river dynamics) is, in some cases, a promising new avenue of practice. It reverses this process and restores coastal ecosystems and dynamics (see, for example, Esteves 2014). Room can be found either by giving up ground that is currently in economic use (going “inwards”) or by “extending the coastline

outwards” (see below, example “[The Netherlands, Southern North Sea Coast: ‘Building with Nature’ in a Sandy Environment](#)”). In any case, physical planning is crucial, especially in the case of an eroding coastline, which is the case in ca. 80% of the world's present coastlines.

Coastal practices often reflect the (history of) culture of coastal population. Back in history, when there were few coastal issues to deal with, there was hardly any management (necessary). Later on, management comprised a kind of “wait and see.” Actions were only taken in response to acute and local problems or hazards that occurred. This reactive attitude can, unfortunately, still be seen in many places in the world. With memory of a hazard (for example, a disastrous flood) fading over time, awareness in the minds of the people does so too, including the urgency to take measures.

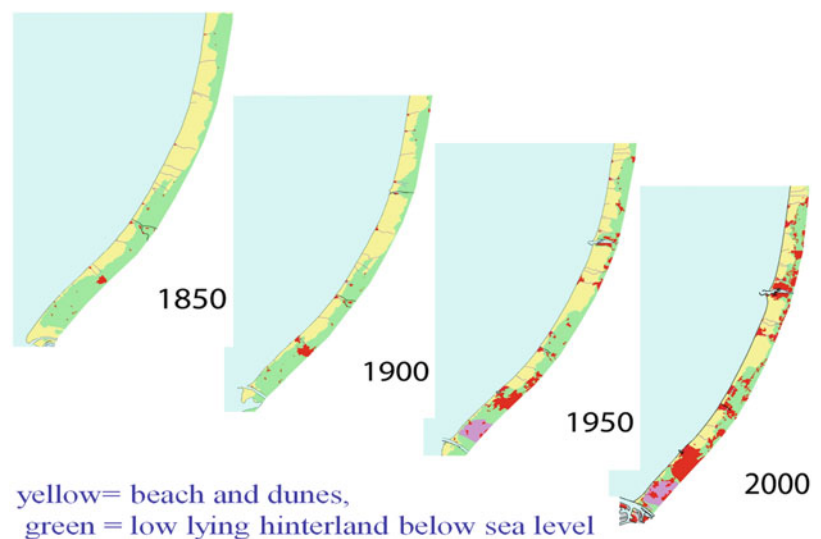
More and more countries, though, tend to change their mind-set and develop proactive attitudes in CM. These countries want to anticipate imminent danger and be ahead of the problems. Proactive practices require measuring, monitoring, modeling, and evaluation, together with public education on the issues and their implications. The longer the time spans of measuring, the better the predictions on future changes, and the better the appropriate management practices. In many places, there is still a lack of sufficient coastal data to support sound management (Example “[São Tomé and Príncipe, Southern Atlantic Ocean: Small Fishermen Communities at Risk](#)”). A proactive attitude, emphasizing the collection and scientific analysis of coastal data, will improve policy design and decision making in the future. This is necessary also with a view towards coastal adaptation to climate change.

Engineered Versus Ecological Design

Civil engineering solutions have been traditionally prominent in coastal management worldwide (see, for example, Pranzini

Coastal Management

Practices, Fig. 2 Coastal squeeze and disruption at the Dutch mainland coast between 1850 and 2000. Note the increase of build-up areas (red) in and around the dunes (yellow). Entrance piers to harbors of Rotterdam (bottom right) and Amsterdam (top right) can be seen



and Williams 2013). Such solutions, for example, those to counteract erosion and prevent flooding from the sea, use hard structures. Examples are groins, dikes, dams, breakwaters, and barriers. An increase in hardening and armoring of shorelines has been the result worldwide. At the same time, the ecological functions and dynamics of the natural system were overlooked or disregarded, and sometimes seriously destroyed. As a result, coastal biodiversity declined (see, for example, Rachel et al. 2016; Dethier et al. 2016).

In recent years, an important shift is seen from hard constructions to nature-based designs, where this seems a possible option. Such designs aim to make use of natural processes, materials, and elements that are present at a given location. This approach has been termed “building with nature” (Waterman 2008, 2010) or ecological engineering. Using this approach means that coastal managers aim to work “with” nature, rather than “against” it. This might be an option in many places with soft coasts, for example, beaches and dunes, but also wetlands, like mangroves and estuaries. In the long term, nature-based management might turn out to be more long lived and cost effective. It certainly increases system adaptability (for example, coastal adaptation to climate change) as well as biodiversity (Temmerman et al. 2013, Nat et al. 2016). Borsje et al. (2010) illustrate how ecological engineering can be used for coastal protection at different scales. From optimizing texture and structure of hard constructions in the intertidal zone on microscale to employing mussel and oyster beds on the meso-scale and using vegetation stands and landscape elements on the macro- and megascales.

The growing momentum for nature-based solutions is also illustrated by two recent publications (World Bank 2017 and Esteves 2014). The World Bank publication attempts to take a step towards standardized guidelines for all nature-based solutions. Five basic principles that came out of recent projects are discussed: (i) take a system-scale perspective; (ii) assess the risks and benefits of the full range of solutions; (iii) use standardized performance evaluations; (iv) integrate with ecosystem conservation and restoration, and (v) apply adaptive management.

Esteves (2014) explains the need for adaptation to nature/natural forces in coastal protection while shifting from hard engineering to nature-based designs. This implies so-called managed realignment. More room is given to habitat creation and coastal ecosystems with their intrinsic dynamics. Examples mainly come from Europe. In reality, one often sees a combination of hard and soft solutions. Along increasingly crowded shorelines, hard responses may be necessary, both politically and economically. A crucial point, however, is that those involved must recognize that their benefits may be short lived as climate change unfolds.

Examples of Management Practices

A limited number of examples of practices and “lessons learned” are selected to illustrate various methods of CM (Fig. 3). They give the reader a feeling for the opportunities for coastal management in the twenty-first century.



Coastal Management Practices, Fig. 3 Location of examples of management practices discussed in this entry. Bangladesh 1, Marshall Islands 2, Mekong Delta Vietnam 3, The Netherlands 4 and 6, São Tomé and Príncipe 5

In general, all the examples illustrate, be it in different contexts, aspects of the features of CM that are highlighted above. But in each example, some features stand out more than others. São Tomé and Príncipe (5) together with Marshall Islands (2) present a small-scale case in a developing cultural setting, scarce on data and coastal experience. Physical and hydrological modelings are tools used. The few field data were combined with desk studies. In the case of Marshall Islands, the emphasis is on a building with nature solution. In ST&P on stakeholder involvement, risk analysis and public acceptance of the results. Although still on a small scale and in early stage, this example shows the principle of risk and benefit assessment (World Bank 2017). Bangladesh (1) and Vietnam (3), as more developed countries, present large-scale examples with dense coastal populations, where building with nature can also be a more sustainable solution. It involves entire coastlines. Stable governance over time and a long-time undertaking are needed. The principle of Integration with ecosystem conservation and restoration (World Bank 2017) applies to these examples. The case of The Netherlands (4, 6) show what can be done in a developed setting with much data and a large coastal experience. Again, building with nature is a component, both at the coast and along rivers. Long-term monitoring, modeling, a proactive (government) approach and learning by doing are CM features here. The example also shows the principle of taking a system-scale perspective (in space as well as in time) (World Bank 2017).

Bangladesh, Northern Bay of Bengal Coast: A Low-Lying, Densely Populated and Vulnerable Delta

Almost the entire country of Bangladesh consists of one of the largest delta floodplains of the world, reaching more than 100 km inland. It presents very dynamic hydrological, morphological, and landscape-ecological characteristics. Three of

the world's largest rivers, the Ganges, the Brahmaputra, and the Meghna, together contribute huge volumes of water and sediment. About 80% of the country is covered with rivers and their floodplains. These rivers are the arteries for life, livelihoods, and economy. Water management in terms of quantity and flood protection is a very complex and challenging issue in Bangladesh. Proper management of water resources is essential for the future development of the country. The Bangladesh Delta Plan (General Economics Division, Bangladesh Planning Commission 2017, draft not for publication) is a guideline for such development. It stresses stable governance and a long time undertaking.

During the monsoon season of just 4 months, hundreds of millions of cubic meters of water flow through the country. More than 90% of this comes from countries upstream in the Himalayan highlands. During the dry season, human interference with the river system is wide spread, due to the requirements of agriculture. The upcoming impacts of climate change are affecting river flows, complicating river management to a large extent. Sea-level rise in combination with land subsidence already cause extensive low-lying areas to flood. In the near future, Bangladesh will face increased flood risks as well as challenges regarding water quality, droughts, and salinization and greater land subsidence.

In some areas, tidal pumping is affecting the downstream tracts of tidal rivers, clogging them with sediment, as polder dikes prevent the spreading of this sediment (Fig. 4). However, exactly this sediment can be used to increase the elevation of parts of the floodplain that have become isolated (by the polder dikes) from sediment supply for half a century by now.

A combined and coordinated management approach on water and sediment could provide a way out. Due to the extent of the floodplain and the prevailing natural conditions (mega-rivers loaded with sediment, monsoon flooding, drought, high tidal levels, effects of cyclones, dense population, the need for

Coastal Management

Practices, Fig. 4 Sediments choking a tidal river in Khulna District. The sediment can be utilized to increase the elevation of the low-lying areas behind the banks. (Photo: Bert van der Valk)



agriculture, and aquaculture production), this management approach needs long-term planning and adequate funding. Stable governance, without frequent changes, is a prime requirement. Proper water and sediment management, from a system-scale perspective, needs to be a key to this process. Temporarily opening up of polder dikes to allow controlled sediment transport into low-lying areas is one of these measures (Tidal River Management: see, for example, the Uttaran plan 2013).

In some areas, short-term measures can be training of shrimp farmers for better use of mangroves for the improvement of the natural conditions of their shrimp ponds, as well as for better yields and quality of shrimps. This involves sediment management. Long-term measures involve more drastic ways of “artificial” (re)distribution of delta sediment using the forces of nature.

Marshall Islands, Pacific Ocean: Survival Techniques for Atolls

The sustainability of human habitation of thousands of low-lying atoll islands is threatened. Threatened by sea-level rise, but more so by salinization of their fresh water resources due to wave-driven over wash (Storlazzi et al. 2018).

The Republic of the Marshall Islands consists of an atoll archipelago located in the central Pacific, stretching approximately 1000 by 1300 km. The archipelago consists of 29 atolls and 5 reef platforms. They are host to approximately 1200 islands with an average elevation of 2 m above mean sea level. Over 3/4 of the population is concentrated on the two atolls of Majuro and Kwajalein. These islands (size 10 resp. 16 km²) have low adaptive capacity. The adaptation costs are high, relative to the gross domestic product (GDP). Consequently, they are very vulnerable to natural hazards and to effects of climate change.

A coastal hazard assessment was made for the islands of Ebeye and Majuro (Giardino et al. 2018). The following

hazards were analyzed as part of a desk study: waves, storm surges, typhoons, and tsunamis. The direct impacts (inundation and coastal erosion) were evaluated, specifically for the island of Ebeye. The analysis was carried out for different return periods and expected climate change scenarios and sea-level rise. The hazard assessment was then used as a basis to carry out a coastal risk assessment, in which information on hazards was combined with information on exposure and vulnerability. The output from the risk assessment allows the determination of weak spots where adaptation measures may be required. A conceptual design and cost-estimate is presented for present or near future measures. The method is used to test the effectiveness of each of these measures, in achieving the predefined objective in terms of risk reduction. This is a very effective strategy for reducing the total investment costs and maximizing the future benefits of the interventions (Giardino et al. 2018).

Technical measures include: (i) dredging coral sand from the reef to raise the island’s level and to reclaim land for the extension of living space; (ii) construction of a rock revetment breakwater around the island to protect against over wash by ocean storm waves, and (iii) provision of construction materials (coral sand and gravel) to the island’s community. These measures will hold for other low-lying islands in general, of course focused on local requirements (Fig. 5).

Coral atolls have a very rich and vulnerable marine biodiversity. It should be safeguarded through special measures, in particular during dredging and reclamation activities. Special monitoring programs are required to monitor environmental changes before, during, and after the work has been completed.

Vietnam, Mekong Delta: Mangroves for Coastal Defense

In part of the Mekong Delta (Southern Vietnam), sedimentation still prevails near the estuarine inlets. But further inland, coastal erosion (due to natural and human-induced causes) is occurring. This erosion is anticipated to increase in the future

Coastal Management

Practices, Fig. 5 Supplementing sediment to Thinhadho Island (Maldives), setting a scenario for other low-lying islands in the world’s seas (Photo courtesy Van Oord, Richard Heerschap)



for several reasons: (i) the increasing number of dams in the Mekong River basin, capturing bed load as well as suspended sediments that will no longer reach the coast, (ii) the increasing human-induced land subsidence in the lower delta due to groundwater extraction for agriculture, aquaculture, and human consumption, and (iii) climate-change induced sea-level rise (IUCN 2011).

The decrease of the suspended solids content of the river and the decline of mangrove health, due to mangrove squeeze (Phan et al. 2015), are of particular impact. Coastal recession contributes to the loss of (mangrove) area, increases salt water intrusion and is detrimental for activities like agriculture and aquaculture and for coastal infrastructure. The best possible measure to mitigate coastal erosion is planting of mangroves in suitable (accretionary) areas, using the sediment entrapment and wave-attenuation properties of mangrove tree stands. Model studies confirm the applicability of this type of measure (Phan et al. 2015). Studies along degraded (and erosional) mangrove coasts have demonstrated the vulnerability of such coasts to almost any kind of human land-use other than nature oriented (Winterwerp et al. 2013).

Spalding et al. (2010) observed the role of mangroves in risk reduction: wind and swell waves are being reduced while passing through mangrove stands. The wider the mangrove belt, the stronger the attenuating effect. Sediment binding and fixation by mangrove roots is also important, especially over time. For an increase in coastal safety, mangrove stands should be recognized as valuable elements of coastal protection. They need to be incorporated in long-term coastal planning by all stakeholders. Phan et al. (2015) found an average critical width of 140 m for the southeastern and eastern Mekong Delta to sustain a healthy mangrove that attenuates wave energy.

New mangrove forestation is not an easy practice. It certainly implies more than simply planting the trees. Mangrove protection and restoration may be more effective, because ecological conditions are dynamic and very subtle. Mangrove stands may better recover in areas where they were growing

previously. This illustrates the principle of integration with ecosystem conservation and restoration (World Bank 2017). The costs for protection are far lower than those for new forestation. Natural mangrove forests are not very rich in species but have a high regeneration capacity. Use this capacity. In erosional areas, additional measures may be needed (keep sediments, plant actively). Green funds may be very helpful in financing combined projects for mixed purposes of mangrove recovery, (economic) livelihoods, and safety of the population.

The Netherlands, Southern North Sea Coast: “Building with Nature” in a Sandy Environment

The Netherlands present interesting recent examples of “building with nature” practices. They illustrate a system-scale perspective (World Bank 2017). They can be found in the south west along the North Sea coast (Fig. 6). Here, the coastal zone is narrow (just 0.2–0.5 km) with beaches and dunes that protect the low-lying hinterland (1–4 m below sea level). The area is very densely populated with glasshouse culture as main land-use. Large towns are The Hague and Rotterdam. This coast has been erosive for centuries. In medieval times, it was part of the estuary of the river Rhine and Meuse with a coastline extending some 1.5 km further seaward than today. Persistent coastal erosion is the reason why the present coast is so narrow. Since at least 200 years, various hard methods of coastal defense have been applied (dikes, groins). Two recent examples, using larger scale sand nourishment, show “building out into the sea” as a measure. They are the Sand Motor (www.dezandmotor.nl) and Spanjaards Dune.

The Sand Motor

The Sand Motor (www.dezandmotor.nl) is an innovative method for coastal protection, taking the system-scale of a whole coastal sector into account. It consists of a huge volume of sand (ca. 20 ml m³) and was constructed in 2011 (Fig. 6, middle). The original form was a hook-shaped peninsula, extending 1 km into the sea and 2 km alongshore.



Coastal Management Practices, Fig. 6 Left: The coast between The Hague and Hook of Holland (see inset for location). This coast is narrow and protects densely occupied hinterland, lying below sea level. Reflective white surfaces represent intensive glasshouse culture (Google

Image). Middle: the Sand Motor, 2013. Right: Spanjaards Dune, 2013. Black lines indicate the original alignment of the coast before construction. (Air photo courtesy Rijkswaterstaat)

Wind, waves, and currents are spreading the sand in a natural way, with the prevailing longshore current (mainly to the coastal sector northwards), thus reinforcing this coastline and creating a dynamic area for nature development and recreation. The Sand Motor has been gradually changing its form, diminishing in size, and will eventually fully merge with the original alignment of the coast. The large-scale Sand Motor nourishment will replenish the local sediment budget for a much longer time than the routineous smaller scale nourishments planned here and to the north. Nourishing a large volume in one go is supposed to be cheaper and better for the benthic ecology than repetitive smaller nourishments. The first evaluation report (Taal et al. 2016) concluded a.o. that the life span of the nourishment will possibly be twice as long as the 20 years that had been estimated by modeling before construction.

The Sand Motor is an integrated solution. It combines coastal defense through maintaining the sand budget on an erosional coast, scientific research, nature development, and recreation.

Spanjaards Dune

This dune area (35 ha) was constructed in front of the existing coast (Fig. 6, right) in 2008/2009, as compensation for the expected loss of valuable existing dune habitats nearby. The loss, mainly because of increasing deposition of traffic emissions, is due to the extension of the Port of Rotterdam with Maasvlakte 2 (van der Meulen et al. 2014). The existing dune habitats in question are protected under the EU Natura 2000 legislation and therefore had to be compensated (van der Meulen et al. 2015). The compensation was realized by constructing a whole entire new dune ecosystem that, over time, will develop into the compensation habitats. This approach illustrates the ecosystem-scale principle (World Bank 2017). The basic foundation of the new area is reclaimed from the shallow North Sea by shore face-, beach-, and dune nourishment (6 million m³). After the construction of this abiotic “ground plan,” natural processes (wind, rainfall, groundwater dynamics, and vegetation succession) have free play to shape the area further, developing it into the dune habitats that need to be compensated. These “target” habitats are calcareous, nutrient-poor dune marsh, and dry dune grassland. The marsh may take 5–10 years to complete but the grassland takes 15–20 years. Careful management, monitoring, and evaluation are needed to assess whether the desired development is indeed taking place and to adjust when necessary. Important questions for the monitoring are: (i) Is the eventual damage to existing dunes going to be as it was predicted? (cf. Royal Haskoning Nederland 2006); (ii) Is the quality of the new compensation area going to be enough to compensate for this damage?

Because of the EU obligation to compensate, this project also has a political dimension. It is a unique example of a one-

to-one field experiment (learning by doing) where politicians, engineers, ecologists, and managers work and learn together.

“Learning by doing” in building with nature is important because more experience is needed on ecosystem behavior (both physical as well as biological) to be able to apply this knowledge.

The compensation aspect is worth considering worldwide where ecosystem damage occurs as a result of (large) engineering projects, such as harbors.

São Tomé and Príncipe, Southern Atlantic Ocean: Small Fishermen Communities at Risk

The islands of São Tomé and Príncipe present an example of coastal management practices at a less developing and isolated location with hardly any available data. And also of how to involve the local population so that results are truly appreciated and accepted by the receivers.

The Democratic Republic of São Tomé and Príncipe (ST&P) comprises two main islands and four islets located at the equator, in the Gulf of Guinea, 350 km off the west coast of Africa. It is one of Africa’s smallest nations with a total area of 1000 km² and population of 166,000. Angolares villages are small fisherman’s communities built at the coast, often at very isolated locations (Fig. 7). They have since long suffered from coastal inundation and erosion. The flooding increased in recent years, possibly due to increased storm intensity and shifts in the rainy seasons that have begun to overlap with the storm season. Climate impacts have probably been further increased by offshore and beach sand mining. In view of its isolation, small size, and limited capacity, ST&P was given the least developed country (LCD) and small island developing state (SIDS) status. The island’s fishermen communities are considered highly vulnerable to climate change and sea-level rise. Between 2010 and 2012, possible adaptation options were studied. The focus was at the fishermen villages of Malanza, Praia Sundy, Ribera Alfonso, and Santa Catharina (https://www.researchgate.net/publication/266216911_Sao_Tome_and_Principe_Adaptation_to_Climate_Change-Climate_variability_and_hydrogeomorphological_study, UNESCO-IHE and Deltares 2011).

At first, a study design with stakeholder plan was made. Potential hazards were assessed by making a climate variability scenario analysis and a coastal hydrology, oceanography, and geomorphology survey and subsequent analysis. The hazard assessment formed the basis for the risk assessment, where information on hazards is combined with information on exposure and vulnerability. This resulted in a coastal communities adaptation plan and recommendations for National Government. Also, a social assessment was carried out and risk maps were made (Fig. 8).

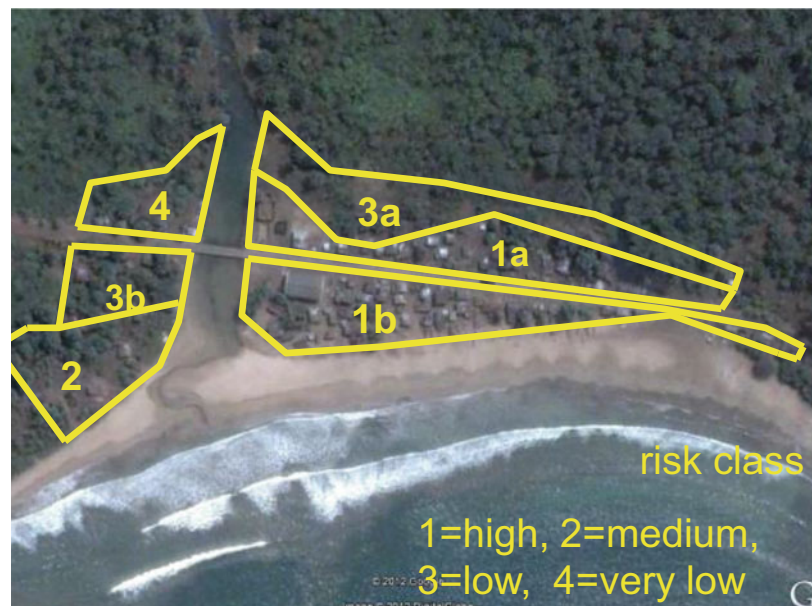
The technical analysis was done with the help of simple bathymetric profiles, surveyed with an Echo sounder (from the shoreline to –20 m) and a digital



Coastal Management Practices, Fig. 7 Left: View on Praia Sundy from the boat. Right: Fishermen houses on the beach and on higher ground

Coastal Management

Practices, Fig. 8 Risk map of Malanza fishermen's village (2012). Risk factors are flooding from the sea, from the river and coastal erosion. Flooded areas were qualitatively defined with help of terrain elevation, river discharge data, and 1/50–100 years storm surge event with 80 cm sea level rise by the end of the century. Risk class 1 involves flooded by the river and the sea and with a large number of houses. Class 4 involves only flooded by the sea at sea level rise of 80 cm by 2100 with only a few houses



elevation model of the coastal zone, using “storm data” (<http://strm.csi.cgiar.org>) for sediment transport and storm erosion calculations. Climate data were obtained from open sources and modified to fit the local situation as much as possible.

During the participatory meetings, the risk zones in the villages were shown on aerial photos and people were asked to comment on it. Three-dimensional mapping was used as a tool to interact with the communities and to discuss the various risks and potential adaptation measures. Positive and negative sides of the measures were written down. In addition, people were asked in which way they could contribute to implementation of the measure. It was advised to

proceed with the implementation in steps, in order to evaluate the efficiency of the measurements. Long-term monitoring and a stable government are needed to improve CM at the islands.

Room for the River: Ecological Engineering in the Lower Reaches of the River Rhine

In 1993, and again in 1995, two near-flooding events occurred in the lower Rhine flood plain. As a precaution, many people were evacuated. This urged an integrated plan for the area to prevent flooding in the immediate and more distant future. Climate models predict that by 2100, the peak winter discharge of the Rhine will increase from 16,000 to



Coastal Management Practices, Fig. 9 Overview of the area through which peak river flows will be diverted, Nijmegen, Eastern Netherlands. The area to be flooded is designed to conduct floodwaters for some

time until river levels are back to normal. In the meantime, the temporarily flooded areas function as wetlands. (Photo: <http://www.ruimtevoordewaal.nl> (Johan Roerink/Aeropicture.nl))

18,000 m³/s (Delta Commission 2008). Appropriate legislation for drafting such a plan was quickly initiated and passed by the government. Measures were implemented shortly after. The program was called “Room for the River” (see [RfR brochure](#); Sijmons et al. 2017). Major storage areas for controlled flooding were allocated (and in some cases “depolderized,” which is the reverse of reclaiming land from a wetland) to temporarily accommodate excess water during the peak of river floods. By design, these areas will start to function when the water level in the river exceeds a given threshold. The storage areas have the shape of real river courses, parallel to the normal river bed (Fig. 9). Or they are situated in the lower reaches of the river, where previously reclaimed polders were handed back to the river to become flooded-storage area during high water. The added advantage of these measures is that deposited river aggregates (sand, clay, gravel) become available for ready use during the construction period. The areas-to-be-flooded are wetland nature areas when not needed (van Staveren et al. 2014). This is another example of “building with nature” in a delta.

The Future of Coastal Management

Coasts will become more and more engineered with the construction of cities, harbors, and other infrastructure, pushing away room for nature. More emphasis will be needed to put on compensation of damage to nature and on co-development with nature. The concept of Green Port development is an example. Older harbor areas can be subjected to Brown-field development but equally focused on using “building with nature” type of measures.

The major challenge for the future lies in proactive thinking and governance. This involves integration of solutions, collecting and exchange of sufficient data to evaluate and use for prediction, awareness, and stakeholder involvement. Physical planning should reflect all ideas about coastal management. When it comes to practical measures in the field, one should be aware when, where, and how traditional engineering designs can be combined with new nature-based designs. More experience with the latter in different coastal environments is needed, because, at an increasing rate, coasts will be used and at risk (Scheffers et al. 2012).

“Man uses nature for all kinds of purposes. That is quite all right as long as we realize that we depend on her” (Saeijs 2008, p. 113).

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Coastal Modeling

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Definition

Coastal Modeling includes physical modeling and mathematical modeling. Physical modeling is to build a replica of a coastal domain, use scale-down ocean and atmospheric forcing to drive the system and study physical processes in the system. Mathematical modeling is to use numerical methods to solve the mathematical equations for conservation of mass and momentum and to simulate waves, hydrodynamics, sediment transport, and morphology change in the coastal zone. Various coastal models provide coastal engineers and scientists an efficient tool for understanding coastal processes and for designing and managing coastal inlets, beaches, navigation channels, ports, and coastal structures.

Introduction

Coastal physical processes include waves, storms, tide, current, and sediment transport. These processes can cause shoreline and beach changes, coastal inundation, and damages to coastal structures and properties.

Coastal waves are generated in open oceans due to winds, storms, or seismic activities, which carry energy and propagate to coastal areas. As approaching coasts, water becomes shallow, waves interact with sea bed, current, shoreline, and coastal structures. Due to shoaling and refraction, wave height and direction will change. When encountering coastal structures, stronger processes like wave diffraction, reflection, breaking, run-up, and overtopping will occur.

Ocean tide induces the regular rise and fall of sea surface and is driven by the combined gravitational forces of Moon, Sun, and Earth. Tidal periods may range from a few hours to more than 10 years, while the main tidal components include diurnal and semi-diurnal periods. Associated with tidal changes in water surface elevation are the horizontal water movement and tidal current. The regular water ups and downs in coastal regions correspond to the convergence and divergence of tidal current, including the interaction between tidal current and shoreline geometry.

Ocean current is driven by two classes of natural forces (Pond and Pickard 1983). Primary forces include gravitational, wind, atmospheric pressure, and seismic forces, and secondary forces are Coriolis and friction forces. On the contrary, coastal circulation are mainly driven by tide, wind,

and density gradient. Depending upon the nature and the relative importance of forces, and geographical locations, coastal currents can be vertically uniform or varying with depth.

Storms generated in tropical oceans or in the polar regions move past coastal regions, induce water level rises and falls, and enhance coastal and estuarine circulation. Extreme water levels due to storms are storm surges, which, combined with storm winds and peak tides, may result in extensive coastal flooding, land recession, and massive loss of human lives and properties (Nummedal et al. 1980; Halverson and Rabenhorst 2013; Li et al. 2013a).

Coastal structures are built to protect coastal residents, shorefront properties, and infrastructure, to prevent shoreline erosion, and to improve environmental conditions for coastal community and ecosystems. These structures include seawalls, jetties, breakwaters, groins, weirs, culverts, and tidal gates.

The combination of physical forces acts on coastal systems and interacts with complex coastlines and coastal structures, which cause coastal sediment movement. Sediment transport processes in coastal zones include longshore and cross-shore sand migration due to wave action, navigation channel scouring and infilling, changes of beach face, coastal shoaling, and sandbar formation.

Accurate modeling of coastal physical processes is required in engineering studies for coastal inlets, shore protection, nearshore morphology evolution, harbor design and modification, navigation channel maintenance, and navigation reliability.

Physical modeling has been a powerful means used for coastal studies. With the fast development of computer technology and its cost efficiency, mathematical modeling has become more popular in the last couple of decades. In this write-up, coastal numerical modeling will be described in different aspects of coastal studies.

Wave Modeling

Waves possess energy and propagate towards shoreline. As approaching surf zones, waves will break and dissipate energy. In coastal regions, released wave energy is partially transferred into the driving force for coastal current and coastal sediment movement. Therefore, waves are the most important factor in studying coastal processes. Because of the complexity in wave generation and nearshore wave transformation, understanding and predicting wave action always present a challenge.

Among various approaches conducted for wave analysis, numerical wave modeling has been widely used in recent years. Two types of wave models have been developed and applied for solving scientific and engineering problems in

deep or coastal waters. The first type is phase-averaging wave models and the other is phase-resolving wave models.

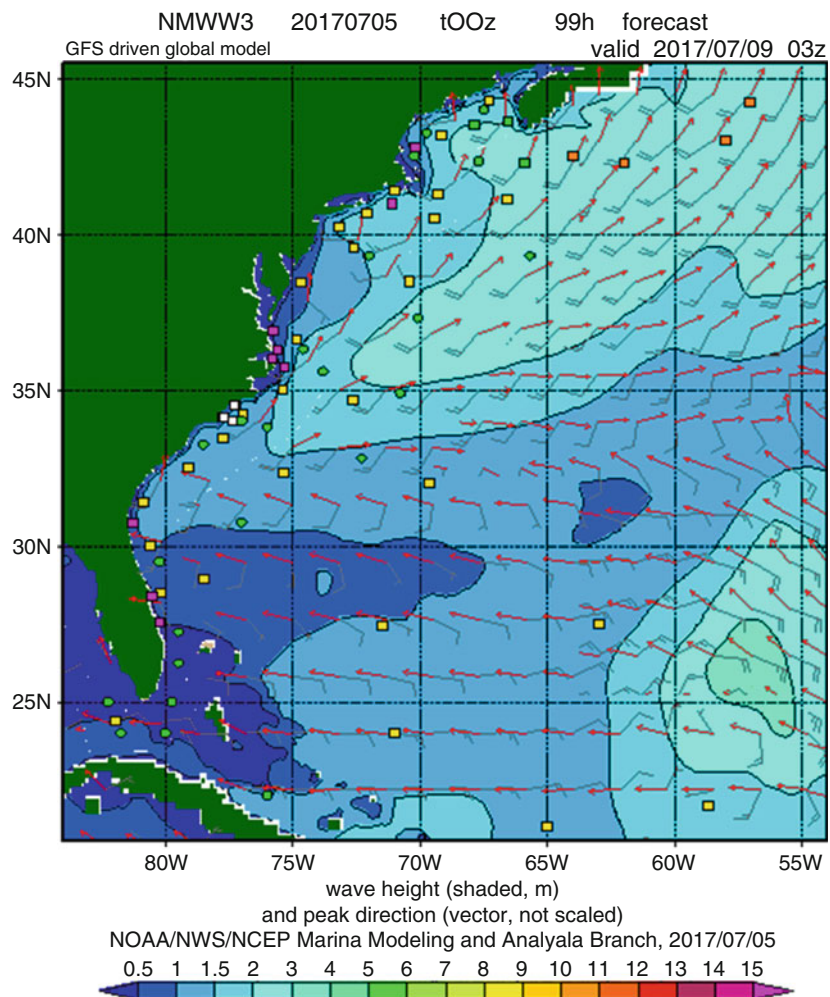
Phase-averaging models calculate energy spectra of waves and are used to simulate waves in a large regional to global scale or a small coastal scale. Popular phase-averaging models include WAM (WAMDI Group 1988), WAVEWATCH III (Tolman 1991), SWAN (Booij et al. 1999), STWAVE (Smith et al. 2001), and CMS-Wave (Lin et al. 2008). The WAM and WAVEWATCH III models were developed to simulate wave generation for large scale, deep water applications. The National Centers for Environmental Prediction of NOAA operates the WAVEWATCH III model and provides wave analysis and forecast over global oceans and US coasts. Figure 1 shows a one-hour forecast snapshot of significant wave height on the East Coast of US. Three- to four-meter waves are observed in the open ocean and wave height is reduced to less than 0.5 m close to coastal areas. Waves propagate primarily in the wind direction, indicating that wind-waves are generated in open ocean areas.

While the SWAN and STWAVE models are used for wave generation in deep water and wave transformation in shallow water, the CMS-Wave model focuses on coastal wave processes, including diffraction, refraction, reflection, wave breaking and dissipation mechanisms, wave-current interaction, and wave generation and growth. In nearshore zones, wave energy dissipation and transfer due to bottom friction and depth-induced breaking is a great contributor to coastal currents, coastal sediment transport, and coastal structure impact. Four depth-limited breaking formula were examined and implemented in the CMS-Wave model (Lin et al. 2011). Figure 2 shows significant wave height distributions associated with different wave breaking criteria.

Phase-averaging models calculate wave generation and transformation by assuming uniformly distributed wave phases. In order to calculate wave propagation and wave processes more accurately, phase-resolving models focuses on detailed phase information. Because of the requirements for high temporal and spatial resolution,

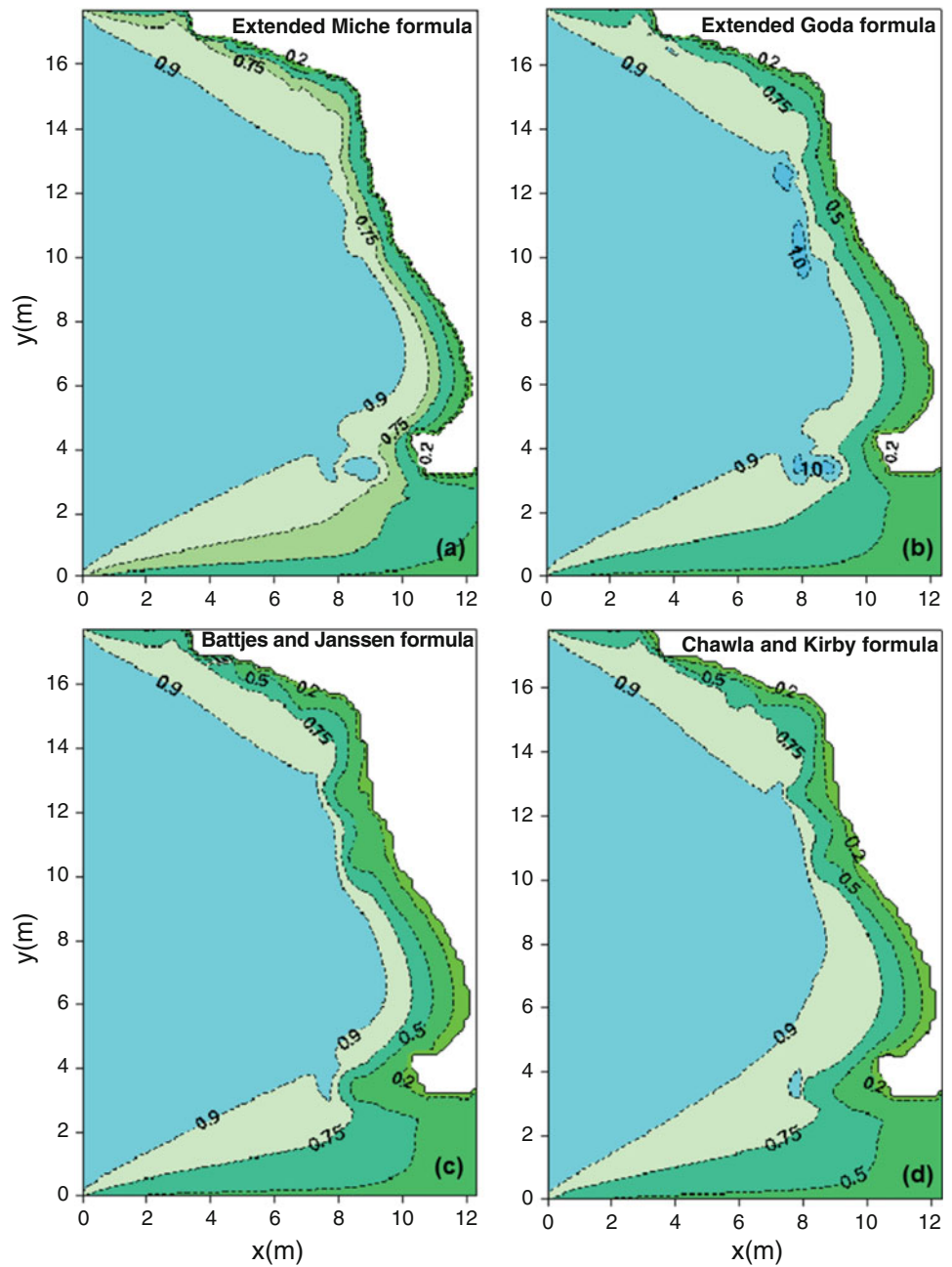
Coastal Modeling, Fig. 1

WAVEWATCH III model product: one-hour forecast snapshot of significant wave height on the East Coast of US (http://polar.ncep.noaa.gov/waves/viewer.shtml?multi_1-latest-hs-US_eastcoast, accessed 11 July, 2017)



Coastal Modeling, Fig. 2

Calculated wave height distributions using (a) the extended Miche formula (Battjes 1972), (b) the extended Goda formula (Sakai et al. 1989), (c) the Battjes and Janssen formula (Battjes and Janssen 1978), and (d) the Chawla and Kirby formula (Chawla and Kirby 2002)



phase-resolving models are limited to the calculations of short-term processes in small-scale coastal domains. Such models include the mild-slope model (Berkhoff et al. 1982) and the Boussinesq model (Nwogu 1993). Demirebilek and Nwogu (2007) carried out Boussinesq model simulations to investigate wave energy transformation at a site along Guam's southeast coast near Ipan. The calculated wave propagation and corresponding significant wave height over the Ipan reef are shown in Fig. 3. The Boussinesq model simulations clearly display that wave shoaling and breaking processes and asymmetric bores over the reef.

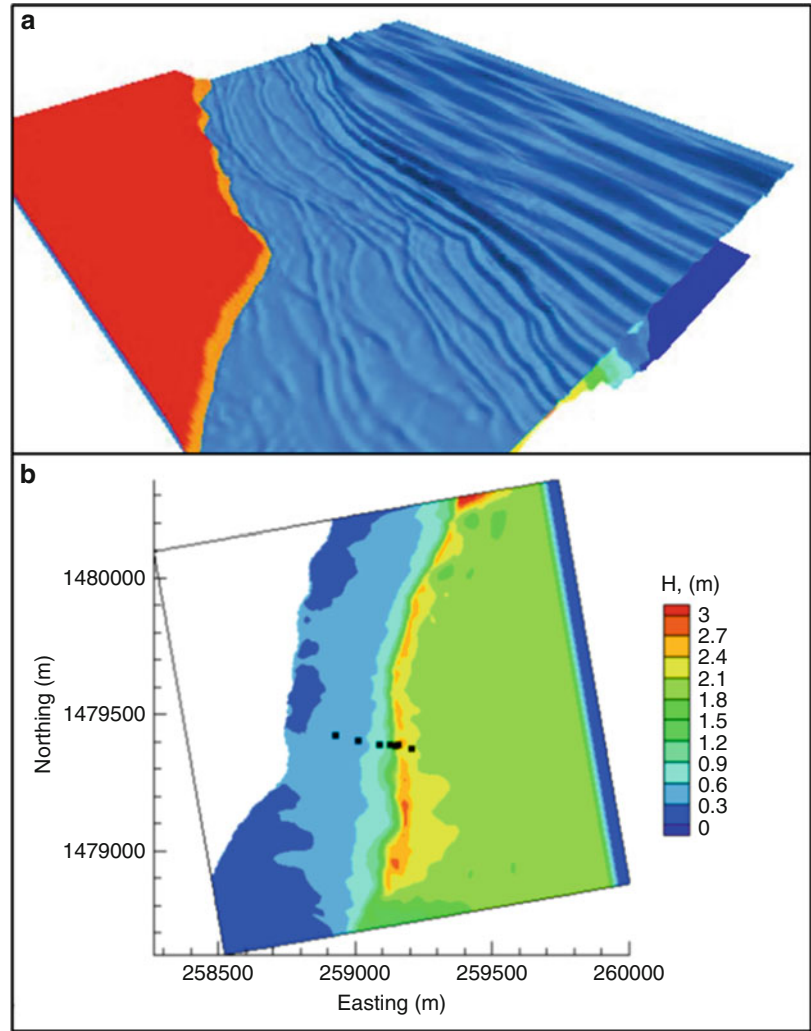
Circulation Modeling

Shallow coastal areas experience strong tidal mixing, receive high wind energy input, and are very often categorized as vertical well-mixed water bodies. Therefore, two dimensional modeling is a common practice for coastal applications, in which the shallow water equations are derived by depth-averaged equations of conservation of mass and conservation of momentum in the Cartesian coordinate system as follows:

$$\frac{\partial D}{\partial t} + \frac{\partial Du}{\partial x} + \frac{\partial Dv}{\partial y} = 0 \quad (1)$$

Coastal Modeling, Fig. 3

(a) Wave propagation and
(b) significant wave height over
Ipan reef, Guam during a storm
event (Demirbilek and Nwogu
2007)



$$\begin{aligned} \frac{\partial Du}{\partial t} + \frac{\partial Du^2}{\partial x} + \frac{\partial Du v}{\partial y} + \frac{1}{2} g \frac{\partial D^2}{\partial x} - f D v \\ = \frac{\partial}{\partial x} \left(A_x \frac{\partial Du}{\partial x} \right) + \frac{\partial}{\partial y} \left(A_y \frac{\partial Du}{\partial y} \right) + (\tau_{sx} - \tau_{bx}) \end{aligned} \quad (2)$$

$$\begin{aligned} \frac{\partial Dv}{\partial t} + \frac{\partial Duv}{\partial x} + \frac{\partial Dv^2}{\partial y} + \frac{1}{2} g \frac{\partial D^2}{\partial y} + f D u \\ = \frac{\partial}{\partial x} \left(A_x \frac{\partial Dv}{\partial x} \right) + \frac{\partial}{\partial y} \left(A_y \frac{\partial Dv}{\partial y} \right) + (\tau_{sy} - \tau_{by}), \end{aligned} \quad (3)$$

where x and y are the horizontal coordinates; u and v are the horizontal velocity components in the x - and y -directions, respectively; D is the total water depth and is the sum of the still water depth, h , and the free surface deviation from the still water surface, η ; g is the gravitational acceleration; f is the Coriolis coefficient; A_x and A_y are the diffusion coefficients in the x - and y -directions, respectively; τ_{sx} and τ_{sy} are

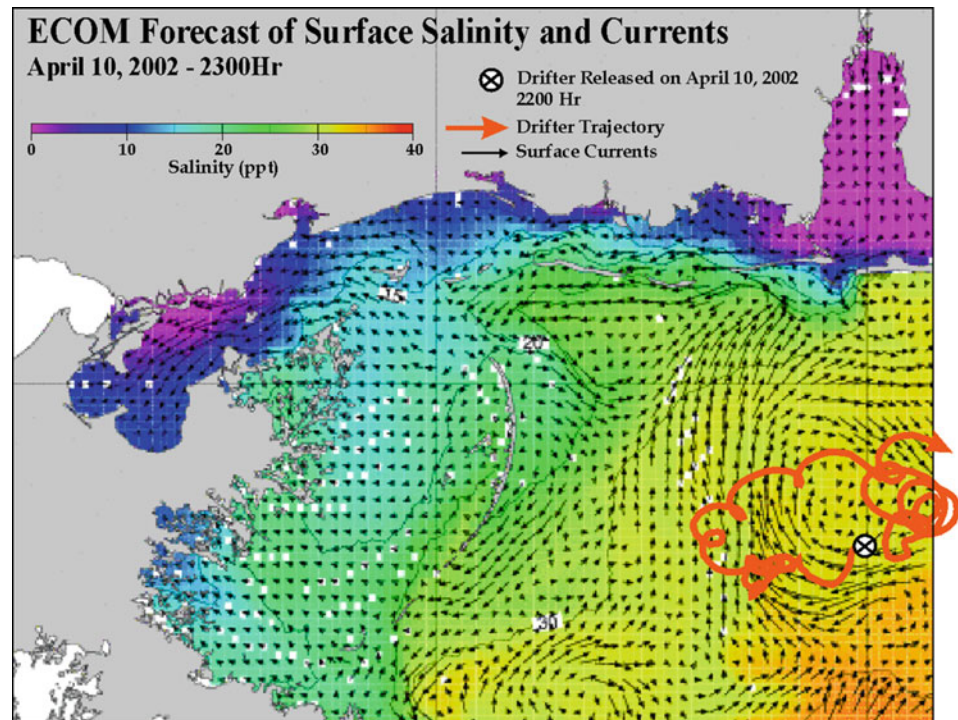
the surface stress in the x - and y -directions, respectively; and τ_{bx} and τ_{by} are the bottom stress in the x - and y -directions, respectively.

Numerical solutions of water surface elevation and depth-averaged current are obtained in Eqs. (1), (2), and (3). Princeton Ocean Model (POM) is an early ocean circulation model (Blumberg and Mellor 1987), based on which the Estuarine Coastal Ocean Model (ECOM) was developed (Blumberg et al. 1992). Successful applications of POM and ECOM include many coastal and estuarine sites in US and around the world. ECOM is the core model of an operational forecast modeling system for the Mississippi Sound/Bight (Blumberg et al. 2002). Figure 4 shows the model forecasted surface current and salinity and the drifter trajectories. Considerable spatial variability in current and a persistent ocean eddy were exhibited in the figure.

The coastal modeling system (CMS) was developed at coastal and hydraulics laboratory of U.S. Army Engineer Research and Development Center. The CMS is an integrated suite of numerical models for simulating water surface

Coastal Modeling, Fig. 4

ECOM forecasted coastal current and salinity fields (Blumberg et al. 2002)



elevation, current, waves, sediment transport, and morphology change in coastal and inlet applications. This modeling system includes representation of relevant nearshore processes for practical applications of navigation channel performance and sediment management at coastal inlets and adjacent beaches. The CMS consists of a hydrodynamic and sediment transport model (CMS-Flow) and the previously mentioned spectral wave transformation model (CMS-Wave). CMS-Flow is a two-dimensional (2D) finite-volume model that solves Eqs. (1), (2), and (3) on a telescoping grid (Sánchez et al. 2011; Wu et al. 2011). Typical applications of the CMS include analyses of past and future navigation channel performance; wave, current, and wave-current interaction in channels and in the vicinity of navigation structures; and sediment management issues around coastal inlets and adjacent beaches.

The CMS was set up to conduct numerical modeling investigation adjacent to Merrimack Inlet, Newburyport, and nearshore in the vicinity of Salisbury Beach and Plum Island, Massachusetts (Li et al. 2014). Concerns at the site include beach erosion, shoreline retreat on Plum Island down-drift of and within the inlet, and reduced navigability of the inlet. Figure 5 shows the CMS configuration and calculated residual current around the inlet system.

Sediment Transport Modeling

Combined action of waves and current in nearshore areas causes significant shoreline change and sediment migration.

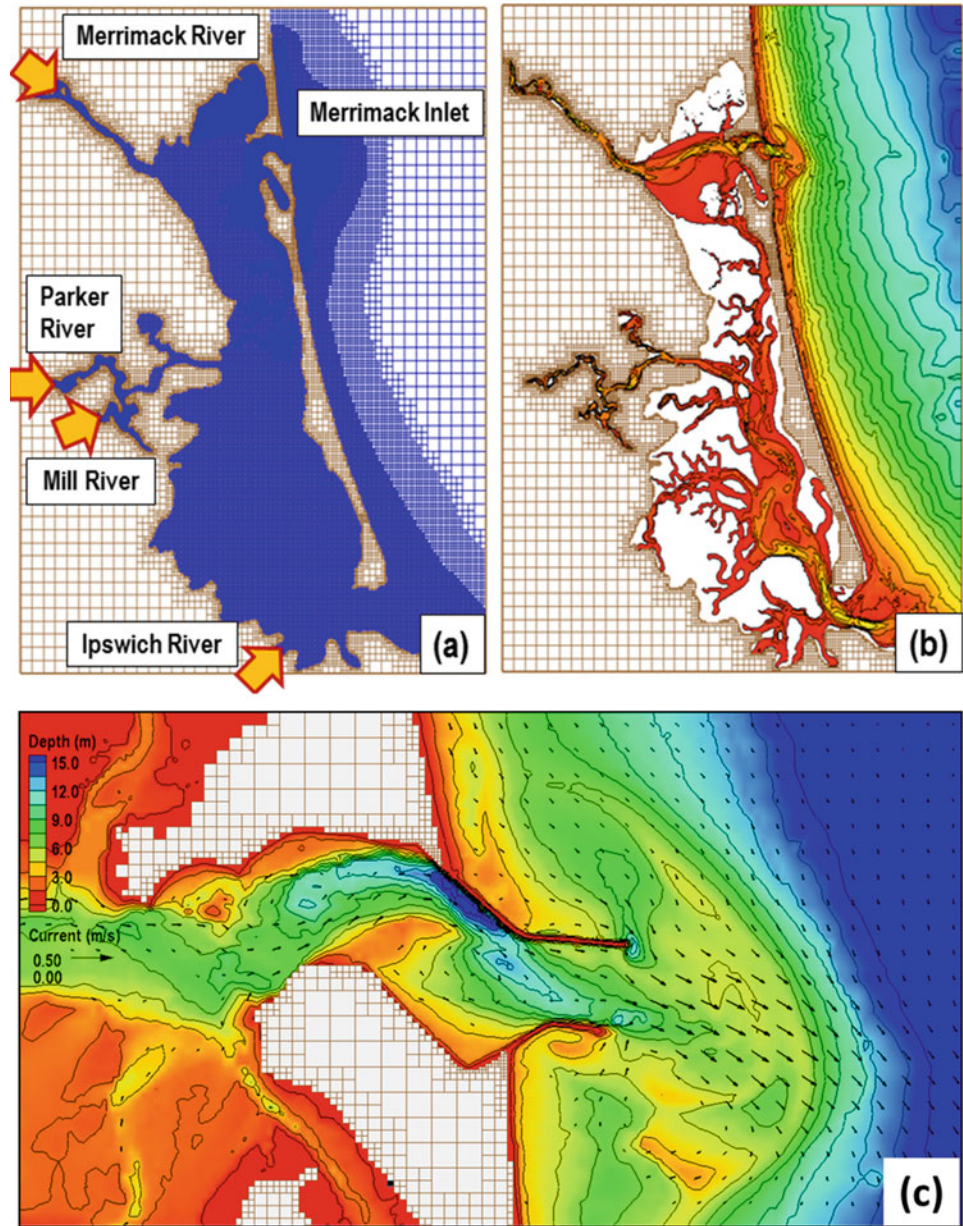
Simulating and understanding coastal sediment transport and morphodynamic processes is an important component in coastal modeling practice.

Sediment transport models simulate bedload, suspended load, or total-load (combined bedload and suspended load) transport of noncohesive and cohesive sediments. For muddy sediment bed (cohesive material), suspended load transport is the process to be considered; for sand sediment bed (noncohesive material), both bedload and suspended load transport processes have to be taken into account. As displayed below, the depth-averaged advection-diffusion equation is solved for sediment concentration and sediment flux in water column.

$$\begin{aligned} \frac{\partial DC}{\partial t} + \frac{\partial CDu}{\partial x} + \frac{\partial CDv}{\partial y} = & \frac{\partial}{\partial x} \left(K_x D \frac{\partial C}{\partial x} \right) \\ & + \frac{\partial}{\partial y} \left(K_y D \frac{\partial C}{\partial y} \right) \\ & + (E_b - D_b) \end{aligned} \quad (4)$$

where C is the depth-averaged sediment concentration; K_x and K_y are the diffusion coefficients in the x - and y -directions, respectively; and E_b and D_b are the erosion and deposition rates, respectively. The bedload transport in noncohesive sediment or in mixed cohesive and noncohesive sediment calculations have been obtained by researchers using empirical formulas in laboratory experiments (van Rijn 1984, 1993; Watanabe 1987; Soulsby 1997; Camenen and Larson 2005).

Coastal Modeling, Fig. 5 (a) The CMS telescoping grid and (b) bathymetry at Plum Island Sound and Merrimack Inlet. (c) Calculated residual current for January 2011 (Li et al. 2014)



In sediment transport models, the advection-diffusion equation and various transport formulas are used to calculate total load sediment transport. Once transport rates are determined, depth change (bed elevation change) can be calculated by the sediment continuity equation,

$$(1 - p) \frac{\partial z_b}{\partial t} = \frac{\partial q_{t,x}}{\partial x} + \frac{\partial q_{t,y}}{\partial y} \quad (5)$$

where p is the porosity of sediment; z_b is the bed elevation; and $q_{t,x}$ and $q_{t,y}$ are the total load sediment transport rates in the x - and y -directions, respectively.

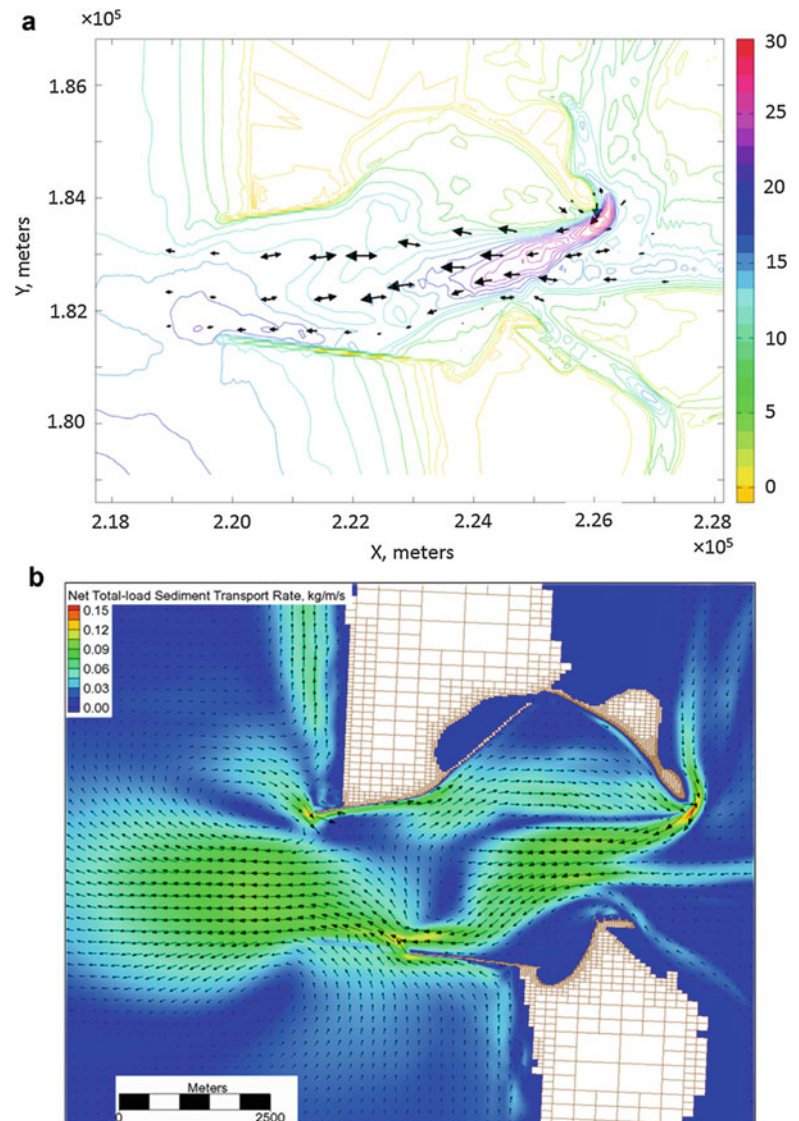
CMS-Flow in the CMS has three noncohesive sediment transport models which differ mainly in the

assumption of the local equilibrium transport for the bed and suspended loads (Buttolph et al. 2006; Sánchez and Wu 2011). CMS-Flow can simulate any number of sediment size fractions, the interactions between size fractions, bed sorting and layering, and morphology change. The sediment transport model also includes processes such as avalanching, nonerodible surfaces (hard bottom), and bed slope effects.

Grays Harbor estuary is located on the southern coast of Washington, USA. A large natural inlet on the west connects the harbor to the Pacific Ocean through a deep draft navigation channel. In recent years, the elongation of Damon Point, a spit on the northeast portion of the harbor entrance, has caused the channel thalweg to migrate. The purposes of the

Coastal Modeling, Fig. 6

Calculated sediment transport pathways and net total-load sediment transport rates (Li et al. 2013b)



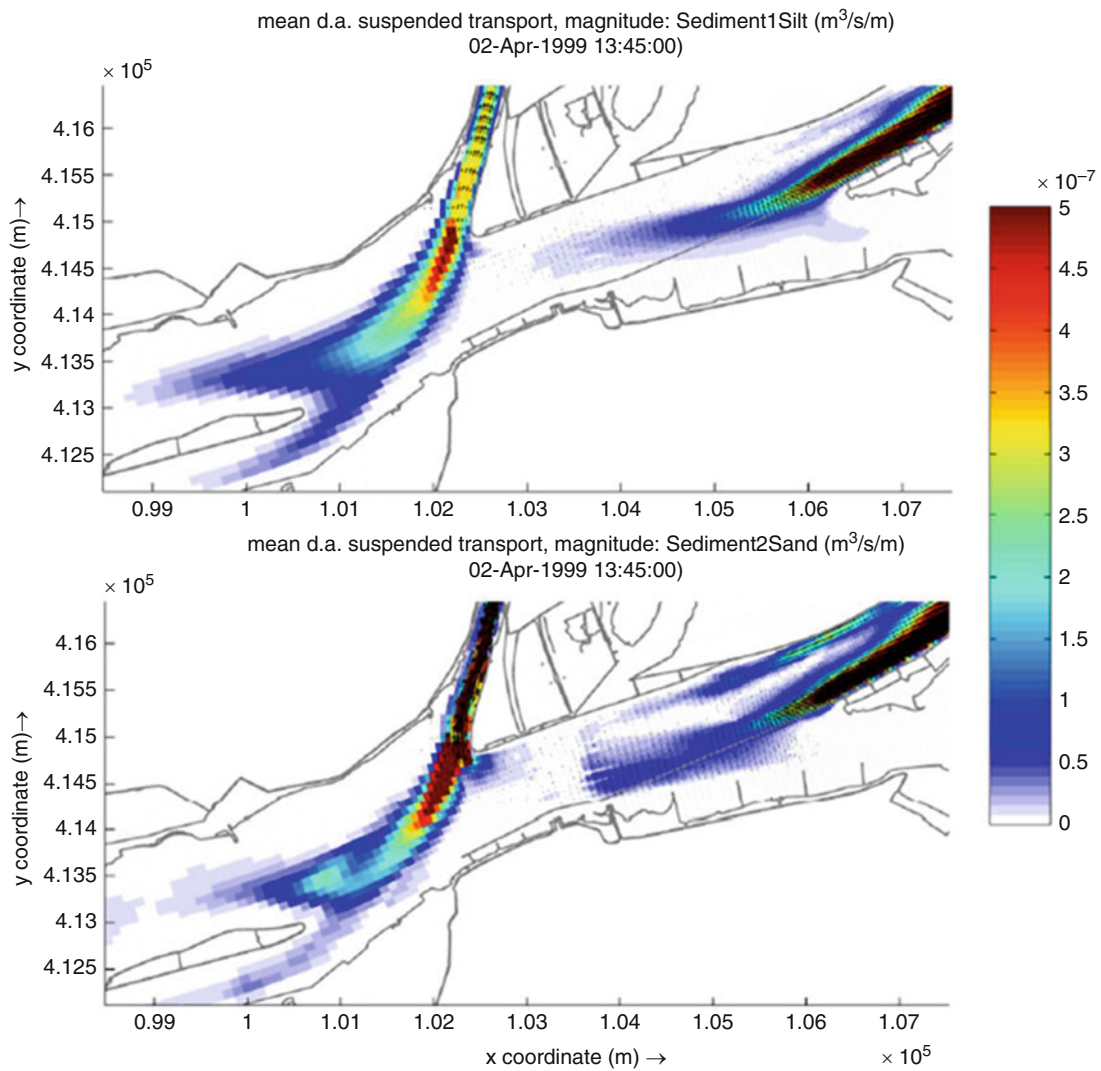
CMS development for the Grays Harbor estuary are to evaluate potential shoaling in the navigation channel and its possible link with the growth of the Damon Point spit and to investigate physical consequences of the Damon Spit encroachment and associated channel migration. Figure 6 shows the conceptual sediment transport pathways and the calculated net total-load sediment transport rates around the inlet system (Li et al. 2013b).

Delft3D-SED is the sediment transport and morphology module in Delft3D modeling suite (DELFT HYDRAULICS 2006). For noncohesive sediment (sand), both bedload and suspended load transport are calculated; for cohesive sediment (mud), suspended load transport is calculated with the incorporation of flocculation and consolidation processes.

Sloff et al. (2012) developed a numerical morphological model at the mouth of Rhine and Meuse Rivers, which is an

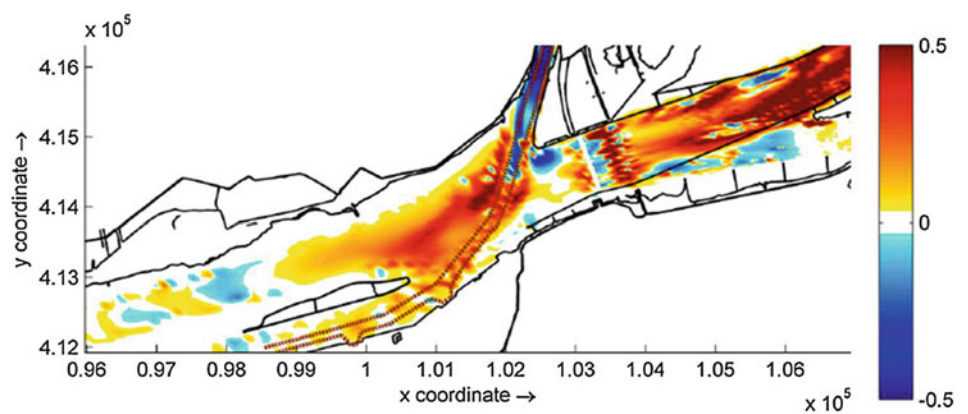
extension of the existing Delft3D model of the Rhine branches in the North and the West of the Netherlands. The modeling study is to improve the understanding of hydrodynamics and morphodynamics in this estuarine system and help develop local sediment management strategies in response to flow pattern changes and erosion variations due to the construction of the Deltaworks.

The morphological model was implemented by specifying fractions of both cohesive and noncohesive sediments from silt to coarse sand. Figure 7 shows the calculated suspended sediment transport rates of silt and sand material. As found in the study, Sloff et al. (2012) indicated that suspended load transport is the dominant process comparing with bedload transport in the estuarine system. Therefore, bed elevation change shown in Figure 8 is well corresponding to the distribution of sediment transport rates in Fig. 7.



Coastal Modeling, Fig. 7 Cohesive and noncohesive suspended sediment transport (Sloff et al. 2012)

Coastal Modeling, Fig. 8 Bed elevation change. The warm color represents accretion and the cold color erosion (Sloff et al. 2012)



Shoreline Change Modeling

Coastal engineering practice and regional sediment management require understanding of beach evolution and shoreline change, and numerical modeling of dominant sediment transport and morphologic evolution processes in coastal littoral zones.

GENESIS, GENeralized model for SIMulating Shoreline change, is such a numerical model to calculate longshore sand transport and simulate shoreline change in response to wave action (Hanson and Kraus 1989). The equation of the model is governed by conservation of sediment mass under assumptions of constant beach profile shape, impact of longshore sediment transport, and existence of a long-term trend in shoreline change et al. and is described as below:

$$\frac{\partial y}{\partial t} + \frac{1}{(D_B + D_C)} \left(\frac{\partial Q}{\partial x} - q \right) = 0 \quad (6)$$

where x is the alongshore coordinate; y is the cross shore coordinate, representing the shoreline position; D_B and D_C are the newly formed berm height and the closure depth, respectively, relative to a vertical datum; Q is the alongshore sand transport rate parallel to the x -coordinate; and q is the

sand input in the y -direction. The solution of shoreline position obtained from Eq. (6) is interpreted in Fig. 9.

As a one-dimensional shoreline change model, GENESIS was designed to simulate shoreline/beach evolution over a period from months to several years. With wave input as the major driving forcing, the model can be used to assist coastal engineering projects, such as coastal structure design and evaluations, beach nourishment, and berm placement. While GENESIS focuses on shoreline change studies in a local spatial scale (project dimension) and a relatively short temporal scale of several year, the Cascade shoreline change model was developed to simulate alongshore sediment transport and morphologic evolution of shoreline in a large regional scale and a long-term temporal scale. The Cascade model can cover a coastal domain up to hundreds of kilometers and simulation periods can be extended to more than a century to address long-term influence of natural processes on coastal changes (Larson et al. 2006).

Based on theoretical framework of GENESIS and capabilities of Cascade in calculating large spatial scale longshore transport, morphological change, and interaction between coastal processes and coastal structures (inlets) over a long period of time, GENCADE was developed (Frey et al. 2012).

Coastal Modeling, Fig. 9

(a) Side view and (b) Plan view of solution procedure of Eq. (6) (Frey et al. 2012).

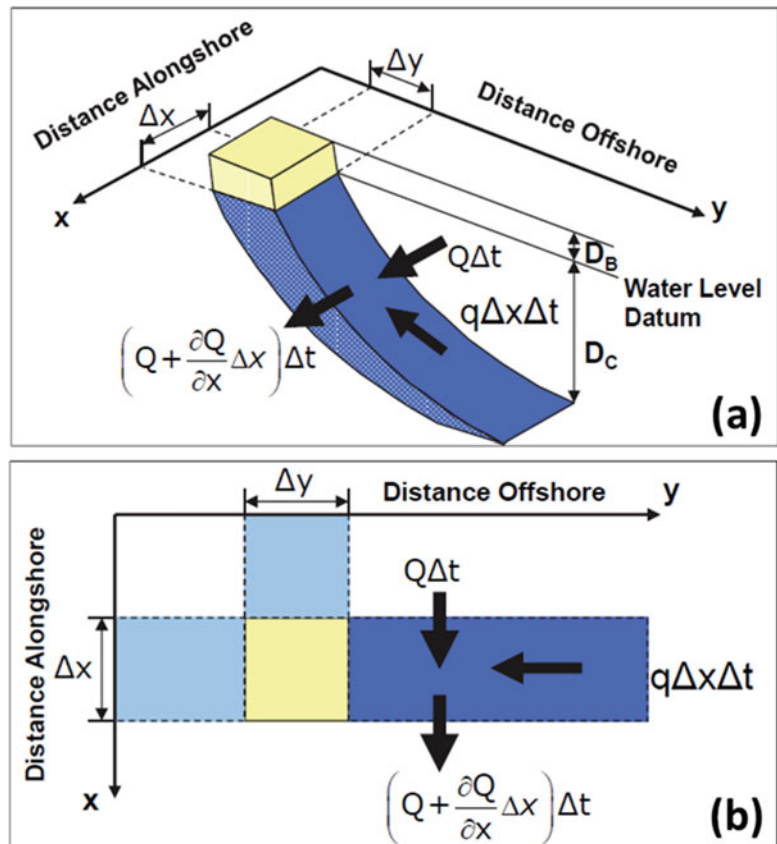


Figure 10 shows a test case with a single inlet along a straight shoreline. Under constant wave forcing with the assumption of equilibrium morphological elements, the simulation results display the shoreline response of updrift sediment accumulation and downdrift erosion.

Coastal change and shoreline evolution have been greatly influenced by coastal flooding due to storm conditions and could be impacted by future sea level rise. Driven by waves and water levels, SBEACH, the Storm Induced BEACH model, is a storm-scale numerical model for simulating beach profile change, the formation and movement of longshore bars, troughs, and berms, and dunes (Larson and Kraus 1989). By neglecting longshore transport components, beach profile change is obtained from the calculation of cross-shore transport rate and the mass conservation equation along cross-shore transects.

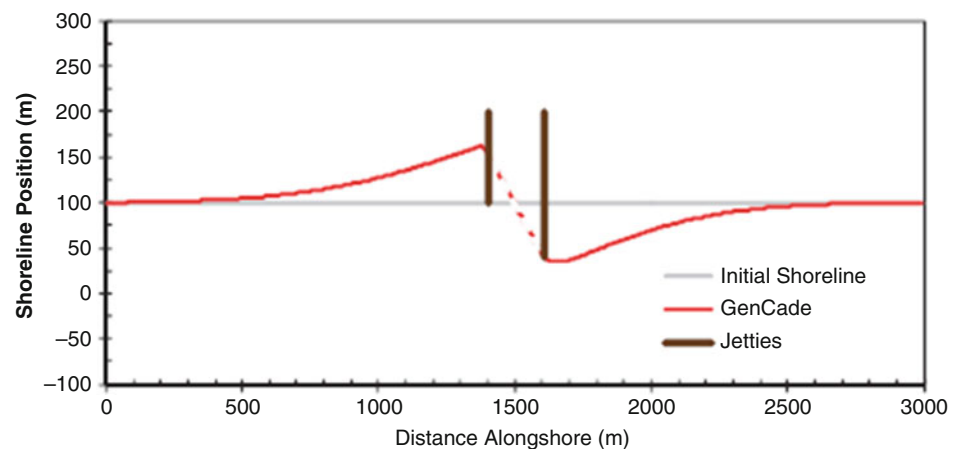
SBEACH can be applied to determine the fate of proposed beach fill alternatives under storm conditions and to compare the performance of different beach fill cross-sectional designs, and to predict volumetric overtopping rates for

catastrophic events in open coasts. King et al. (2011) applied the model to evaluate beach profile and shoreline change in response to storm surges along Wallops Island, Virginia. As shown in Fig. 11, the hurricane induced surge had caused berm and dune erosion.

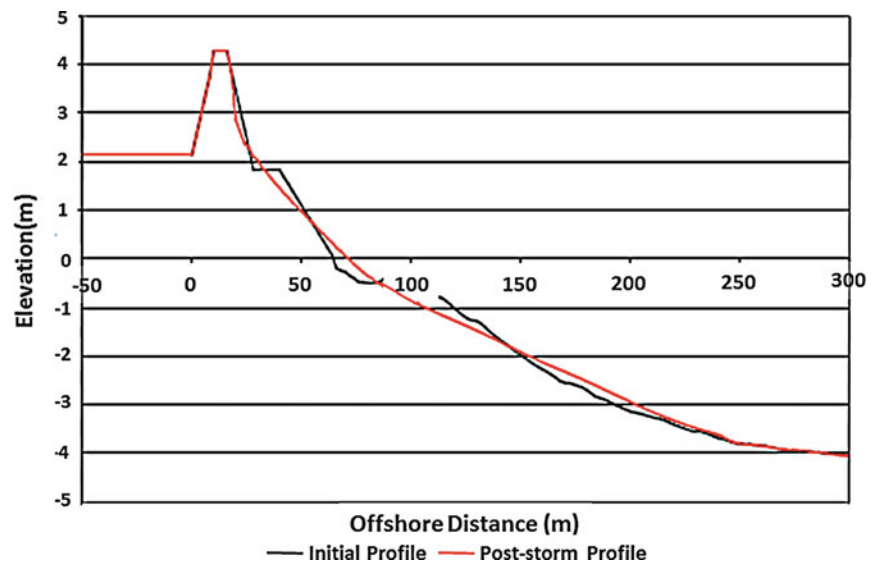
As a one-dimensional numerical model, GENCADE simulates shoreline change and SBEACH calculates cross-shore profile change under various wave forcing. Both models can predict beach profile change, berm, and dune migration in different spatial and temporal scales but with limitations of the assumptions of equilibrium sediment transport and breaking of short-period waves. XBeach is a two-dimensional time dependent swash and surf zone model (Roelvink et al. 2009). This model solves coupled short wave energy, flow, sediment transport, and bed level change equations for cross-shore and longshore hydrodynamic and morphodynamic processes on a spatial scale of kilometers and a temporal scale of storms.

Coastal physical processes included in the model are wave transformation, wave setup and overwashing, longshore, and

Coastal Modeling, Fig. 10
Calculated shoreline position
(Frey et al. 2012)



Coastal Modeling, Fig. 11
Beach profile response for a
hurricane (King et al. 2011)



cross-shore hydrodynamics, bed load and suspended load sediment transport, dune erosion, etc. McCall et al. (2010) set up the XBeach model to calculate morphological change and coastal inundation due to Hurricane Ivan at Santa Rosa Island, Florida. The simulation period is 36 h. Figure 12 shows the comparison between the calculated morphological changes at the end of the simulation and the LIDAR measurements. The figure clearly indicated that the storm surge and wave action resulted in the dune inundation and sediment

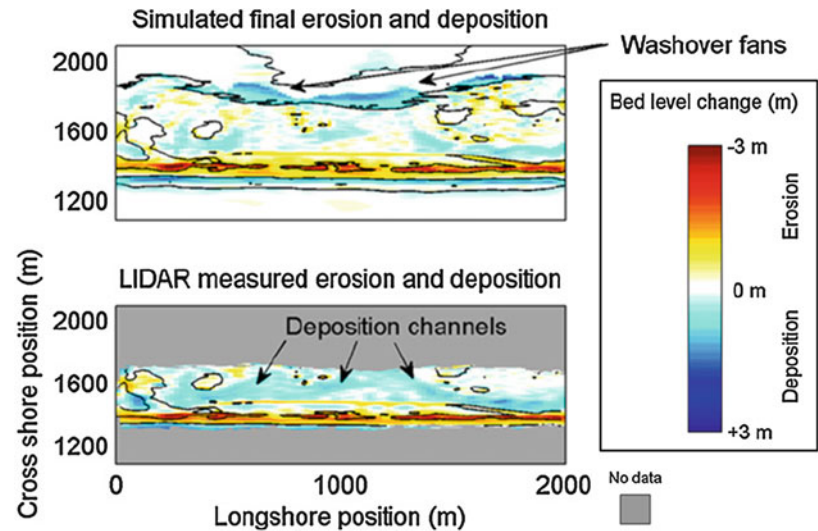
transport over the barrier islands, and two washover fans were created in the back barrier bay.

Coastal Structure Modeling

Various coastal engineering structures are built to protect shoreline, harbors, and navigation waterways, which include rubble mounds, weirs, culverts, tidal gates, et al. (Fig. 13). Since these

Coastal Modeling, Fig. 12

Simulated erosion and deposition at the end of the simulation (36 h). Measured erosion and deposition (McCall et al. 2010)



Coastal Modeling, Fig. 13

Implementation of coastal structures in coastal modeling



structures are a significant component in controlling hydrodynamics and sediment transport in the coastal zone, it is important for a coastal numerical model to properly represent them and to simulate their effects. In coastal models, structure effects are usually treated as the obstructions in flow fields and are represented as energy loss in the governing equations. The term has the form of a friction term with an empirical dimensionless coefficient. Depending on the type of structures, the energy loss can be evaluated by a linear or a nonlinear form.

Rubble mound structures are typically used as seawalls, groins, breakwaters, and jetties. The design of rubble mound structures often consists of a core of small- to medium-size rock or riprap covered with larger rock or riprap to armor against wave energy. In coastal modeling, it is reasonable to assume that the flow through these structures is negligible, since the flow is controlled through laminar flow through small pores, and they are therefore often represented as solid structures, impermeable to both flow and sediment transport. However, some designs may implement larger diameter riprap in the core, and the resulting structure can be sufficiently porous to allow flow across the structure. The flow resistance in these structures can consist of both laminar and turbulent components; the pore space can provide significant storage for sediment and act as a sediment trap. Fine, grain-sized sediments may pass through the structure (Li et al. 2015).

The Forchheimer (1901) equation describes unidirectional flow in porous media, in which the laminar (linear) and turbulent (nonlinear) components of flow resistance are represented. The numerical implementation of the formulation in the CMS is to incorporate the resistance equation into the governing equations, representing the drag force of the rubble mounds. Both the linear and nonlinear coefficients in the equation are determined using data from studies by Sidiropoulou et al. (2007).

Weirs are common coastal structures typically used in weir jetties or in wetlands to control discharges, provide flood control, act as salinity barriers, and optimally distribute freshwater to manage salinity regimes and sedimentation rates and deposition patterns. Weirs are also used in inland streams to increase navigation channel depth, collect water runoff from agricultural fields, control sedimentation, and stabilize channel morphology.

Two approaches are developed to implement weir structures in the CMS model. The first approach is based on the standard weir equation for either sharp-crested or broad-crested weirs, with a foundation in Bernoulli's equation. The second approach is to add local resistance force over a weir structure in the momentum equations.

In coastal applications of culverts, the culverts often connect open water bodies of similar water surface elevation with two coupled locations, for instance, to enhance flushing or provide controlled flow through levees or causeways. Therefore, the implementation of culverts assumes subcritical flow

conditions. In the CMS, the implementation of culverts is based on equations developed by Bodhaine (1982).

Gates along with levees, dikes, and roadways are often used in tidal areas to control the water flow. They can increase flushing, improve water quality and also provide passage for fish and other aquatic animals to coastal lagoons and bays. Some gates may be self-regulating (controlled by water level) and some are manually operated according to certain schedules to meet the requirements of flood control, navigation, water quality management, and aquatic ecosystem rehabilitation. Figures 4, 5, 6 and 7 show some typical weirs, tide gates, and culverts used in coastal engineering and other areas.

Similar to the treatment of weirs, two approaches are used to simulate gate functions in the CMS model. The first approach determines the underflow through gates using the orifice flow equation. The second approach adds the local resistance force due to gates in the momentum equations.

Summary

Coastal modeling covers a wide variety of coastal, estuarine, and riverine processes that include wave dynamics, coastal circulation, storm surge and coastal flooding, tidal hydraulics and flushing, sediment transport, and shoreline change and beach erosion. Proper representation of coastal physical processes in physical or numerical models will greatly assist coastal engineering design and improve coastal management.

Presently, as mentioned early, many coastal models are available for applications. However, not one model fits every single situation and it depends on the discretion of coastal modelers (engineers and scientists) to select the right model to apply for a specific site to solve a specific problem.

Cross-References

- ▶ [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
- ▶ [Beach Erosion](#)
- ▶ [Beach Processes](#)
- ▶ [Coastal Changes, Gradual](#)
- ▶ [Coastal Changes, Rapid](#)
- ▶ [Coastal Flood Hazard Mapping](#)
- ▶ [Coastal Management Practices](#)
- ▶ [Cohesive Sediment Transport](#)
- ▶ [Cross-Shore Sediment Transport](#)
- ▶ [Engineering Applications of Coastal Geomorphology](#)
- ▶ [Estuaries](#)
- ▶ [Estuaries: Anthropogenic Impacts](#)
- ▶ [Longshore Sediment Transport](#)
- ▶ [Modes and Patterns of Shoreline Change](#)
- ▶ [Nearshore Sediment Transport Measurement](#)
- ▶ [Storm Surge](#)

- Tides
- Waves
- Wave-Dominated Coasts

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Coastal Risk

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Synonyms

Possibility of loss

Definition

Coastal Risk is defined as the ecological, social, economical, functional, and cultural damages possibly caused to coastal areas due to their geographical location. The risk is conditioned by two components: kind of menace and vulnerability. For that reason, coastal risk means that the coastal area is exposed to the possibility of loss. To know the risks a specific coastal area is exposed to, the possible consequences of a natural or anthropic event impact, and the probabilities of some kind of danger or threat to that area, have to be previously assessed.

Introduction

Large urbanized areas have been founded at coastal zones due to the growing number of population in different countries.

According to Shankardass (Spokesperson of the United Nations Program for Human Settlements (Habitat)), “Coastal cities grow at a rate of 20% faster than others; having also 10–15% greater number of population density.” Coastal cities and small insular States concentrate a population rate of about 643 million inhabitants, representing the 10% of the world population; being, therefore, all those people in great risk of the consequences of the climate change (GN 2007; IPCC 2014).

Vulnerabilities and risks of coastal population have lately increased, not only because of the changes in climate conditions, as the continuous sea-level rise, but also because of the great number of natural catastrophes and the creation of nonplanned urban areas (Cardona 2001; Milanés et al. 2017a). Analysis of the risks and vulnerabilities of urban coastal zones might give a solution to such complex problems.

In the present chapter, a critical review about various methodological frameworks used for establishing coastal risks is carried out. The following five aspects were taken into account for the analysis: (1) kinds of actual risks; (2) disciplinary approaches about risks; (3) relationship among coastal risk–menace–vulnerability; (4) current tools for coastal risk assessment; and (5) coastal risk management.

Kinds of Risks

The risk concept and associated terminology have varied not only throughout time, but also as a consequence of the different disciplinary approaches used (Cardona 2001). Up to the present, different kinds of risks have been studied from the theoretical and scientific points of view, such as: *urban risk* (Allen et al. 2017; Herslund et al. 2016; Milanés and Pacheco 2011; Bull-Kamanga et al. 2003); *urban environmental risk* (Lavell 2009a); *nuclear risk* (EPM 2010; GE 2009); *economical risk* (Mascareñas 2008); *marine or fishing risk* (Rezaee et al. 2016; Shahrabi and Pelot 2007; Wu et al. 2005); *geological risk* (Milanés et al. 2017a); *earthquake risk* (Guasch et al. 2008; Soloviev 1978); *coastal risk* (Milanés et al. 2017b; Zanuttigh et al. 2015; Botero and Milanés 2015; Milanés 2014; Bárcena et al. 2012; Gornitz et al. 1994), among others.

Some professionals working on themes related to disasters, use the term risk with only one concept, without making reference to the particularities involved in that term. For effective risk management, it is very important to know the differences among the various kinds of risks and corresponding variables.

Coastal risk has also been defined as “Expected damages in case of one or various kinds of menaces, no matter the origin, having the probabilities of affecting at a specific time and place, one or several vulnerable elements located within the coastal zone boundaries” (Milanés 2014; Milanés et al.

2015a). The variable called “geographical space” and the specific coastal zone characteristics are comprised in the previous concept, which needs essential knowledge about the territorial coastal boundaries object of analysis to be carried out into practice.

Following the previously mentioned concept, some nations take into account a narrow zone between land and sea for establishing coastal boundaries (MMA 2001; GORC 2000; Steer et al. 1997). However, some authors and coastal polices follow arbitrary distances to establish coastal boundaries (Henríquez et al. 2011; Abogado and Mendez 2003; GORC 2000). A new method, on the other hand, was designed for that task following differentiated integral and integrated criteria with the aim of proposing new programs for integrated management, planning and ordering of coastal zones (Milanés et al. 2017b). In all cases, the vulnerable elements to coastal risk, such as: housing, social buildings, infrastructure networks, economical goods, ecosystems, biodiversity, among others, have to be preventively assessed against different menaces to decrease damages, manage the risks, and increase resilience to the threat.

Approaches and Evolution of Coastal Risk

At present, great advances have been achieved regarding risks; however, any of the approaches followed by different disciplines have proposed a consistent and coherent theory about the disaster approach (Cardona 2001). The present climate-change effects, gradual sea-level rise, saline intrusion, erosion processes, and loss of coastal ecosystems have led to new studies, methodologies, and projects for analyzing coastal risk following an integrated, multi-transdisciplinary approach. Table 1 shows different disciplinary approaches about risk, as well as their characteristics and evolution.

Relationship Among Coastal Risk–Menace–Vulnerability

The concept of risk continues being misunderstood by some professionals, since the terms threat or danger are also equally used when making reference to risk (Cardona 2001). The term risk really means the possibility of people or systems of undergoing some kind of damage when exposed to a natural or anthropic phenomenon at a specific time period and place. The risk is a consequence of the mutual conditioning between threat and vulnerability and depends not only on social factors, but also on natural factors.

Identification of threats constitutes an essential component in studies about coastal risks because of the importance of knowing the probabilities of occurrence of a natural or

anthropic event of a specific magnitude at a particular geographical zone and time (Milanés 2014). Threats might occur at recurrent time periods, meaning the intervals of time of an event repetition. Events of great magnitudes generally have long recurrent time periods. Threats to people or territories depend on the following factors: event magnitude, frequency of occurrence, time duration, extent, speed, and particular space to be affected (Botero and Milanés 2015).

In some countries, e.g., in Cuba, the term danger is used as a synonym of threat (CITMA 2005; Cardona 2003; Lavell 2009a; DCC 2012). Some authors, like Milanés (2014), however, use those two terms differently, defining danger as any kind of natural or anthropic event affecting a particular territorial area, and considering the danger to become a threat when the event has the possibility of affecting one or more vulnerable elements, e.g., people, economic, and social goods, among others. Furthermore, the threat becomes a risk, coastal or not, when the previously mentioned factors coincide at a specific place causing social, environmental, and economic damages. A risk becomes coastal risk when the damages to people and territories are caused by the interaction between land and sea.

According to the origin and possibility of prediction, threats to insular or coastal regions are classified as follows: natural menaces, technological threats, and sanitary threats (Botero and Milanés 2015). For a threat to become a risk, it is necessary a scenario of vulnerability, meaning the presence of elements with probabilities of undergoing some kind of damage caused by a natural or anthropic event.

Vulnerability stands out at the urban coastal context as a relevant risk element. The economic, political, social, and environmental realities of every country have to be taken into account when studying coastal risks. Analysis of vulnerabilities is the starting point for getting knowledge about risks (Batista Matos 2006). The term vulnerability and its corresponding categories have been analyzed following various approaches, in correspondence with the institutions responsible of their assessment. For that reason, specific studies about the following categories or vulnerabilities have been done: urban vulnerability (Rosales 2004; Lavell 2009a; Milanés and Pacheco 2011; Herslund et al. 2016) and seismic vulnerability (Guasch et al. 2008; Milanés et al. 2015b), carrying this last category the analysis of social and physical vulnerabilities (Cardona 2003; Batista Matos 2006; CITMA 2005; Milanés 2014; Milanés et al. 2015b). Structural and Nonstructural vulnerabilities are included in studies about physical vulnerability, also called functional vulnerability by some specialists (CITMA 2005; Milanés et al. 2015b). For other authors, functional vulnerability is the same as institutional or political vulnerability (Botero and Milanés 2015; Milanés et al. 2017a). However, all the categories of vulnerabilities previously mentioned differ in behavior and results depending on the variables and contexts used for the analysis.

Coastal Risk, Table 1 Different approaches about Risk and Evolution to the Integrated Coastal Risk Approach (Source: Modified from Cardona 2001)

Disciplinary approaches	Characteristics and evolution
Natural sciences	Study of severe natural phenomena, (e.g., floods, landslides, earthquakes, hurricanes, tsunamis, volcanic eruptions)
	Object of study by different specialists, like geologists, seismologists, geophysicists, meteorologists, hydrologists, biologists, among others
	Reduced view, or partial approach about risk, considering the threat only “part” of the risk (Cardona 2001)
	Analysis of geodynamic and hydrometeorological menaces, considered remarkable contributions to risk assessment
	Great technological advances, although prediction of events, like earthquakes and tsunamis, is still very difficult
	Use of advanced technological systems, like sensors, for alerting and warning about intense events (e.g., hurricanes, floods, tsunamis, and volcanic eruptions) at real time (Cardona 2001)
Applied Sciences	Use in engineering and “hard” sciences
	Special attention to the physical properties of systems with possibilities of being damaged, or technologically destroyed by external phenomena (Cardona 2001)
	Concept of vulnerability incorporated from the disaster and modeling perspectives; use of probabilistic methods (UNDRO 1979)
	Recognition of the importance of analyzing threats and elements exposed to damage, including vulnerabilities, e.g., people and systems (Milanés et al. 2017a, b)
	Use in different disciplines, like geography, physical, urban or territorial planning, economy, environmental management, among others
	Use focused on the event effects associated to the threat, and not only on the event
	Restrictive view of the social, cultural, economical, and political aspects in assessing vulnerabilities and risks (Cardona 2001)
Technical sciences	Design and use of geographical information systems for the first time to draw maps with zones of danger or menaces in relation to the areas of influence of natural phenomena (Maskrey 1998)
	Use of specific and differentiated studies and methodologies to analyze the different kinds of risks
	Use of matrixes of damages to relate the phenomenon intensity with the damages caused (Soloviev 1978)
	Assessment of scenarios for potential losses, e.g., earthquakes in urban areas
	Use of decision-making processes for territorial ordering and physical planning
Social sciences	Risks estimated by analysis, or empirical information of the menace probabilistic modeling and possible damages (Cardona 2001)
	Study of people’s individual and collective reactions and perceptions toward risks (Mileti 1996; Drabek 1986)
	Study of the inhabitants’ reactions or responses to general emergencies, not only to risks (Cardona 2001)
	Analysis of the social character of vulnerabilities, not only the physic-potential damages, or demographic aspects
	Assessment of vulnerabilities, making reference to the inhabitants’ ability to respond and recover by themselves after an event impact (Westgate and O’Keefe 1976)
	Analysis of social factors, like fragility of family economy, absence of basic social services, lack of accesses to properties and credits, presence of ethnical, political, or other kinds of discrimination, high rate of illiteracy, lack of educational centers, among others (Maskrey 1994; Lavell 1996; Cardona 1996, 2001; Wilches-Chaux 1989; Mansilla 1996)
	Prevention and reduction used to decrease the pressure of social factors from global to local levels. Therefore, risk mitigation is directly involved in security conditions, dynamic pressures, and in other factors of influence (Wisner 1993; Cannon 1994; Blaikie et al. 1996; Cardona 2001)
	Risk conceived as something imaginary; thus, individual and collective perceptions, attitudes, and motivations vary according to the corresponding contexts (Johnson and Covello 1987, 2012; Slovic 1992; Luhmann 1993; Maskrey 1994; Adams 1995; Muñoz-Carmona 1997)
	Environmental impact and potential physical damages considered no essential when estimating risks (Cardona 2001)
Holistic approach	Integration of all disciplines
	Not only the expected physical damages, victims, or equivalent economical losses taken into account to estimate risks, but also the social, organizational, and functional factors related to the communities’ development (Cardona 2001)

(continued)

Coastal Risk, Table 1 (continued)

Disciplinary approaches	Characteristics and evolution
Integrated, multimethodological, and transdisciplinary approach	Dialogue among specialists of different sciences
	Interaction of different disciplines in relation to approaches, categories, and specialties (Milanés et al. 2017b)
	Respect for specialists' knowledge and multiple opinions when expressing different perspectives
	Unity of relationships, reciprocal actions, and interpretations of the different disciplines, keeping always their limits
	Search for unity of factors concerning opinions, methods, and languages
	Understanding the ways of dealing with flood risk management, or coastal risk interventions according to the physical, conceptual, and organizational borders (Nicholls et al. 1999; Bracken et al. 2016)
	Addition of multidisciplinary perspectives for integrated management of coastal risk and integrated management of coastal zones due to their essential link (Milanés et al. 2015a; Botero and Milanés 2015)
	Coastal risk, like flood risk, is considered an unavoidable and significant economic and social challenge to communities all over the world; therefore, the use of progressive and holistic approaches is necessary for adaptable management (O'Hare et al. 2016)
	The multidimensional nature of the assessment approach allows integration of at scale-information about cities, communities, and houses. This approach is novel in leading assessments about physical, social, institutional, and economic themes, as well as in understanding the ways of interaction among the underlying processes (Herslund et al. 2016)

Several definitions for the term vulnerability have been found in the bibliography reviewed (Cardona 2001, 2003; Batista Matos 2006; Lavell 2009a, b; Ruiz Rivera 2012; Guasch et al. 2008; Milanés 2014; Milanés et al. 2015a, 2017a), although two issues are analyzed in all of them: (1) probability of the territory exposed to risks of undergoing damages in case of threats and (2) preparation of the territory to face threats.

Current Tools Used for Coastal Risk Assessment

At present, various methodologies and tools are being used for determining coastal risks before the occurrence of natural threats or other kinds of danger (ICAM 2009; EMNDC 2005; AMA 2014a, b; Botero and Milanés 2015; Botero et al. 2017; Milanés et al. 2017b). Additionally, those tools are very helpful for taking preventive actions in order to avoid disasters, give quick responses, identify vulnerabilities, and calculate possible risks. Table 2 shows the positive results of coastal risk management when some current methods were validated at four coastal geographical regions of North America, Central America, South America, and of the Caribbean Islands (see Table 2).

Structural, nonstructural, functional, social, ecological, and economical vulnerabilities are all assessed in the three methods currently object of analysis in Cuba (CITMA 2005; EMNDC 2005; AMA 2014a, b). To calculate coastal risks, all those six different kinds of vulnerabilities are taken into account to get total vulnerability. Maps of danger, vulnerability, and risks are constantly drawn throughout all stages of the methodological tools, which require active participation of

experts, including scientists, technicians and decision-makers for their analysis and application. During validation of the Cuban tools in some of the Caribbean Islands, permanent presence of members of “Estado Mayor Nacional de la Defensa Civil” was really outstanding.

Coastal Risk Management

Managing coastal risk means analyzing, identifying, characterizing, and studying the different coastal menaces with possibilities of occurring at a given time (Milanés 2014). This kind of management requires a whole process of interinstitutional and multidisciplinary negotiations, and is supported by strategies, politics, and actions directed to prevent disasters and favor effective planning of coastal territory. Coastal risk management is also aimed to avoid or reduce an event impact and the consequences to the society of the dangers caused by the threats. Additionally, the threat magnitude, intensity, frequency, duration, repercussion on human life, economy, and on other social activities are equally taken into account for coastal risk management; procedure expected to have the following results:

1. Identification of different kinds of threat scenarios, being floods, tsunamis, storm surge, hurricane, landslide, sea-level rise, among others, the most frequently predictable coastal risks. Constant assessment during the hurricane season is extremely important for Caribbean countries because of the great impact tropical cyclones cause on shallow seas and lowland coasts (see Fig. 1a–c).

Coastal Risk, Table 2 Some international current tools used for coastal risk assessment

Tools	Objectives	Extent and work scales	Steps
Guide for assessing risk management using ICAM framework (ICAM 2009)	To facilitate practical and realistic range-risk assessment by the States Members	Procedures directed to assess probabilities of coastal risks according to the event location, frequency, magnitude, and possible factors influencing on magnitude modification	Introduction
	To give reachable and sustainable strategies	Risk assessment or local physiographic factors to know boundaries in potential risk of undergoing an event impact; thus, authorities can be alerted about priorities to mitigate risks	ICAM context
	To develop people's resilience to emergency situations in coastal communities	Changes in levels of vulnerability throughout time, e.g., environmental changes and those in coastal urban zones	Risk description
	To reduce or mitigate risks	Both risk and vulnerability assessment, indicate the policy and decision-making processes to follow	Identifying and quantifying risks
		The geographical scale of vulnerability depends on the resolution and extent of the assigned coastal management area in the ICAM process, called "macro scale" for the regional level, and "micro scale" for the local level	Vulnerability calculation Risk assessment Increasing awareness and risk arrangement Risk mitigation
Methodology for studies about danger, vulnerability, and risk of disasters by heavy rain and floods (AMA 2014a)	To calculate danger for different return periods	Analysis and evaluation of hydrographic basins, giving the results to the corresponding Province, Municipality, or Popular Council	Danger calculation:
	To mark dangerous zones on cartographic maps using geographical information systems	Work scale at 1:25,000, or higher	Identification of zones prone to flooding
	To identify all elements exposed to risk, calculate vulnerabilities and estimate risks using geographical information systems		Calculation of danger intensity and return periods Vulnerability calculation Risk estimation
Methodology for studies about danger, vulnerability, and risk of disasters by sea floods (AMA 2014b)	To calculate danger for different return periods	Identification of vulnerabilities and risks of elements exposed to several scenarios of danger	Calculation of danger by coastal floods:
	To mark dangerous zones on cartographic maps using geographical information systems	Presentation of results to the corresponding Province, Municipality, and Popular Council	Numerical modeling of surge
	To identify all elements exposed to risk, calculate vulnerabilities, and estimate risks using geographical information systems	Work scale at 1:25,000, or higher in case of detailed studies	Numerical modeling of waves
	To analyze and compare risk of every Province, Municipality, and Popular Council using the cartographic maps drawn by geographical information systems		Vulnerability calculation
	To do technical reports using the cartographic maps of every unit of analysis		Risk estimation

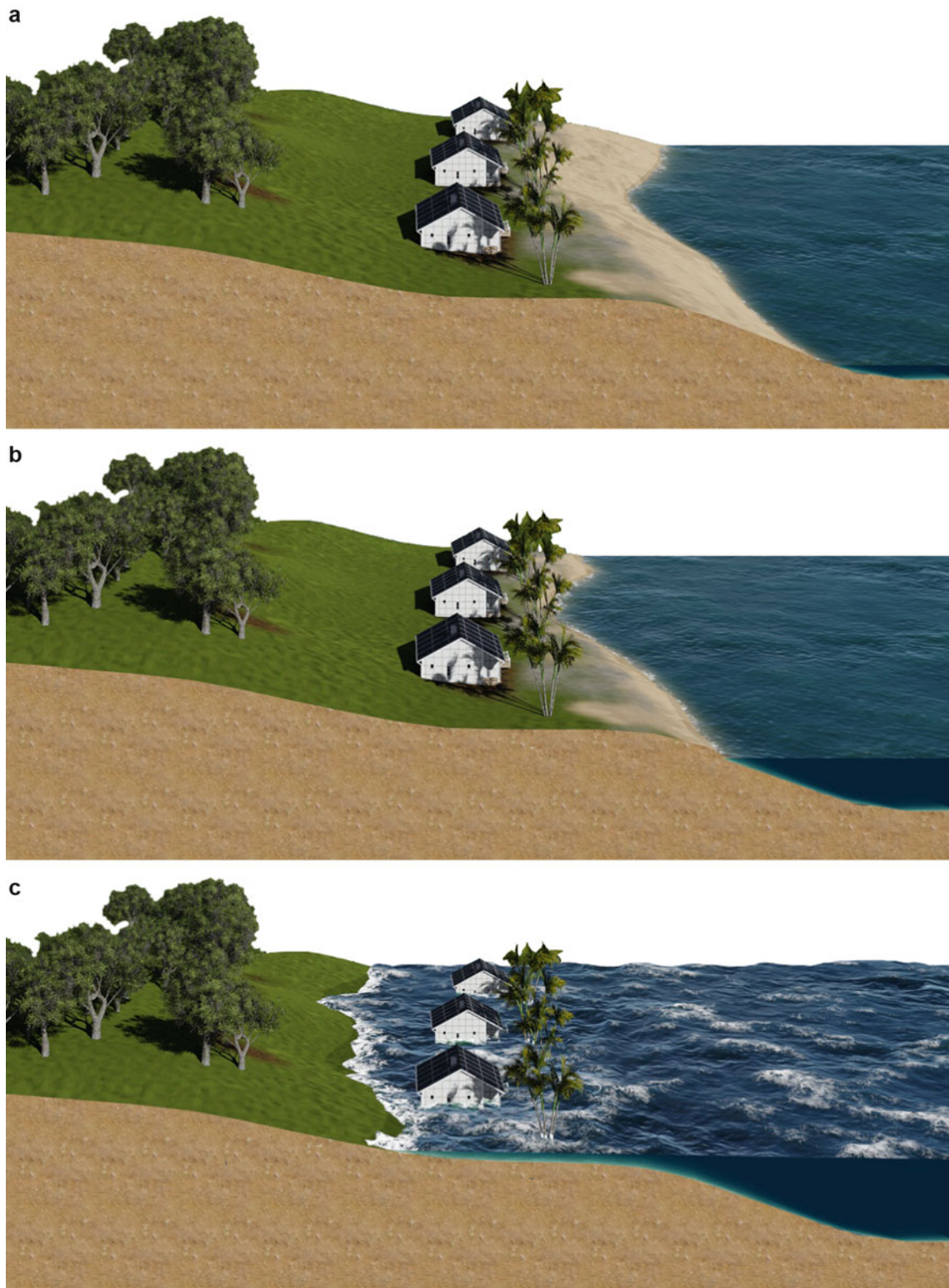
(continued)

Coastal Risk, Table 2 (continued)

Tools	Objectives	Extent and work scales	Steps
Methodology for studies about danger, vulnerability, and risk of disasters by intense wind (EMNDC 2005)	To calculate danger for different return periods	Use of the existing information and studies to work at every Popular Council. Presentation of results to different levels: Province, Municipality, and Popular Council (PC)	Danger calculation using:
	To identify elements exposed to danger, calculate vulnerabilities, and estimate risks using geographical information systems	Work scale at 1:25,000, or higher	The climate variable
	To draw cartographic maps marking areas of vulnerability and risks		Characteristics of the observational series
			Statistical modeling
			Vulnerability calculation
			Risk estimation
Methodology for formulating marine–coastal indicators associated to risk management (Botero and Milanés 2015; Botero et al. 2017)	To use the four marine–coastal indicators and corresponding sub-indicators for calculating interaction between governmental actions and coastal risk management	Use of methodological files with the four designed indicators	Enough knowledge about risks
		Work done at municipal coastal zones	Measures for risk mitigation
		Assessment of scenarios includes coastal particularities and variables associated to integrated coastal management	Preparation of institutional responses in case of emergencies
			Interinstitutional coordination and cooperation
Procedure used in four general processes for risk management and disaster mitigation (Milanés et al. 2017b)	To identify and diagnose different kinds of coastal risks	Holistic character	Four general processes to follow:
	To define future actions for improving present situations and introducing result analysis based on the principle of continuous improvement	Continuous assessment of various kinds of risks to reduce vulnerabilities in case of disasters	Diagnosis of institutions and organizations involved in risk management and disaster mitigation
		Scale variation according to the object of analysis, oscillating from 1:1,000,000 for national scale, up to 1:2,000 for local scale	Risk identification and assessment
			Establishment of strategies and plans of actions to reduce risks
			Continuous improvement in relation to risk management

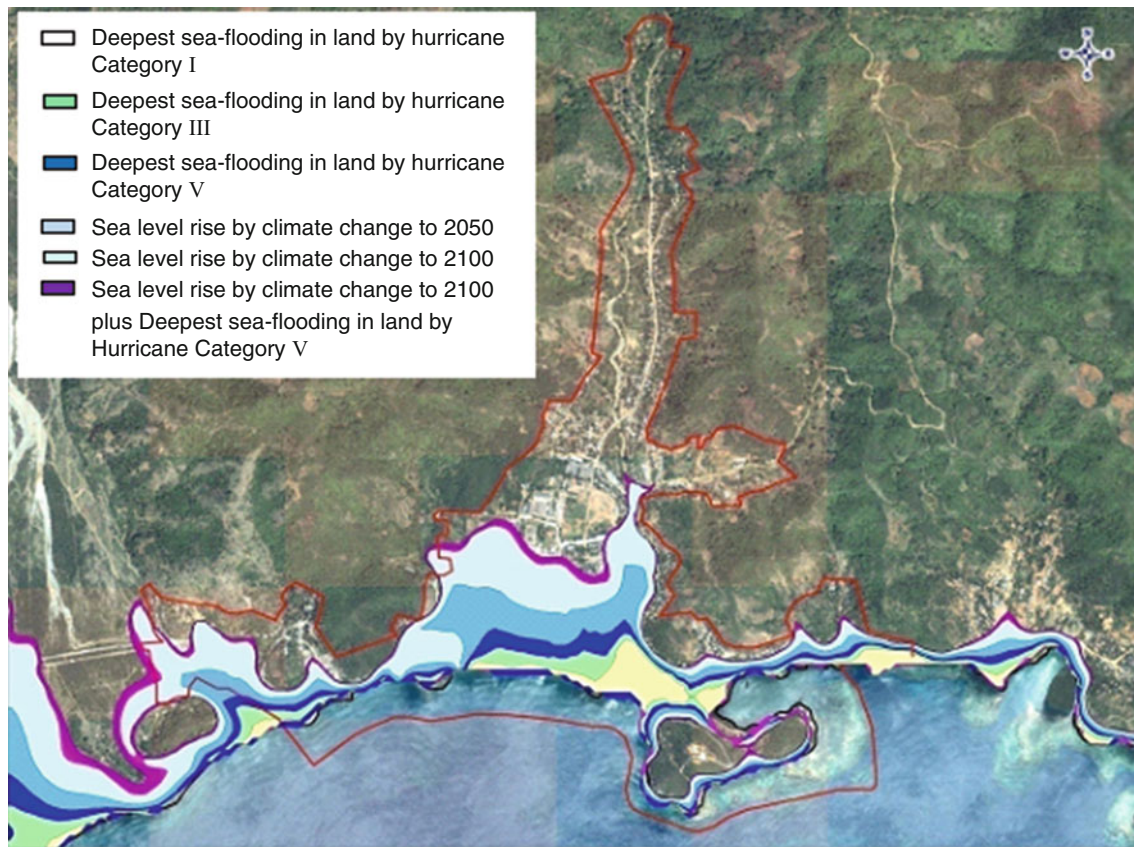
2. Determination of geographical scales and assessment of risk boundaries (Milanés 2014; Milanés et al. 2017b). Use of geographical information systems to analyze combined dangers and draw maps including areas in danger of temporary coastal sea-flood caused by hurricanes of different categories (I, II, and V). At the same time, permanent flood scenarios of sea-level rise, up to 2050 and 2100, have to be also represented on multidanger maps (see Fig. 2). In this way, all the assessed information, zones of the territory exposed to danger, boundaries of areas for future zoning, and ordering of coastal regions are also geospatially represented on the maps.
3. Presentation of proposals about preventive, prospective, and corrective actions, including specific actions for

financial protection (Guasch et al. 2008; Milanés et al. 2017a). Those actions must include tasks for preparation and responses to the destructive effects of the threat to guarantee protection to population and economic goods. Once the event is over, a recuperative phase must be declared to rehabilitate and reconstruct all vulnerable urban and eco-systemic elements, among other necessary actions. The measures taken for risk and damage mitigation (ICAM 2009) must be assessed for repetition of other kinds of risks in the near or far future. Three strategies for damage mitigation: protection, accommodation, and retreat, have been stated by Bijlsma et al. (1996) (see Fig. 3a–c).



Coastal Risk, Fig. 1 (a) Normal tidal conditions: low water in coastal lowlands (Source: Modified from ICAM (2009)). (b) Tropical cyclone conditions: high water in coastal lowlands (Source: Modified from

ICAM (2009)). (c) Tropical cyclone conditions: high water in coastal lowlands; storm surge conditions or extra-tropical storm; coastal lowland flooding and surge level (Source: Modified from ICAM (2009))



Coastal Risk, Fig. 2 Analysis of scenarios of combined danger in case of sea-level rise caused by climate change and hurricanes of different categories

Conclusion

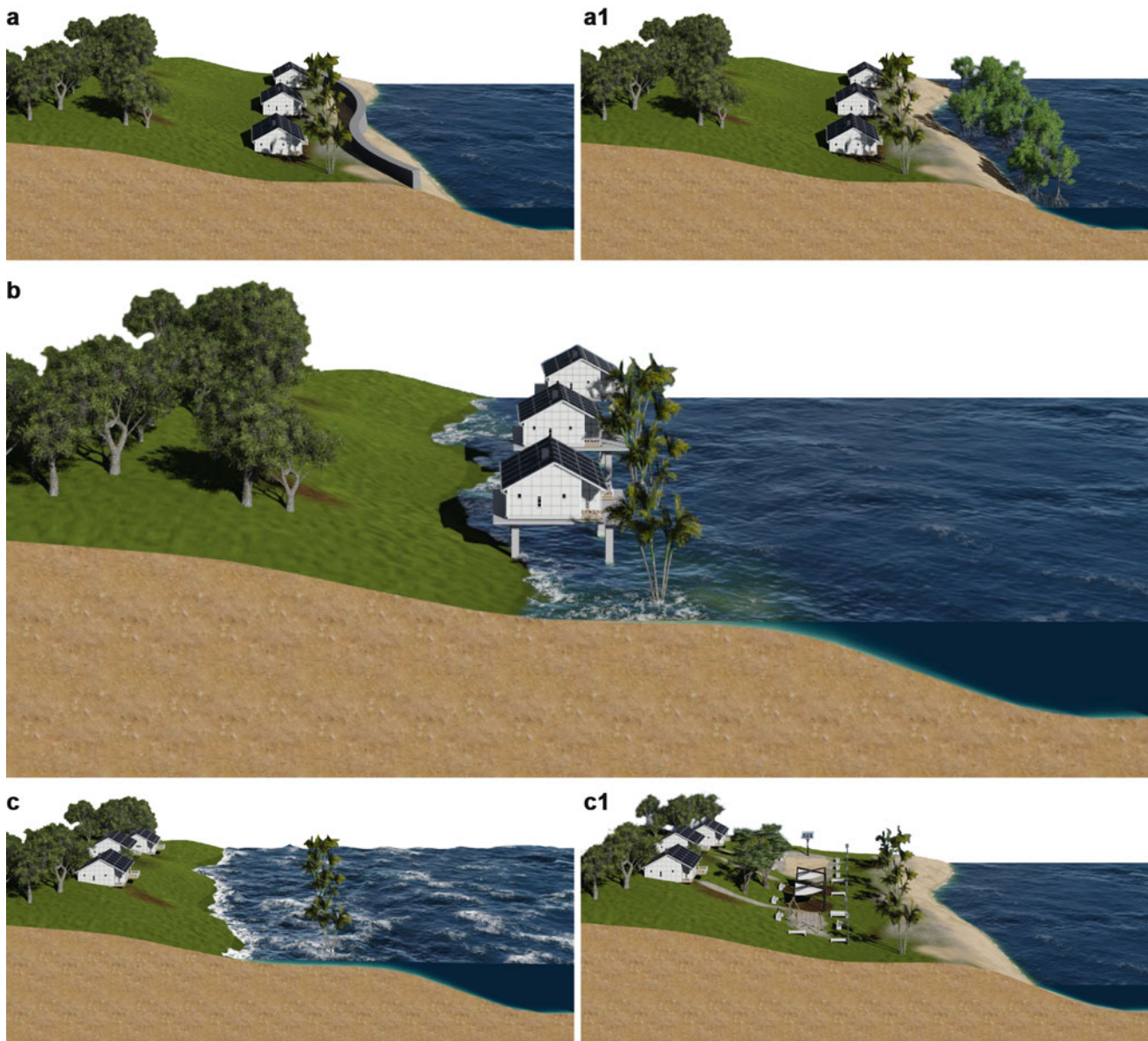
The peculiar characteristics of coastal zones of being fragile geographical areas exposed to threats of different origin, which can cause numerous disasters if not managed properly, give great importance to studies about coastal risks.

The theoretical frameworks and various terms about coastal risk analyzed in the present chapter, lead to conclude that vulnerability constitutes an inner risk factor of urban coastal zones because of the probability of those areas to be affected by natural or anthropic threats, and the capacity of the corresponding authorities and institutions to face them. As it is generally impossible to modify threats, and eliminate or minimize coastal risks, the alternative for reducing their consequences is, therefore, to increase knowledge about vulnerabilities.

Vulnerabilities of a coastal urban environment are conditioned by the following eight factors: (1) location of human settlement at the first coastal line; (2) number and density of

settlement population; (3) restriction to access or mobility of resources; (4) absence of basic medical services; (5) degradation of habitat and ecosystems of natural or built environment due to the climate change; (6) poor economic, political, institutional, and governmental will, at different levels, to manage risks; (7) poor, or null, urban – territorial planning; (8) absence of integrated approaches in preventive studies about threats, vulnerabilities, and risks.

Numerous advantages were observed when assessing some methodologies for coastal risks, e.g., to help in using the land properly, in taking actions for urban ordering and planning, and in proposing future options for areas free of risks; to be significant indicators for environmental management of urban-coastal zones and for the establishment of governmental politics and plans of response at different levels; to help in determining the environmental impacts to be taken into account for future projects, and in keeping the sustainable development of all the spaces used; to help in doing holistic analysis of threats, vulnerabilities, and risks as a contribution for urban zones to increase resilience to all kinds of threats.



Coastal Risk, Fig. 3 (a) Anthropic structure for protection (pier or jetty). Hard solution. (a1) Protection with natural element (mangrove). Restoration or *soft solution* (Source: Modified from Bijlsma et al. (1996)). (b) Accommodation (Source: Modified from Bijlsma et al.

(1996)). (c) Retreat. (c1) Controlled retreat and territorial planning. Urban proposal for external public areas at vulnerable zones (Source: Modified from Bijlsma et al. (1996))

Recommendations about the best practices for risk mitigation include responses of Hard and Soft Engineering to possible impacts, reducing, therefore, exposure to different kinds of dangers. Additionally, nonstructural and adaptable actions, environmentally sustainable, like: reforestation, rehabilitation, and, if necessary, relocation of vulnerable communities to safer places, are proposed in the present chapter. Measures aimed to build special structures for emergency shelters have to be taken in case of communities having no other choice than living on land in great risk of floods.

Cross-References

- [Changing Sea Levels](#)
- [Coastal Boundaries](#)
- [Coastal Flood Hazard Mapping](#)
- [Coastline Changes](#)
- [Geomorphology and Sea Level](#)
- [Global Vulnerability Analysis](#)
- [Natural Hazards](#)
- [Sea-Level and Climate Change](#)
- [Storm Surge](#)

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Coastal Scenery

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Definition

Scenery has been analyzed in a variety of ways, and a lack of consensus exists as to the precise source of its aesthetic quality via the expert and perception-based approaches, culminating in landscape/seascape character assessment approach techniques being assessed. A semi-objective methodology utilizing weighting parameters and fuzzy logic mathematics is given as an example to assess coastal scenery, in which the main parameters (26) that influence people's assessment of coastal scenery were obtained from over >1000 beach interviews.

Introduction

The importance of landscapes to society for recreation, spiritual nourishment, and posterity has long been recognized. Some landscapes are intrinsically more important than others and designated as such by the plethora of existing classifications, e.g., "national parks," "heritage coasts," "areas of outstanding natural beauty," "wilderness areas," "protected landscapes," and "green belts" all reflect scenic beauty. Because of coastal tourism, beautiful coastal scenery is an asset to any community and an economic resource and therefore not a disposable luxury resource.

In recent years the number of people visiting scenic coastal landscapes for recreational purposes has increased dramatically, and correspondingly it is necessary to protect and conserve these attractive landscapes. Environmental managers have been faced with a dilemma: should coastal landscape development be impaired for the sake of conserving the

natural scenery or not? This revolves around which landscapes are favored by society and requires the use of techniques that evaluate the relative quality of scenery so that comparisons can be made according to resource user's needs. This is particularly significant to coastal scenery, as coastlines currently face enormous pressure from anthropogenic factors seeking to develop the area to facilitate activities, such as tourism and residential housing, and studies have shown that excellent scenery is a prime tourist factor in choosing a beach holiday – the others being water quality, safety, no litter, and facilities, the weightings varying with destination selection, resort to remote (Williams and Micallef 2001).

Establishing a coastal scenic evaluation technique that is credible and accurate has often proved difficult. Scenic evaluation can be as basic as observers pointing to features, such as, trees, water that enhance the view, as against pylons, industrial chimney stacks that disfigure it. Observers will not/cannot see the same landscapes, and the effects of elements in the form of numbers, colors, etc. might be in agreement, but the facts only take on meaning through association. There is a lack of consensus regarding the precise source of aesthetic quality: is it inherent in the physical landscape or is it a psychological product? The former theory is termed the objective paradigm and the latter the subjective paradigm, and these opposing premises underpin two approaches to any landscape evaluation: the "expert" and "perception-based" approaches, respectively. The expert approach relies on a standard of assumptions made by an expert surveyor regarding attractive and unattractive scenery, and each landscape is evaluated according to this standard, e.g., Fines (1968) and Linton (1982). Since Leopold's (1969) seminal paper, the academic study of scenic assessment has developed, and it is now possible to apply some form of scenic assessment to any landscape or scene, and this is increasingly relevant to coasts. This approach has been largely rejected by the academic world because it assumes that scenic preferences are homogeneous between individuals. However, scenic appreciation is dependent on a complex interaction of factors with the precise influence of each being invariably interlinked. These factors include psychological constructs and sociodemographic variables, which combine in different ways to trigger a unique emotional response when an individual views a landscape. Perception of coastal scenery is therefore by no means universal and varies from person to person, and for this reason, the methodology of an objective "expert" evaluation can be viewed as erroneous.

Landscape/scenery is bound up in semantics. Sauer (1963, 319) defined landscape as "... an area made up of distinct association of forms, both physical and cultural... The content of landscape is found...in the physical qualities of the area that are significant to man and in the form of his use of the area... In facts of physical background and facts of human culture," i.e., the natural landscape is present in preman's activities and a cultural landscape fashioned from the natural by the activities

Srøksnes, T (2017) 'The landscape is not in front of me. It's all around me', 'Sharkdrunk', Jonathon Cape, 320 pp.

of a cultural group. It is composed not only of what lies before the eyes but what lies within the heads. Different groups of people view landscapes in different ways, e.g., a geologist will look first at rocks and a botanist at flora. Scenery on the other as defined by any dictionary is the aggregate of picturesque features in a landscape. Scenic beauty is deemed to be the central dimension of man's relationship with the environment representing the highest quality common factor derived from focused acts of environmental appreciation, and it is a prime factor in landscape evaluation. It evokes aesthetic responses that form part of a hierarchy of values representing a source of self-actualization for which *Homo sapiens* strives once basic biological/psychological needs are met.

Evaluation simply implies choice/comparison between two or more defined landscape areas, and the evaluation procedure should have clear aims, which should involve certain features: the ability to observe, assessment/measurement criteria, a theoretical basis, a means of recording/summarizing/ interpreting, and communicating the measurements. So, what and why do people like about landscapes, as these words are of concern for coastal management, waterways, land areas, etc.? Frissell et al. (1980, 160) referring to Yosemite Valley National Park, the first in the USA and founded in 1864, wrote "... it is the will of the nation embodied in the Act of Congress, that this scenery shall never be private property, but that like certain defensive points upon our coast, it shall be held solely for public purposes."

Different approaches have been devised mainly of a subjective nature, but more and more public preferences and the use of photographs have taken center stage; however, two main problems exist:

- Defining a landscape that people enjoy and responses to a given landscape and the psychology of environmental perception varies enormously.
- Describing aesthetic values quantitatively – essential for resource users.

History and Problems Associated with Landscape Evaluation Techniques Regarding Scenery

No one doubts the need for landscape evaluation, but no one can agree as to how to go about it. Evaluation, having diverse objectives encompassing preservation, protection, recreation, etc., is concerned with quantifying matters into numbers and distributions which can then be turned into maps, graphs, etc. Initial attempts at scenic classification aimed at direct measurements (Fines 1968; Leopold 1969), but with no underlying theory, the opinion of "users" was initiated. This did not decry the use of "experts" but provided a level playing field for both professionals and the man in the street. Objectivity was frequently claimed, but subjective judgments became the norm. Most techniques failed to take into account

the scenic quality at a point and the views from all directions from that point. Can one consider a piece of land in isolation? Many direct methods depended upon a gross scoring method with broad generalization and no formal recognition of basic aesthetic principles in the determination of scores assigned.

(a) **Measurement techniques.** Three environmental conditions have to be accounted for in any technique: the state of mind of the observer, the context of observation, and the landscape itself (environmental stimulus). Landscapes are dynamic elements constantly changing day to day, year to year, via weather, erosion/deposition processes, etc., so surveys must be done in a short time span and in comparable weather conditions.

(i) *Observer choice.* Differences in personality usually mean that visual beauty is felt differently. Training could increase the receptivity to beauty, but it could also constitute a potentially influential variable. Is there a "good fit" between a small group of "experts" and the public?

(ii) *Observer number.* Traditionally one or two people were involved in order to have consistency, but these were not necessarily a representative sample. Higher numbers are needed to give composite judgments.

(iii) *Scoring scales.* A lone observer could use a personal scoring scale but not if the results were to be compared with another observer at another site with his/her personal scale. Usually the observer gave a score within set limits, e.g., equal interval scales (1, 2, 3, 4, 5, etc.), geometric scales (1, unsightly; 2, 4, 8, 16, 32, spectacular; Fines 1968), or nominal grades (low, medium, high). Problems with ordinal/rank scales are that they are not open ended making an observer reluctant to give the highest/lowest score. Comparisons between areas are difficult due to the large variety of these scoring scales. Linton (1982) divided landscape into landform (1, lowland, to 6, mountains) and landscape (1, urban/industrialized, to 7, wild) and in addition gave a number suitable for planning purposes. Leopold (1969) in a response to Snake River dam building did away with numbers by utilizing five attributes (ugly–excellent) to expand upon 46 physical, biological, and human factors deemed to be important in scenic assessment, looking at valley and river character, and based assessment upon uniqueness.

(iv) *Measurement units.* Landscape designers required the total field of observation to be defined, and a number of techniques adopted the 1 km² as a base unit, e.g., Coventry-Solihull-Warwickshire (CSW 1971), who delineated 24 prime factors, e.g., 1, hedges/trees; 15, listed buildings; and 23, views, and utilized step-wise multiple regression analysis to find out how far variations between measured factors of 1 km² and

those of another accounted for variation between square scores. The problem was defining the grid square on the ground. Other units included the drainage basin, parish, and landscape tracts.

- (b) **Mathematical techniques.** Landscape beauty is frequently dependent upon subtle characteristics of its components, which the eye can appreciate completely. The visual management techniques of the CSW (1971) study were lengthy and complex including a large number of separate evaluations resulting in a decision rule. Showing a great insight for future work in this field, Shafer and Brush (1977, 255), argued that "... not to use mathematics to help explain phenomenon of landscape preferences would be like trying to fell a tree with a chain saw without turning on the power switch." Multiple regression analysis was a common tool (Shafer and Brush 1977), as it took into consideration not only changes in the landscape variables but also the many subtle and unsuspected interactions between the variables. In environmental impact assessment, scenery is also measured by several methods, which evaluate the loss of beauty or homogeneity caused by a certain project or activity, normally through variable, such as extension, intensity, or accumulation summed in a single equation.
- (c) **Public preference techniques.** These had no practical limitations as the main concern was the use of surrogate measures and definitions of the nature of landscape. Problems arose from the fact that many surveys invariably asked questions of people who were already on sites, with the obvious risk of sample bias by persons who were attracted by the landscape to be present in the first place, but surely it is best to ask questions to respondents about scenery that they knew. Inherent familiarity can affect results either positively or negatively. Scenic preferences can be affected by content, e.g., water, tree, interest, and expertise. Controversy exists as to whether socioeconomic factors (upbringing, attitudes) play a part. Photographs have been used as surrogates for actual site visits, mainly because of cost. In this type of media, filtered reality can be biased, e.g., taken from the photographer's viewpoint. Williams and Lavalle (2001) showed that photos taken with a 50 mm lens when compared with an observer's view of the same scene (a box placed around the observer's head gave the same view as the exact scene photographed) and later tested via a t test showed that the null hypothesis was correct. Public attitude questionnaires (Penning-Rowsell 1989) can be subdivided into techniques having a strong factual basis (Leopold 1969), weak or nonexistent factual basis (Fines 1968), and evaluation based on landscape values (Penning-Rowsell 1989).

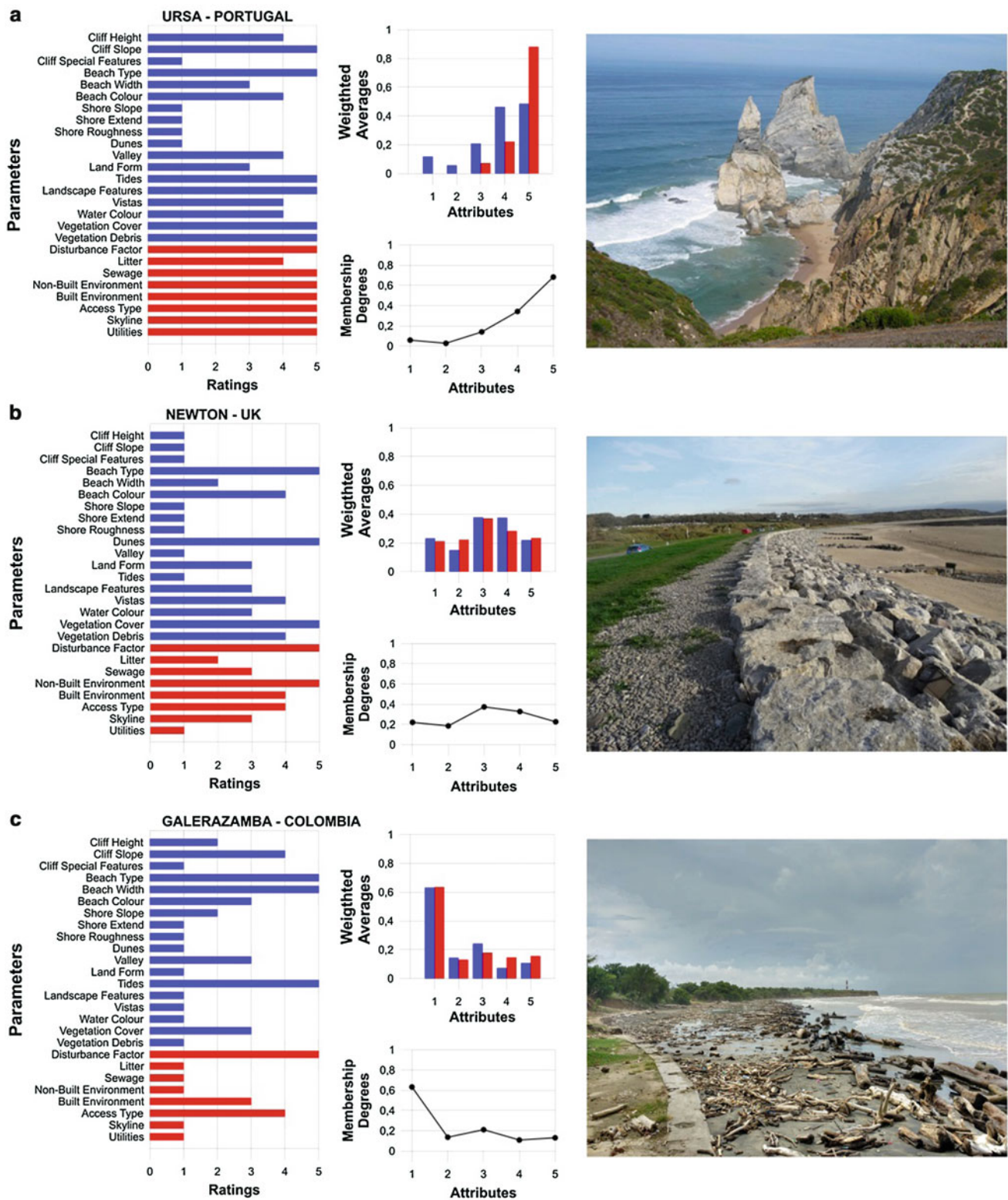
Amalgamation of subjective, objective, measured, and spoken seems to be the way ahead for future techniques in order to produce a technique suitable for areal comparison. Scenic

evaluation emerged mainly in the mid-twentieth century, as a technique for landscape classification and description, identifying particular landscape types or landscape features by referring to specified value criteria. The late twentieth century represented a watershed in the jettisoning of the idea of man's domination over nature to one of more respect, which unfortunately in the twenty-first century does not hold sway in many parts of the world, e.g., the Amazonian or Poland's Bialowieza logging industry a UNESCO Natural World Heritage Site. Logging is banned on the Polish side of the Belarus border, which accounts for only 17% of the whole forest.

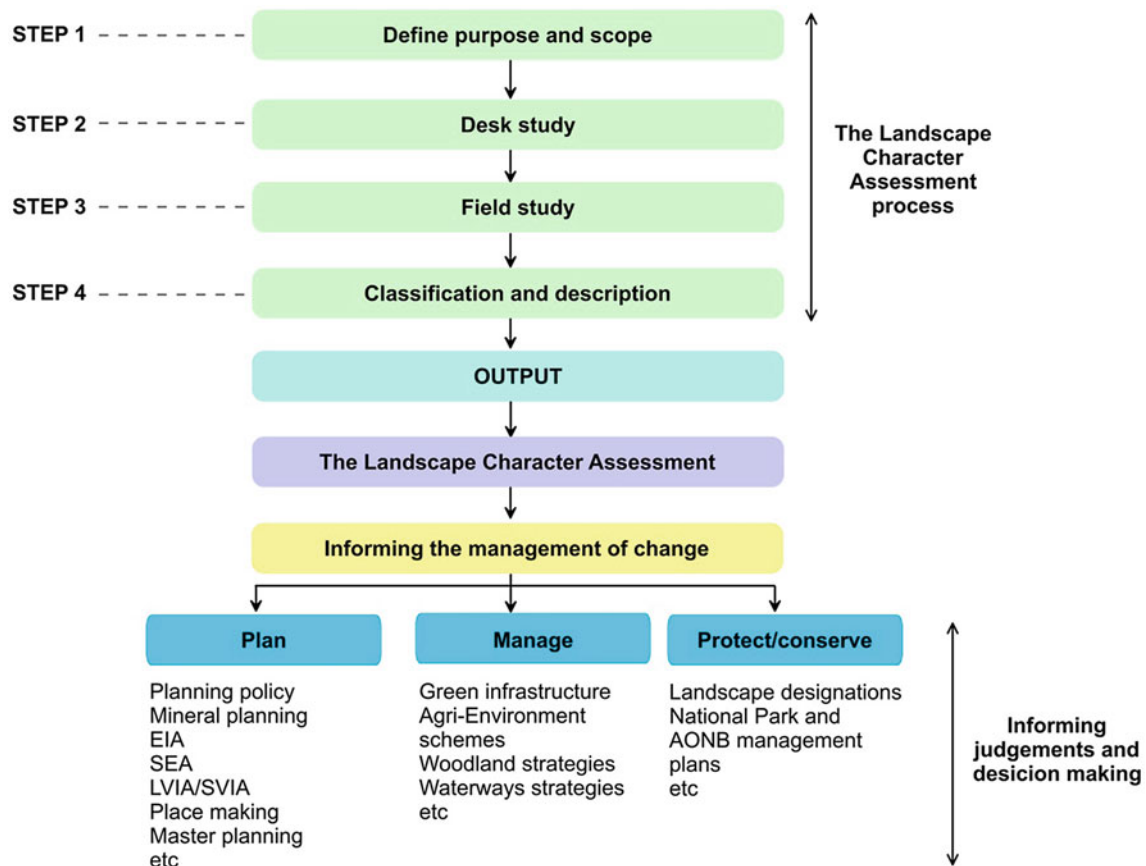
An innovative technique based upon Leopold's (1969) seminal paper produced a coastal scenic evaluation system based upon parameter selection obtained from several thousand interviews (Europe, USA, New Zealand), as to what parameters are needed for "good/bad" coastal scenery (Ergin et al. 2004). Twenty-six were deemed prime factors, and making use of Leopold's (1969) five-item attribute idea, these were weighted and subjected to fuzzy logic analysis to produce histograms, weighted averages, and membership degrees culminating in a "D value." This has been applied to >860 beaches worldwide. The D value classified locations into five scenic classes (I, excellent, D value >0.85 to V, poor, D value <0; Fig. 1).

In the twenty-first century, landscape/scenic assessment mutated into Landscape Character Assessment (LCA) defined as "a distinct, recognisable and consistent pattern of elements in the landscape that makes one landscape different from another, rather than better or worse" (LI 2016, 2). It was a quantum leap in scenic/landscape assessment, aided by the rise of computer power and satellite imagery and one which is still evolving (Fig. 2; Tudor 2014).

Landscape Character Assessment factors, i.e., natural (e.g., geology), cultural (e.g., land use), aesthetic (visual), and associations (e.g., history), have been incorporated into a LANDMAP (Landscape Assessment and Decision Making Process) system, via a partnership program between the Countryside Council for Wales (now Natural Resources Wales) and all the unitary and national park authorities of Wales (CCW 2001). Datasets recorded information about the characteristics, qualities that influence the geology, habitats, visual and sensory, historical and cultural on the landscape managed through a Geographical Information System, i.e., 5 map layers in a GIS format. It was a benchmark publication and a tool for assessing landscape resources to aid decision-making and natural resource planning at a range of levels (local to national) where landscape characteristics, qualities, and influences on the landscape are recorded, and areas were shown as single polygons (or polyline) in a single map layer. Scales utilized were typically 1:250,000, 1:50,000 or 1:25,000, 1:10,000, or larger scales. LCA has now been added to seascape character assessments, i.e., landscapes with coastal/sea views, so that coasts together with the adjacent marine environment which have natural, cultural, historical, and archaeological links with each other are dealt with as a whole, i.e., "an area of



Coastal Scenery, Fig. 1 (a) Ursa, Portugal, Class I; (b) Newton, Porthcawl, UK, Class III; (c) Galerazamba, Colombia, Class V (red colors, physical parameters; blue colors, human parameters)



Coastal Scenery, Fig. 2 Landscape Character Assessment and judgment making (NE 2014)

sea, coastline and land, as perceived by people, whose character results from the actions and interactions of land with sea, by natural and/or human factors” (NE 2012, 8). It covers five key principles:

1. Landscape is everywhere and all landscape and seascape have character (Srøksnes 2017).
2. Seascape occurs at all scales and the process of seascape character assessment can be undertaken at any scale.
3. Seascape character assessment should involve an understanding of how the seascape is perceived and experienced by people.
4. Seascape character assessment provides an evidence base to inform a range of decisions and applications.
5. Seascape character assessment.

Summary

Various techniques (measurement, mathematical, and public preference) have been utilized to assess coastal scenery and an innovative one introduced using weighting and fuzzy logic mathematics, which can be used rapidly in both landscape/

seascape character assessments. Perhaps for the future remotely sensed satellite, data could be a key factor, as data can be digitally processed for landscape classification into major physical divisions. These approaches would allow large landscape descriptions to be made rapidly. It also provides a framework for landscape surveys and evaluations to be made where basic map data is lacking or field survey resources limited.

Cross-References

- [Beach Management Tools](#)
- [Estuaries](#)
- [Marine Litter](#)
- [Rating Beaches](#)
- [Tourism and Coastal Development](#)
- [Tourism, Criteria for Coastal Sites](#)

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Coastal Seafloor Geomorphological Features, Classification

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Synonyms

Bathymetry; Coastal cartography; Coastal classification; Coastal geomorphological maps; Nearshore classification; Seafloor mapping; Seafloor topography

Definition

Coastal seafloor classification is based on the recognition of submarine features depicted by bathymetric patterns that are interpreted in terms of topography of marine origin or the identification of drowned terrestrial landforms. Coastal

seafloor geomorphological maps are usually the products of interpretation, visually displaying digital bathymetric patterns through either cognitive reasoning or the result of computerized autoclassification algorithms. There are advantages and disadvantages associated with each approach (computerized autoclassifications versus cognitive interpretations), and the choice of procedure depends on purpose. Both methods are beneficial classificatory approaches, as they complement each other when interpreting coastal seafloor geomorphological features.

Introduction

Although there are many different methods for classifying seafloor features, most have a special purpose with specific goals that rely on the product of the survey technique (e.g., Finkl and DaPrato 1993; Mumby et al. 1998; Greene et al. 1999; Finkl 2004a, b; Mayer 2006; Collins et al. 2007; Chust et al. 2008; Walker et al. 2008; Achatz et al. 2009; Brock and Purkis 2009; Finkl and Vollmer 2011; Pittman et al. 2013; Makowski 2014; Finkl and Makowski 2015; Makowski et al. 2015, 2016, 2017; Makowski and Finkl 2016). The exemplar of the southeast Florida continental shelf has been studied using remote sensing techniques since the 1960s. The basic framework of these specific shelf features were outlined for the first time in reconnaissance seismic reflection profile surveys conducted by Duane and Meisburger (1969). Spatial distributions of shore-parallel coral reef tracts, inter-reefal sediment-filled troughs, and outcrops of carbonate bedrock were identified in their seismic reflection profiles. Subsequent LIDAR (light detection and ranging) surveys (2001 and 2008) in the form of LADS (Laser Airborne Depth Sounding) produced remarkably detailed depictions of shelf bathymetry. This kind of digital data is amenable to color ramping where tonal variations can be keyed to depth. Combinations of texture, tone, and pattern in the digital data provide a basis for cognitive (manual) interpretation of seafloor features. Maps showing spatial distribution patterns of seafloor features, for example, those prepared by Finkl et al. (2005a, b, 2007), Finkl and Andrews (2008, 2009), Finkl and Vollmer (2011), Steimle and Finkl (2011), Finkl and Makowski (2015), Makowski et al. (2015), Vollmer et al. (2015), Makowski et al. (2016), Finkl and Vollmer (2017), and Makowski et al. (2017) all recognize a range of marine landforms and drowned subaerial forms.

Cognitively derived (i.e., manually interpreted) maps are normally composed by hand and based on visual inspection, after which the mapping units are transferred into some sort of georeferencing informational software (e.g., ArcGIS). Finkl and Makowski (2015) have shown that it is possible to classify bathymetric patterns using automated computer algorithms, such as unsupervised isoclustering and interactive

supervised autoclassifications, which were then compared with previously prepared cognitive maps of field-verified seafloor features.

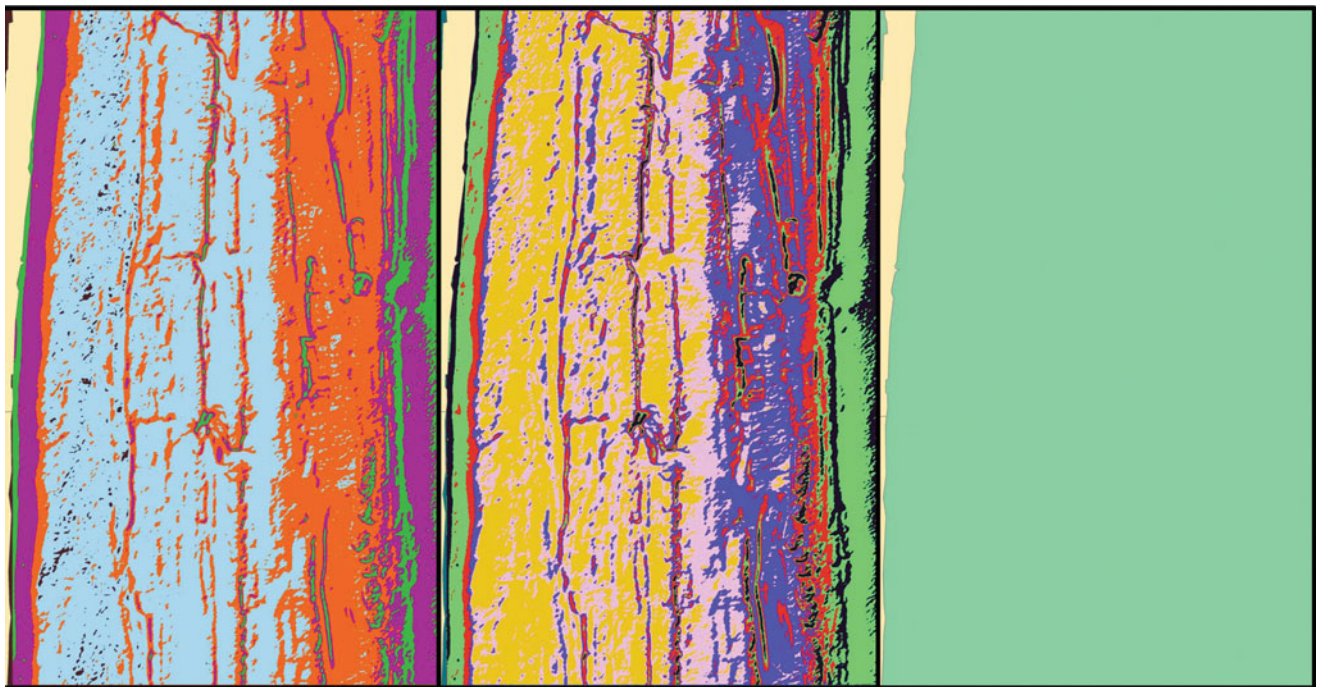
A good example of this involves the digital format of LADS (LiDAR) data, which makes the autoclassification of seafloor bathymetry possible. The basic procedure is to select a subset of the overall digital bathymetric dataset and compare the previously cognitive-identified geomorphological features (as described by Finkl et al. 2004, 2005a, b, 2008; Finkl and Andrews 2008, 2009; Finkl and Banks 2010) to auto-generated classifications of the same area. In order to determine whether autoclassification algorithms produce results comparable to cognitively interpreting, it is advisable to conduct a series of autoclassifying tests in appropriate geographical information system (GIS) software (e.g., ESRI's ArcGIS ArcMap).

In the example of the study area on the continental shelf off southeast Florida, the computerized classification of LADS digital bathymetry was divided into two parts, unsupervised isocluster and interactive supervised methodologies. Iterations of both methods were used to discern comparability with known seafloor geomorphological distribution patterns. Pixel digital numbers (DNs) were based on variations in color derived from color ramps that were applied to the DEM. Critical to this procedure is the number of classes that are discriminated by computer runs. Experimentation shows that

a large number of autoclasses produces patterns that generally are not comparable with previously cognitive-derived geomorphological maps (Finkl and Makowski 2014; Finkl and Makowski 2015). Iterations using a dozen or so classes produce recognizable patterns in interactive supervised classifications, whereas unsupervised isocluster autoclassifications require ten or fewer classes. Details of the autoclassification procedures are briefly summarized as follows.

Unsupervised Isocluster Autoclassification Method

Within ArcMap, the Image Classification toolbar needs to be activated and an unsupervised isocluster autoclassification performed on a subset of the study area. After the appropriate input raster (composite of the RGB bands) is selected and the output-classified raster is renamed in the proper geodatabase, the number of classes is assigned using the same number of features interpreted from cognitive processes. In the LADS subset, for example, where 14 geomorphological features were cognitively interpreted, the number of untrained classes assigned for the unsupervised isocluster autoclassification was 14. Because untrained class counts greater than 10 for unsupervised isocluster analysis of the LADS bathymetry tends to produce unusable results (cf. Fig. 1), additional



Coastal Seafloor Geomorphological Features, Classification, Fig. 1 Uninterpreted spatial distribution patterns derived from unsupervised isocluster analysis of LADS bathymetry DEM using five classes (left panel), seven classes (middle panel), and 14 classes (right panel). The larger number of classes in the third panel shows that

increasing the number of classes does not necessarily improve discrimination of the imagery because, even though the same gross patterns are evident, the increased number of classes requires additional field verification. (Source: Finkl and Makowski 2015)

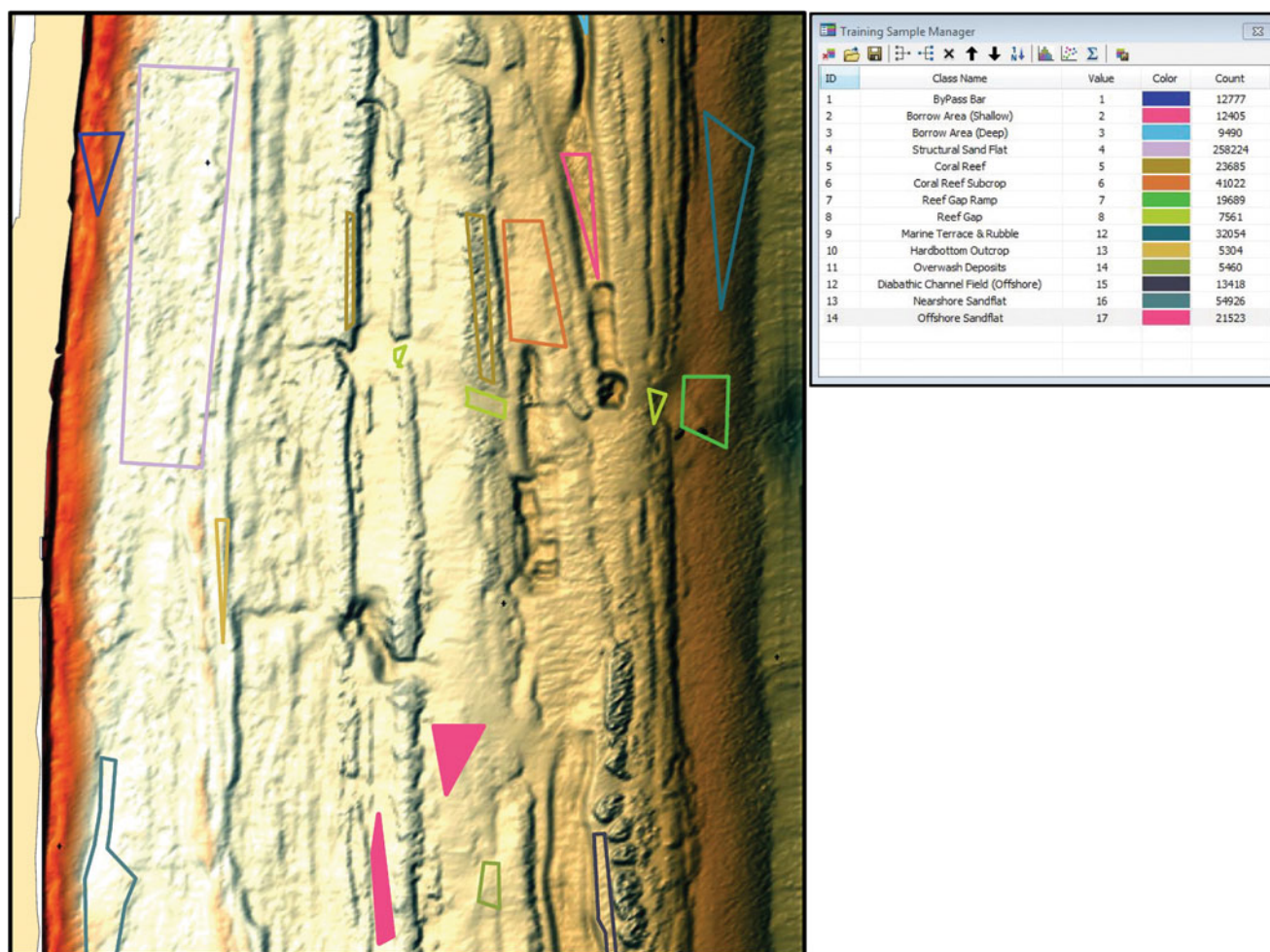
unsupervised isocluster classification analyses were performed with five and seven untrained classes, after which individual output rasters were saved as Tiff image files and imported as layers into ArcMap adjacent to the raw images. These displays can then be exported and visually compared with cognitively interpreted displays of the same imagery to visually determine the degree of correlation of autoclassification.

Interactive Supervised Autoclassification Method

Within ArcMap, the Image Classification toolbar is activated, and an interactive supervised autoclassification performed on the same subset as the unsupervised isocluster method. Once an individual scene is selected in the drop-down window from

the Classification toolbar, the Training Sample Manager window is opened. Training sites can then be established for the interactive supervised classification that corresponds to pixel patterns used previously for cognitive interpretation of the seafloor features. For example, in the LADS digital dataset where 14 geomorphological features were cognitively interpreted, 14 training sites were established for the interactive supervised classification to match the visual cognitive criteria. The training fields shown in Fig. 2 contain variable shapes that are determined by geomorphological landform patterns. Selection of training fields is critical because they have to represent the feature or pattern that is autoclassified. The operator must therefore accurately train the program to recognize specified features or patterns.

After the interactive supervised classifications are performed, individual output rasters can be saved as Tiff



Coastal Seafloor Geomorphological Features, Classification, Fig. 2 Locations of training fields used for the interactive supervised autoclassification of uninterpreted LADS digital bathymetry. Training sites represent the main landform features on the continental shelf, as determined from cognitive interpretation of texture, tone, and pattern

observed in the color ramped DEM, which was then interpreted in terms of field-verified geomorphological features. The training sample manager legend shows matched pairings of landform class names with a training field color. (Source: Finkl and Makowski 2015)

Coastal Seafloor Geomorphological Features, Classification, Fig. 3

Cartographic examples of methodologies based on unsupervised isocluster analysis with five classes (left panel) and interactive supervised autoclassification with 14 classes (right panel). Both examples show organization of bathymetric patterns that variably correspond to cognitively interpreted landform units on the seafloor. Generalization and increased discretization is evident in both examples, as the number of classes required is dependent upon the purpose of the classification and the amount of detail needed. (Source: Finkl and Makowski 2015)

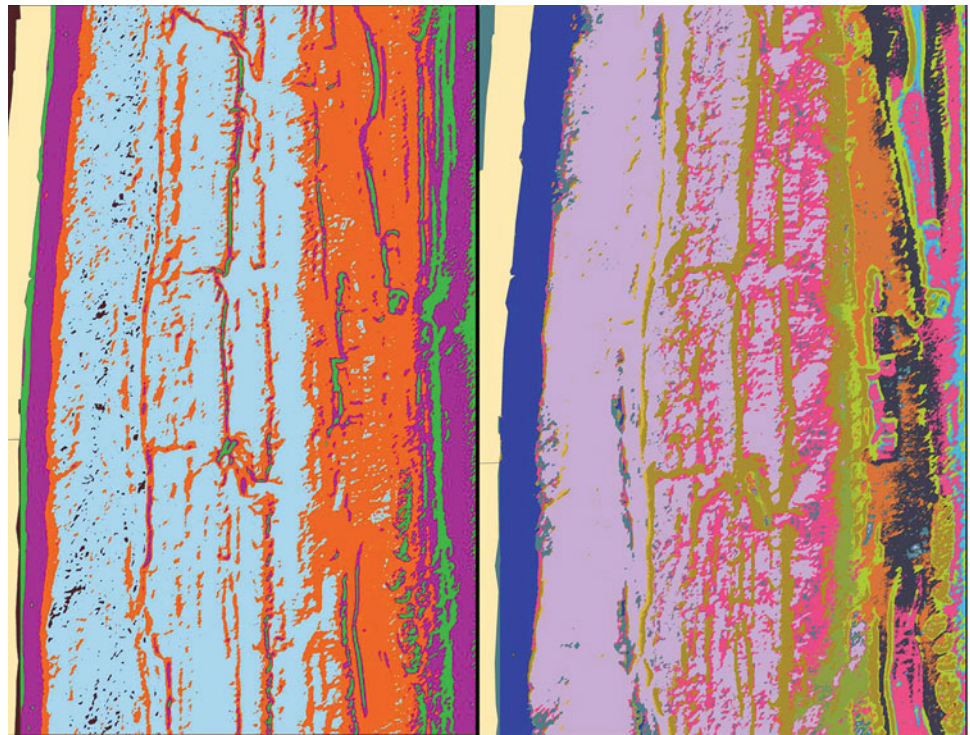


image files and imported as layers into ArcMap adjacent to the raw LADS images. These displays are then exported and visually compared with the cognitively interpreted maps to visually estimate the degree of correspondence with the autoclassification. Results of autoclassifying bathymetric data are shown in Fig. 3 where the left panel contains unsupervised isocluster classification using seven untrained classes and the right panel shows spatial distribution patterns derived from an interactive supervised classification using 14 trained classes.

Comparing Classification Results

Visual comparison of the autoclassification results with a control shows a gross correspondence between the unsupervised isocluster autoclassification and the control. Detailed comparisons can, however, get complicated along the shore where three units are differentiated for the single larger bar and trough units on the LADS imagery (Fig. 4a). Bathymetric texture, tone, and pattern attributes are detected in the unsupervised isocluster classification, suggesting a basis for discerning discrete landform units. Ground truthing is required to verify units in the unsupervised isocluster autoclassification, but in situ familiarity with ground conditions suggest the following features: a surf zone unit close to the shore, crenulated bars alongshore, and fans or sediment spays seaward (Fig. 4b).

On the other hand, the interactive supervised autoclassification shows more complex spatial distribution

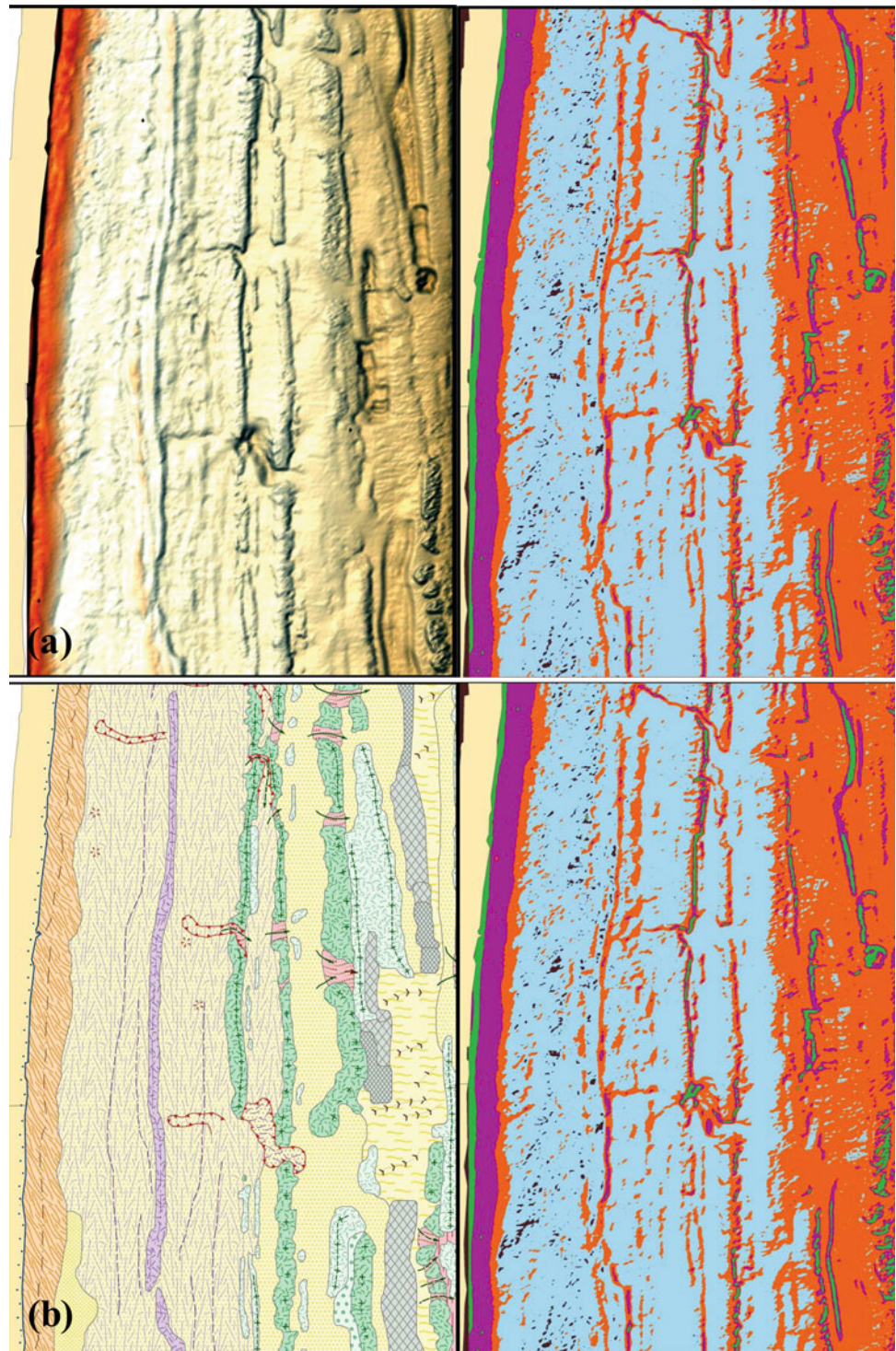
patterns that correspond to parts of the cognitive map, as shown in Fig. 5. This map may seem confusing because it is a montage of two different approaches for depicting seafloor geomorphology. Careful study, compared to a cursory glance, shows many important interrelationships. Starting at the shore and moving seaward, for example, the interactive supervised autoclassification matches the cognitive interpretation. Additional detail is provided in the form of distributions for nearshore sandflat, offshore sandflat, and hard-bottom outcrop.

Classification Determination and Validity

Before modern computerized mapping and GIS software packages, maps were validated in the field and compared with other maps known to be accurate. Although validation and accuracy assessments of computer-aided maps are usually achieved by statistical procedures and mathematical evaluation of self-tests, the visual comparison of machine-classified maps versus cognitively derived maps is simply a measure of correspondence. The advantage of these modern procedures lies in the pros and cons of an unsupervised isocluster autoclassification methodology and an interactive supervised autoclassification methodology when compared with mapping units based on the ratiocinative powers of the human brain. This process shows the possibility of producing composite maps of digital bathymetry.

Coastal Seafloor Geomorphological Features, Classification, Fig. 4

Comparison of unsupervised isocluster analysis (right panels) with unclassified (a) and classified (b) ramped LADS DEM. (a) Five-class unsupervised isocluster analysis (right panel) compared with (uninterpreted) color ramped bathymetry (left panel) showing the general correspondence of shore zone bar and trough features, coral reef tracts, bedrock outcrops, and dredge pits for beach nourishment projects. Symbolization of the geomorphological seafloor features is given in the legend to Fig. 6. (b) The same five-class unsupervised isocluster analysis (right panel) compared with cognitively interpreted geomorphic features on the shelf (left panel). These figures show generalization that occurs in maps hand-drawn at a nominal scale of 1:600 compared with greater detail that is acquired in autoclassification. (Source: Finkl and Makowski 2015)



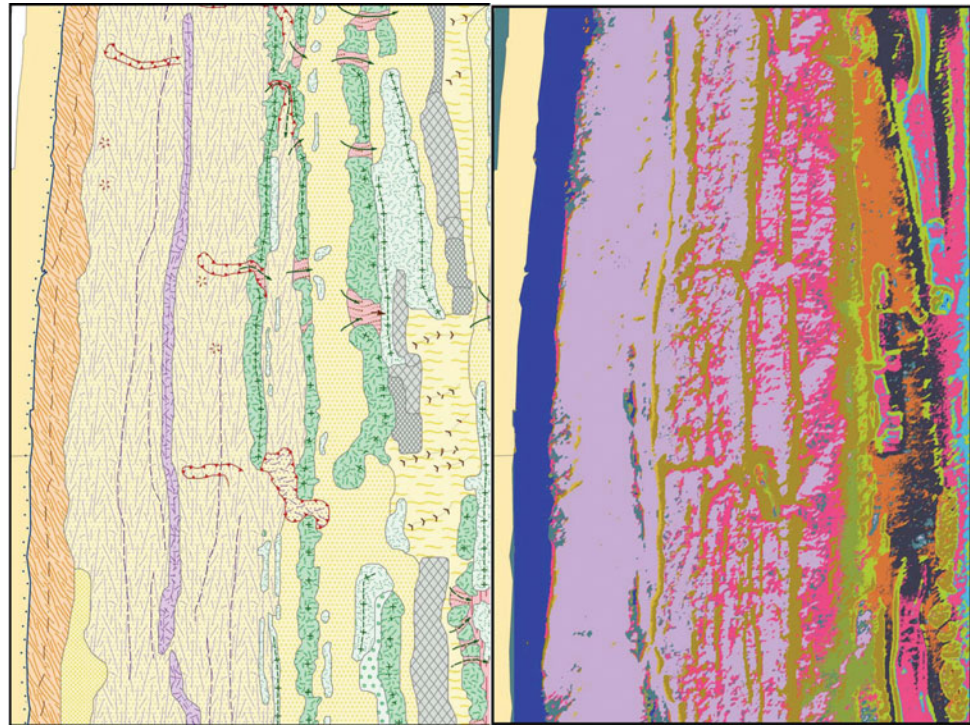
Rationale for Classificatory Design

Digital bathymetric data provide new opportunities to investigate the nature of seafloor topologies. Digital data provide a dense sampling medium that can be manipulated to form a DEM of seafloor bathymetry. Elevations can be

exaggerated to emphasize topographic differences that assist cognitive interpretation processes. A color ramp keyed to bathymetry can be draped over DEMs with an assumed light source to produce shadows. This colored representation of seafloor topography lends itself to visual inspection and consequent interpretation of

Coastal Seafloor Geomorphological Features, Classification, Fig. 5

Interactive supervised autoclassification (right panel) compared with cognitively interpreted geomorphic features on the continental shelf (left panel; a field-verified, color ramped DEM). This comparison shows the benefits of an interactive supervised autoclassification, which enhances the results of cognitive mapping seafloor features on the continental shelf. Cartographic generalizations in the cognitive maps are refined in supervised autoclassifications where the number of classes can be varied depending on special purpose investigations. (Source: Finkl and Makowski 2015)



bathymorphological features. This kind of observation and interpretation was not possible prior to the advent of digital surveys. The sample density of the LADS surveys (4-m resolution) was sufficiently detailed enough to allow construction of maps that allow interpretation of mesoscale geomorphological features (e.g., Finkl et al. 2004, 2005a, b, 2008; Finkl and Andrews 2008, 2009; Finkl and Banks 2010). This procedure is a major advance in the study of clear-water continental shelves and permits researchers comprehensive views of seafloor topologies.

Not all digital bathymetric data are amenable to automated classifications of the type applied to digital aerial and satellite images. In the case of hyperspectral data, for example, unsupervised classifications are more or less inept for interpretation of seafloor features. There are many reasons why hyperspectral data, usually measured with hundreds of narrow spectral bands about 10 nm apart, have prohibitive use for classificatory purposes. For example, in order to accurately obtain spectral signatures from hyperspectral data for different entities, such as different phytoplankton classes, zooxanthellae clades, or scleractinian (i.e., hard coral) species, it is necessary to first provide in situ spectral information. Without this information, the imagery cannot be interpreted because a function of the natural spectral variability is unknown. One such effort took place at Buck Island, St. Croix, U. S. Virgin Islands, where hyperspectral data obtained by AVIRIS (air-borne visible infrared imaging spectrometer)

was used in response to a mass coral bleaching event in the Caribbean. Kruse (2003) used the visible spectrum of AVIRIS light data reflecting off the coral reef and the surrounding reef bottom in order to estimate the extent of the bleaching, as well as the overall health of the coral colonies. Underwater handheld spectroradiometers first had to be used to measure reflected light readings from bleached coral so the hyperspectral imagery data could be calibrated to the in situ reflectance readings for an accurate interpretation. The procedure also requires that special geometric and atmospheric correction techniques are used to recalculate hyperspectral image pixels with data values that correspond to reflectance from the precise locations of those pixels. The main disadvantage is that the nominal spatial resolution of hyperspectral data is necessarily lower in order to maintain an acceptable signal-to-noise ratio. This condition is a function of fewer available photons in the narrow hyperspectral bands to interact with a sensor's detector elements. The consequence of this condition is very little image resolution to aid in the visual interpretation of the coastal environment. Lee and Carder (2005) recognized that a more cost-effective multispectral sensor is preferred over hyperspectral imagery when evaluating the major properties of coastal or shallow-water environments. Because digital bathymetry does not retain these disadvantages, there is an opportunity to investigate the possibilities of auto-classification procedures.

Unsupervised Isocluster Autoclassification Benefits and Limitations

The efficiency of unsupervised pixel clustering depends on the visual properties of the image being classified. Color ramped bathymetric DEMs generally do not offer a specific enough delineation of spectral signatures across the image. That is, if too many pixels with similar spectral properties are detected in multivariate space within the color ramped DEM, then the autoclassification cannot selectively assign different color values to represent specific seafloor landforms. Because the color ramped DEM does not offer enough spectral contrast and there is no cognitive intervention when running an unsupervised isocluster autoclassification, when an increased number of classes is used, each class will inadvertently include a cluster of pixels that carry the same value. When the computer runs the unsupervised isocluster autoclassification and detects the same pixel-valued clusters throughout the color ramped DEM, the same representative color is applied universally.

A potential solution to this problem is to use a greater range of hues in the color ramp applied to the DEM. Although the range of hues selected for a color ramp is arbitrary, the range of hues should be selected for overall visual appearance and arranged by depth. By using a greater range of hues in a color ramp that are keyed to bathymetric variations, it is possible to increase the number of classes in an unsupervised isocluster autoclassification.

Interactive Supervised Autoclassification Benefits and Limitations

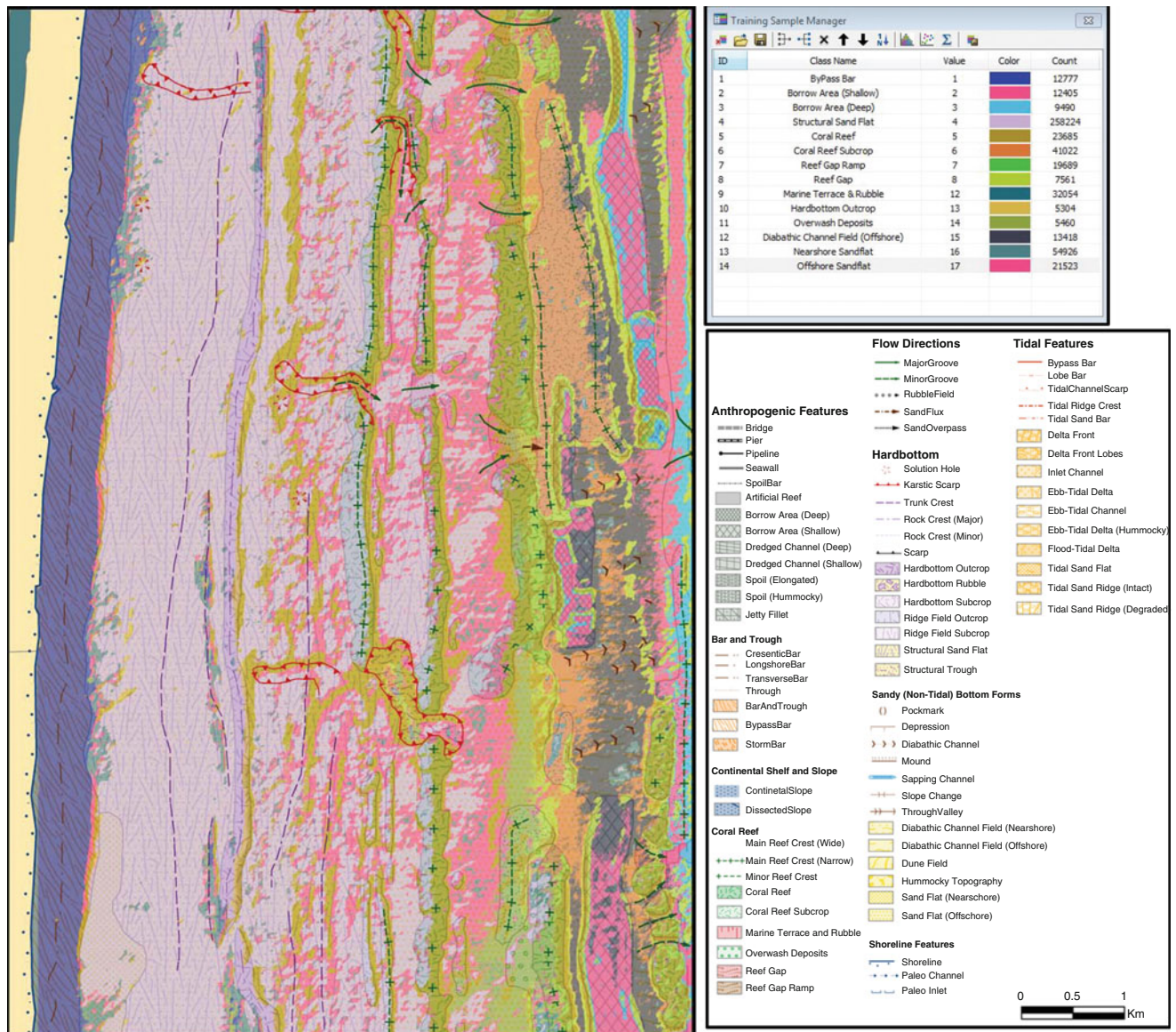
Increased efficiency can be expected when applying the interactive supervised autoclassification method across the color ramped DEM. This is because a cognitive element is introduced to supplement the autoclassification algorithm. That cognitive element is presented in the form of analyst delineated training sites, which enclose specific spectral signatures to teach the autoclassification software how to interpret the remaining pixels in the image. This greatly differs from the unsupervised isocluster autoclassification because the analysis is no longer solely reliant on the undefined distribution of pixel values within the color ramped DEM. Instead, individual geomorphological units are taught to the machine by associating very specific pixels hues in the form of training sites. However, the same limitation of the visual properties in the color ramped DEM that was seen with the unsupervised isocluster autoclassification still persists when applying the interactive supervised autoclassification methodology. For example, because the range of hues in the DEM color ramp shows minimal contrast throughout the image, certain geomorphological units are not easily delineated

from one another, even with the provided training sites to guide the autoclassification process. Regardless of this limitation, the interactive supervised autoclassification provides efficiency in the number of seafloor unit boundaries that could never be achieved with cognitive, hand-drawn cartography.

Cognitive Versus Autoclassification Methods

Cognitive mapping is an historical method of approximating spatial distribution patterns on the ground. These efforts were severely limited in the marine environment, and it was not until the advent of remote sensing techniques during World War II that it was possible to conduct regional submarine mapping. Seismic methods, single beam sonar, sidescan sonar, and multibeam sonar techniques provided great advances in the recognition of submarine seascapes, but it was not until the arrival of LIDAR in the 1960s that it was possible to acquire digital bathymetric data at sufficient resolution to recognize discrete submarine landforms on continental shelves. With the ability to create colorized DEMs from digital bathymetric data, it becomes possible to visualize seafloor features as never before seen. The cognitive maps derived from digital bathymetry provide an interpreted picture of seafloor geomorphology, necessarily generalized to a level that is dictated by the hand-drawing agility of the cartographer. Because these kinds of maps are hand-drawn and based on visual interpretation, it does not make them any more or less useful than other kinds of maps. The quality of the cognitive maps depends on the skill of the interpretation from the cartographer. That is to say, the more experienced and qualified the mapper, the more accurate the results. Because researchers without proper qualification and experience find it impossible to produce useful products, a common “solution” is to turn to automated classifications. Unfortunately, a new problem arises because an automated classification produces a map that differentiates the seafloor into different units or classes based solely on pixel arrangement. The key is to know what those units actually represent and where in the seascape they are correctly represented. Those units, however, can only be properly interpreted by those with appropriate training and experience in geomorphology. This oversight, a major pitfall that is associated with many automated classifications, limits the usefulness of auto-classed maps of seafloor geomorphological features.

On the other hand, supervised machine classifications can be very useful and informative, as described by Erdey-Heydorn (2008) and Rozenstein and Karnieli (2011). As in the analogy of a painting, much critical time and effort is required to prepare the canvas compared with the time spent actually painting. The same is true for the interactive supervised autoclassification, as careful consideration must be given to the selection of training sites. Although training



Coastal Seafloor Geomorphological Features, Classification, Fig. 6 This montage was created by digitally overlying an interactive supervised autoclassification on top of a cognitive (manually interpreted) map of shelf geomorphology based on digital bathymetry in a LADS format. Increased resolution is provided in the overlaid autoclassification which gives better discrimination (more details) of offshore sandflats on

top of hard-bottom (pink areas), diabathic channel fields (gray colored areas), coral reef subcrop (orange colored areas), and coral reefs and reef gaps (light green colored areas). Somewhat problematic are some light green areas for reef gaps that may be confused with several other morphological units, such as coral reefs, coral reef subcrop, and overwash deposits. (Source: Finkl and Makowski 2015)

sites can be selected to represent different kinds of variability in the survey area, they are perhaps more valuable if selected to better define known geomorphological features. Such a procedure requires a priori knowledge of the seafloor, which can be obtained from cognitive observation and mapping.

Both techniques in question, cognitive mapping and auto-classification, can be shown with great advantage when used in concert with one another. Each complements the other to help produce a more accurate, comprehensive product. That is, a cognitive mapping product refined by a software-driven autoclassification algorithm, further details the intricacies of

seafloor spatial distribution patterns in the form of digital bathymetry. The cognitive map provides a comparative control, or reference, for the interactive supervised auto-classification. Both products are useful in their own regard but acquire greater usefulness when blended into a composite or montage (Fig. 6).

Both processes, autoclassification and cognitive mapping, have their own sets of constraints that limit approximations of reality in the form of the kinds of maps that are produced in GIS (see for example Benedet and Finkl 2003; Greene et al. 2005; Finkl and Makowski 2015). Each approach has value or

merit, and greater insight into seafloor features can be achieved by applying the composite approach that melds the two disparate methodologies.

Conclusion

Classification of coastal seafloor geomorphological features can be achieved either through a cognitively derived interpretation or an autoclassificatory algorithm run by geospatial software. The interactive supervised autoclassification with expert-derived training sites provides the most discriminative map when compared with cognitive interpretations. Because of this general correspondence, it is possible to overlay the results of interactive supervised autoclassifications on top of cognitively derived units, thereby merging the two interpretations into a composite that highlights the advantages of both methodologies. Modern georeferenced software packages, such as ArcGIS, contain the appropriate type of tool kit extensions for applying a montage process in the production of a computer refined cognitively derived map. By applying both processes in tandem, a more efficient classification effort is achieved.

Cross-References

- [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
- [Beach and Nearshore Instrumentation](#)
- [Beach Profile](#)
- [Coastal Modeling](#)
- [Geodesy](#)
- [Geographical Coastal Zonality](#)
- [Holocene Coastal Geomorphology](#)
- [Nearshore Geomorphological Mapping](#)
- [Nearshore Sediment Transport Measurement](#)
- [Photogrammetry](#)
- [Remote Sensing of Coastal Environments](#)
- [Shoreline and Coastal Terrain Mapping](#)

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Coastal Sedimentary Facies

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Introduction

Coastal sedimentary environments include a variety of physical and biological processes, which act on coastal sediment to produce associations of composition, texture, primary sedimentary structures, and fossils. These associations constitute sedimentary facies, lithologically distinct, genetically related components of a sedimentary system, or environmental facies, whereby a set of specific environmental processes imparts a distinctive character to a sediment. Coastal sedimentary facies provide the signature of specific types of coastal deposits and facilitate their identification in the geologic record.

This entry describes some of the more common clastic sedimentary facies associated with open coasts and coastal embayments. For a review of carbonate coastal facies, the reader is referred to Demicco and Hardie (1995). The discussion here also does not address facies of high-latitude coasts or low-latitude coasts dominated by mangrove swamps (see Hill et al. 1995; and Cobb and Cecil 1993, respectively). Many of the facies described here form in both marine and lacustrine settings, although tidally influenced facies are restricted to the marine environment.

For more detailed information on coastal sedimentary facies, the reader is referred to the excellent summaries provided by Reineck and Singh (1973), Davis Jr. (1985), Walker and James (1992), Galloway and Hobday (1996), and Reading (1996).

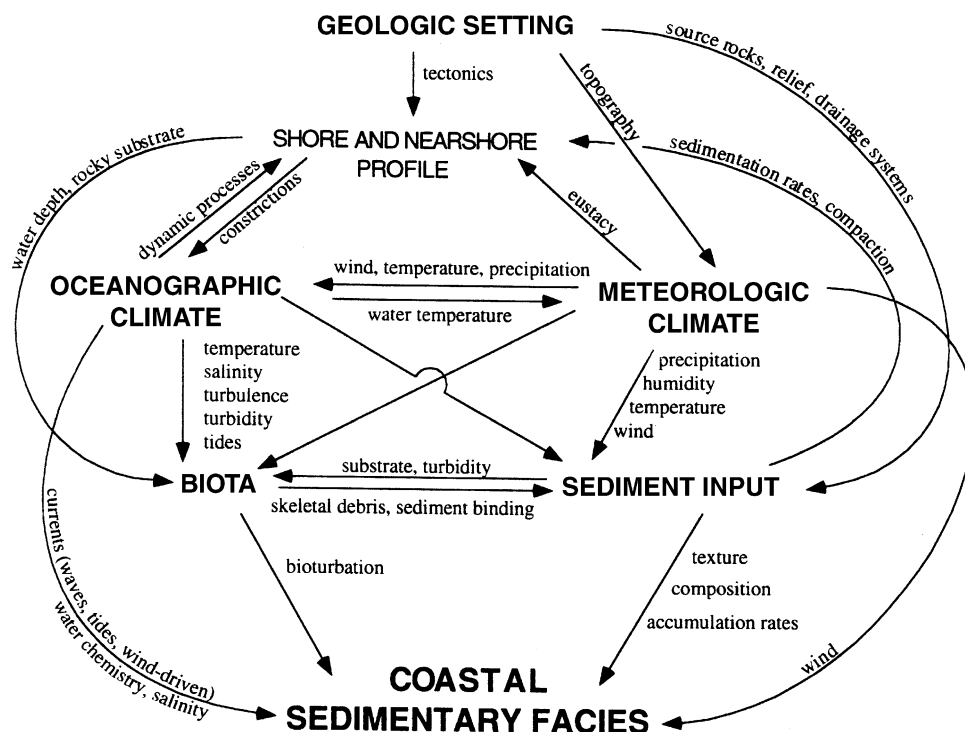
Coastal Sedimentary Processes

A complex array of processes influence coastal sedimentary facies (Fig. 1). Of these the most important are waves, tides, and biogenic processes. Sediment input is also critical to the facies character (grain size) and to the nature of the preserved deposit (sedimentation rate).

Waves may exist as “seas,” driven by local winds, or as “swell,” generated by distant storms. Swell tends to have longer period and to influence the seabed to greater depths

Coastal Sedimentary Facies,

Fig. 1 Interacting environmental factors and resulting processes that control the development of coastal facies (Modified from Clifton and Hunter 1982)



than do local sea waves. “High-energy” coasts are likely to be dominated by swell. “Low-energy” coasts receive smaller everyday waves, but can experience very large waves during storms.

Waves move sediment by two mechanisms. The passage of waves induces a back and forth orbital motion at the bottom. In water that is deep relative to the size of the wave, this motion is symmetrical in velocity and duration of flow. In shallower water, the motion becomes asymmetric, with short relatively strong landward flow under the crest of a wave and a more prolonged, weaker, seaward flow under the trough (for the conditions under which this asymmetric flow occurs, see the “Stokes wave” field in Komar 1976, Figs. 3–17). The orbital velocity asymmetry is important for promoting the onshore transport of coarser material and bed load in general.

Waves also, as they encounter the shore, generate sustained flow in the form of shore-parallel longshore currents and seaward-directed rip currents. These currents, which form nearshore circulation cells, not only transport sediment, but also shape the nearshore morphology into bars and troughs.

Tides influence sedimentary facies in two ways. The rhythmic rise and fall of the sea changes water depth and locally exposes extensive intertidal flats. It also generates tidal currents that flow alternately landward (flood tides) and seaward (ebb tides). In morphologically complex bays, either the flood or ebb tide is likely to predominate at any one location.

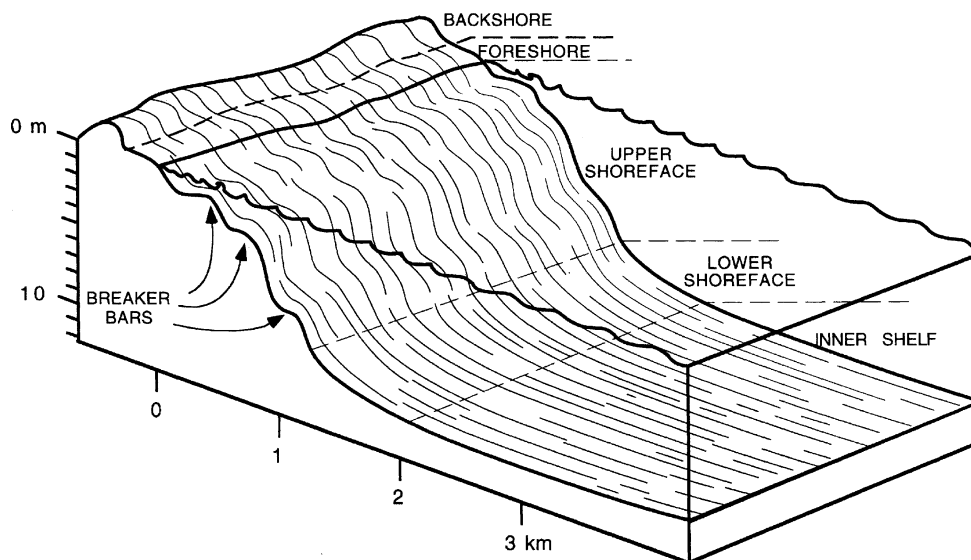
The response of sediment to these fluid motions depends in large part on its textural nature, which is determined partly by local dynamic processes and partly by what is available to the system. Grain size is also a major influence on the rate and

style of bioturbation, which combined with the frequency and intensity of physical processes determine the degree of preservation of physical structures. Bays typically contain abundant silt and clay, which are borne in suspension or move in small aggregates (particularly as fecal pellets) along the bottom. Between the tidal ebb and flood, at high and low water, the flow diminishes, allowing the settling of suspended sediment. This sediment may immediately accumulate on the bottom or it may be concentrated into a dense, highly concentrated fluid that hugs the bottom before deposition.

Biogenic processes are important in the development of coastal facies. Biogenic effects include bioturbation, the generation of distinctive traces, contribution of shell detritus, the fixing of suspended fine sediment by filter feeding and vegetative baffling, and the binding and bioerosion of the substrate. These processes depend on many factors, but faunal activities are particularly sensitive to variations in salinity. Faunal traces in brackish environments tend to be simpler, less diverse, and more likely to be dominated by a single ichnogenus (Pemberton and Wightman 1992) than those made in oceanic salinities.

Sedimentary Facies of Wave-Dominated Open Coasts

Waves are most important on open coasts, where the tides mostly only raise and lower the water level. A fundamental aspect of most open clastic coasts is the *shoreface* (Fig. 2), a relatively steeply inclined (1:200) ramp that extends offshore

Coastal Sedimentary Facies,**Fig. 2** Typical morphologic zonation of a wave-dominated open coast. Considerable vertical exaggeration

from the intertidal *foreshore* zone and merges with a nearly flat (1:2000) inner shelf or basin floor. Its origin is discussed by Niederoda and Swift (1981). The shoreface-shelf transition lies at depths that range from 2 m to 30 m. On prograding coasts, where the sedimentation rate is commonly sufficient to allow for a wave-induced equilibrium profile, the base of the shoreface rarely exceeds a water depth of 10 or so meters.

The *upper shoreface* is equivalent to the nearshore zone, where wave-generated currents predominate. Upper shoreface facies are generally the coarsest part of the coastal system and are characterized by high-angle trough cross-bedding, that is directed parallel to the shoreline (in response to longshore currents) or offshore (in response to rip currents). Modern medium- to coarse-grained nearshore zones are dominated during fair-weather by medium-scale lunate megaripples (dunes) that face and migrate in an onshore direction. Cross-bedding produced by these structures seems rarely preserved, however, in ancient upper shoreface deposits, which seem dominated by storm effects. If gravel is present, it typically is well sorted into discrete beds and layers, and gravel and shells commonly form erosional lags within the deposit. Small, shore-normal gravel-filled gutter casts exist at the base of some gravel beds. Burrowing is common and is dominated by the *Skolithos* ichnofacies, characterized by mostly vertical cylindrical and U-shaped burrows, the dwelling structures of suspension feeders or passive carnivores (Pemberton et al. 1992). The trace *Macaronichnus* is common in fine sand of high-energy coastal deposits and feeding pits of rays or other vertebrates may occur. The thickness of upper shoreface deposits generally ranges from 1 to 8 m, depending on the intensity of wave energy.

A *lower shoreface* facies is identifiable in most pebbly shoreface deposits. This facies, 1–3 m thick, consists of

small pebbles of relatively uniform size in a matrix of well sorted fine to very fine-grained sand. The pebbles occur as scattered clasts, in small clusters, or as thin discontinuous stringers 1–2 pebbles thick. The associated sand may be laminated or structureless, and is indistinguishable from inner shelf sand. The facies results from a combination of offshore transport from the upper shoreface during storms and onshore transport of most of this material by the shoaling waves in the aftermath of a storm. Winnowed by the waves, the coarser material moves landward more slowly and some is trapped in burrows, swales, or other depressions on the seafloor. In the absence of pebbles, the upper shoreface facies may be the only recognizable shoreface component.

On transgressive coasts, *shoreface-attached ridges* extend obliquely into the sea. These ridges, composed of fine to coarse sand, are tens of kilometers long, several kilometers wide, and 5–15 m high. Cores through shoreface-attached ridges show that they consist of an upward-coarsening accumulation of storm-event beds (Rine et al. 1991). Some shoreface-attached ridges show evidence of both storm and tidal influences in their internal structures.

The *beach foreshore* facies lie within an intertidal zone subject to the swash and backwash of the waves. Foreshore sands are typically very well sorted and characterized by planar lamination. Foreshore deposits are generally 1–2 m thick, and if the grain size permits, show an upward-fining. Individual lamina may show inverse grading. Placers of heavy minerals define the upper part of some foreshore deposits. Small vertical burrows and the trace *Macaronichnus* typify the ichnologic signature of the foreshore.

Nonmarine coastal facies of the open coast include eolian dunes, characterized by large-scale landward dipping foresets and backshore facies of wind/tidal flats. The character of the *backshore* facies depends on the specific setting and the

amount of vegetative growth. Backshore sand is commonly fine and well sorted; layers of pebbles and coarser sand are introduced during floods, and layers of muddy and/or peaty sediment can develop under wetter conditions. Structures such as climbing adhesion ripples or vertebrate footprints document subaerial exposure. Roots are common, and substantial coals may directly overlie the foreshore facies in some ancient deposits.

Mudflats dominate some open coasts proximal to, or down-current from, the mouths of large mud-bearing streams (see Augustinius 1989). In settings such as the coast of Surinam, inshore fluid mud maintains its presence by frictionally reducing the energy of incoming waves. Episodically, mud deposition ceases, either during storms or because the mud moves along the coast in discrete waves, and sandy beaches develop. A return to muddy deposition isolates the sandy sediment in shore-parallel ridges or *cheniers*. Cheniers on the coast of Surinam are about 2 m high and contain a combination of steeply and gently landward dipping cross-beds or, where fine-grained, gently landward and seaward-dipping cross-beds. Cheniers adjacent to the Mississippi River contain shell aggregates concentrated by storm waves. Mud in the associated flats may be laminated to structureless; mudflats covered by salt marshes will contain root or rhizome structures.

Sedimentary Facies of Deltaic Coasts

Deltas form where sediment discharged at the mouth of rivers exceeds the capacity of marine processes to redistribute it fully. As sites of substantial sediment accumulation, deltas form a volumetrically important part of the sedimentary record. Deltas fall into three end-member types depending on whether they are dominated by rivers, waves, or tides. River-dominated deltas are lobate and commonly quite muddy; the modern Mississippi River delta is a prime example. Wave-dominated deltas have a classic deltoid shape and are more likely to be sandy; the Niger River delta is a commonly cited example. Tide-dominated deltas are less common, and tend to be muddy or sandy and highly embayed. The Fly River delta of Papua New Guinea is a primary example.

Distributary channels, which develop on delta plains where a primary fluvial channel diverges into a series of straight or sinuous distributaries, are common to all types of delta. A balance of fluvial discharge and tidal range determines the extent of salt water intrusion into the lower reaches of the distributary and the degree of tidal influence. Distributary channel floors are erosional and typically littered with lags of mudclasts and/or plant detritus. Three-dimensional subaqueous sand dunes in the thalweg produce trough cross-bedding. Higher on the channel banks the sediment becomes finer and is likely to be dominated by ripple bedding.

Reversals of cross-bedding direction will occur where tidal range is sufficiently great. The nature of burrowing depends on salinity ranges in the distributary. If salinity is low or highly variable, the burrow types are likely to be greatly restricted.

The deposits of the distributary channels typically display an upward coarsening, and may pass upward into muddy levee deposits where complete. Clinoforms may mark the upper part of a distributary channel fill. Cross-sections of the deposits formed by relatively straight distributary channels show a deep central channel flanked by broader lateral extensions ("wings") near the top.

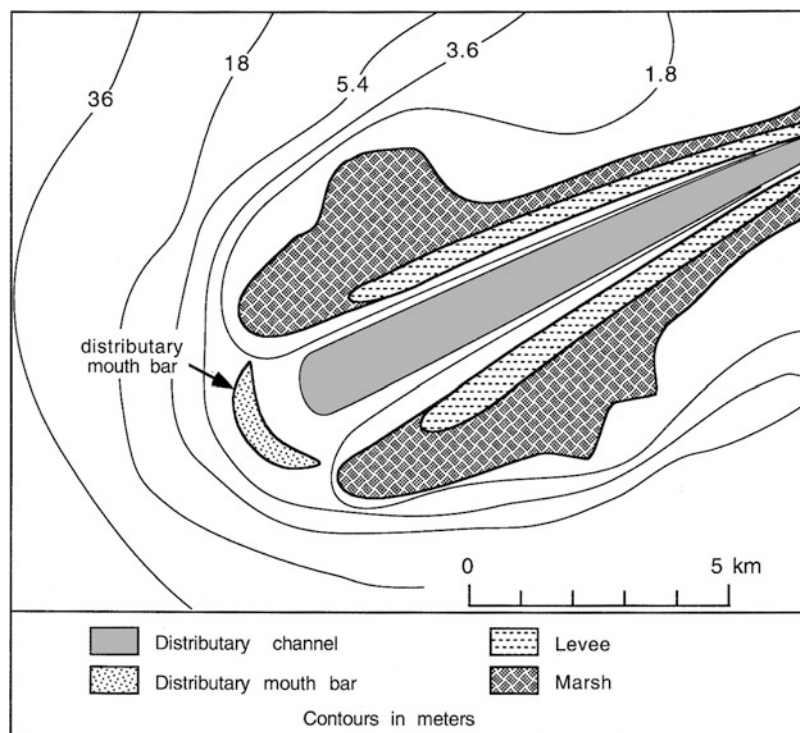
The banks of the distributary channels are capped by *levees*, ridges composed of fine-grained sediment at either side of a channel. Levee deposition occurs during submergence associated with major floods. Internal structures of levees include ripple lamination and small-scale cross-bedding, roots, and in some cases bioturbation.

Deltaic bays are associated with lobate (typically river-dominated) deltas and represent settings of low energy in which the prevailing sedimentation is from suspension, including much organic matter. Episodic influxes of silt and fine sand during floods produce coarser laminae, which may be well preserved in brackish settings where faunal reworking is diminished. The shallower parts of deltaic bays commonly are somewhat sandier and may display evidence of wave reworking in the form of hummocky cross-stratification. Plant growth along the edge of a bay may inhibit the development of normal beaches.

Crevasse splays and deltas accumulate where floodwaters breach a levee and carry fine sand and silt into an adjacent deltaic lake or bay. The resulting lobate deposits become progressively thinner and finer with distance into the bay. Internal sedimentary structures include ripple- and planar-lamination, climbing ripples, trough cross-bedding, and mud drapes. Graded beds, and, near the breach, erosional surfaces, define the deposits of separate floods. Dewatering or other deformational structures are fairly common in these rapidly deposited sands. The upper part of individual flood deposits may be bioturbated, where salinities permit. Splays may amalgamate into coarsening-up crevasse deltas, which resemble other small deltas.

Distributary mouth bars develop where distributary channels of a river-dominated delta debouch into a basin (Fig. 3). They consist of coarsening-up accumulations of fine-grained or silty sand. Deposition occurs mostly during flood stage, but the resulting deposit can be partly reworked by waves or tidal currents. Internal structures include ripple, climbing ripple, and planar-lamination. Flaser bedding may be present, as well concentrations of wood or finely divided organic detritus.

Beyond the actual mouth of a distributary lies the *delta front*, an area dominated by flood deposition and possible reworking by waves. Delta front deposits of river-dominated

Coastal Sedimentary Facies,**Fig. 3** Morphologic elements and facies at the mouth of a river-dominated delta lobe

deltas typically consist of interbedded fine to very fine grained sand and mud. Ripple lamination is a common internal structure of the sands, although storm waves can also generate hummocky cross-bedding in this environment. Some graded sand beds appear to be deposited by sediment gravity flows, and deformational structures are common. Bioturbation may be intense to limited depending on the volume of freshwater discharged into this setting. Delta front deposits grade down into *prodelta* muds that form the deeper-water foundation of a delta. Facies on the fronts of wave-dominated deltas may be indistinguishable from those of an open-coast wave-dominated shore.

Tide-dominated deltas have received less study than river-dominated deltas. Here, tidal influence on river flow may extend up-river well beyond the distributary system. Many of the facies (tidal channels and flats) associated with tide-dominated deltas are similar to those discussed in the section on embayment facies. Tidal *sand ridges*, 10 km or more long, several kilometers wide and as much as 10 m high, are common in the funnel-shaped mouths of tide-dominated deltas. Where studied, these ridges display an upward coarsening from interbedded sand and mud to cross-bedded sand. Mud drapes and flaser bedding are common.

Fan deltas are a special type of delta that exists where an alluvial fan builds into a marine or lacustrine basin. Alluvial fans are constructed by a combination of fluvial processes and episodic debris flows; because they typically lie proximal to highlands, alluvial fans are generally composed of coarse sediment. Where waves are too small to redistribute this

sediment, it accumulates on the shoreline in the form of a *Gilbert delta*, with steep, nearly planar, foresets that extend to the floor of the adjacent basin. Such foresets may define a deltaic unit tens of meters thick. Sediment slides down the foresets in grain flows, in which the coarser material outruns the finer component. As a consequence, Gilbert deltas are likely to show an overall upward fining from their bases. Individual foreset layers, however, tend to display an upward coarsening normal to the stratification in response to the grain flow process that created them.

Gilbert deltas form in settings other than fan deltas, wherever coarse sediment is introduced into a basin with relatively low wave energy. Braided streams carrying a coarse sediment load commonly create Gilbert deltas where they encounter a shore. Alluvial fans differ from braid plains by containing debris flow deposits composed of unsorted layers of sand, mud, and gravel with little internal structure. Subaerial debris flows on fan deltas flows can extend across the strand line and compose much of the subaqueous part of the delta.

Sedimentary Facies of Coastal Embayments

Coastal embayments are bodies of water somewhat restricted from the open sea. They include river mouths, drowned river valleys, and lagoons or other bodies of saline water isolated from the open sea by a barrier island or spit. Estuaries are semi-restricted bodies of water in which river discharge

measurably dilutes the salinity. Lagoons may have no significant freshwater input.

Embayments are subject to a complex interplay of waves, tides, and river discharge. Of these processes, tides typically predominate. Generally, the limited wind fetch in embayments precludes the development of large waves. River discharge, an important factor in many embayments, need not be present. Dalrymple et al. (1992) provide a useful model of estuarine facies that recognizes an inner area dominated by fluvial processes, a central area of mixed energy in which both river flow and tidal currents are significant, and an outer area dominated by waves and tides.

Within embayments, specific sedimentary environments produce distinctive facies. In addition to the factors noted above, bay facies depend on the degree of subaerial exposure provided by the tides. Subtidal facies are never exposed, intertidal facies are exposed by the astronomical tides, and supratidal facies are inundated only during coincidence of high astronomical and high meteorological (storm surge or river flood) tides. This section addresses the most common sedimentary facies formed in coast embayments.

Tidal basins as used here are subtidal parts of a bay, which lack discrete tidal channels. As a consequence, tidal currents tend to be feeble, and the sediment largely undisturbed by physical processes. The substrate here is typically fine-grained and dominated by biogenic processes. Commonly the sediment is totally bioturbated. Mollusk shells are common, either as clasts or in situ. During periods of minimal sedimentation, articulated and disarticulated shells can form a pavement on the bay floor leading to discrete shell layers in the resulting deposit.

Tidal channels provide a focus for tidal flow within a bay. Their size ranges from tens of meters to kilometers across and meters to tens of meters deep. Their substrate ranges from gravel to mud depending on the nature of river influence and the texture of sediment available to reworking by tidal currents. Typically, tidal channels cut into older sediment or bedrock, and their floors are littered with coarse detritus:

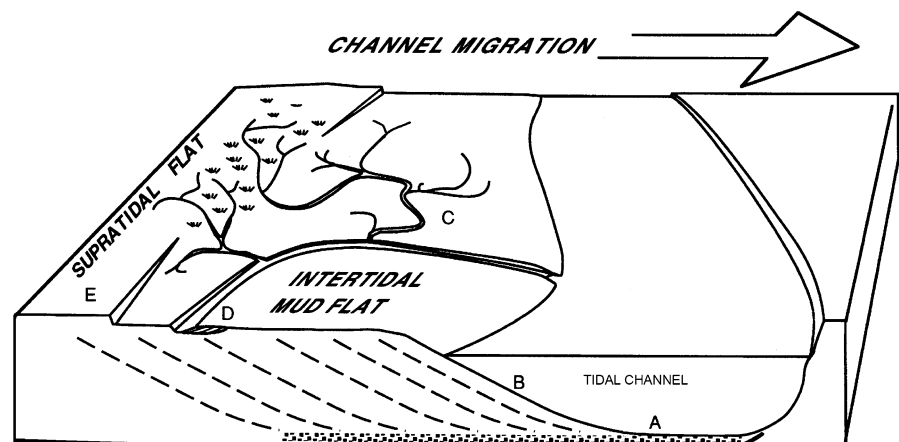
shells, pieces of wood, mud clasts, and, today, human debris. Where tidal channels erode into a relatively consolidated muddy substrate, organisms, such as decapods may bore into the older material. The filling of these borings by channel-floor detritus generates a *Glossifungites* ichnofacies (MacEachern et al. 1992). Less cohesive substrates are likely to become colonized by bivalves, producing a thin interval of in situ shells below the base of a channel. Where sand is abundant, two- and/or three- dimensional dunes typically occupy a channel floor, producing trough or tabular cross-bedding. Otherwise, channel-floor deposits are likely to be thoroughly bioturbated, owing to slow rates of sediment accumulation.

Because tidal channels tend to migrate, their banks are either erosional or accretionary (Fig. 4). Accretionary tidal channel banks form a volumetrically important part of many ancient bay fills. Because accretionary banks are sites of deposition, internal physical structures are likely to be preserved. The nature of accretionary bank stratification depends greatly on sediment texture. Sandy banks are likely to preserve cross- bedding or ripple lamination. With increasing mud, the bedding style changes progressively from wavy (scattered clay drapes) to mixed (roughly equal quantities of sand beds and clay drapes) to flaser (mud encasing sand ripples) to laminated (mud with scattered sand laminae). Rhythmic stratification is common, owing to the regularity of tides or seasonal processes. Tidal channel bank deposits generally (but not invariably) show an upward fining. The migration of tidal channel banks produces large clinoforms that are evident in good exposures.

Tidal sand bars are ridges of sand within tidal channels. Such bars may be entirely subtidal or include both subtidal and intertidal components. Subtidal bars are 1–15 km long, have length to width ratios greater than three, and have a relief of as much as 20 m (Dalrymple and Rhodes 1995). The larger ones approach the size of tidal sand ridges of large open embayments and seaways (see Houbolt 1968). Sand waves (dunes) typically mantle the bar surfaces and internal

Coastal Sedimentary Facies,

Fig. 4 Components and stacking process of a mesotidal muddy tidal channel system



structure consists of trough and tabular cross-bedding. Where bars are less active, their internal structure is likely to consist of sets of cross-strata separated by bioturbated intervals, owing to low rates of aggradation and bedform migration. Larger bars commonly separate areas of flood- and ebbdominance. Like the tidal sand ridges in the North Sea, these ridges commonly are asymmetric whereby the steeper side faces in the direction of dominant sand transport. As a consequence the dip cross-bedding foresets is unidirectional in the direction of subordinate sand transport. Where tidal sand bars extend into the intertidal range, the vertical sequence likely to result if the bar were fully preserved consists of a lower subtidal and intertidal cross-bedded interval overlain by tide flat and salt marsh deposits (Dalrymple et al. 1990).

Intertidal flats are sandy or muddy areas of low relief that are alternately exposed and inundated by astronomical tides. Sedimentation rates are low and bioturbation predominates unless physical or chemical. A. Tidal channel floor composed of bioturbated shell and other lag; Numerous burrows/boring into firm ground beneath the channel floor form a *Glossifungites* ichnofacies; B. Tidal channel accretionary bank, composed of stratified mud and sand; C. Tide flat, underlain by bioturbated mud/or sand; D. Accretionary bank of runoff channel, composed of finely interlaminated sand and mud rich in organic detritus. Low-angle mud across-stratification; E. Supratidal flat, inundated only during combinations of high spring tides and meteorologic tides, such as storm surges. Underlain by laminated very fine sand/silt and mud. Numerous root or rhizome structures.

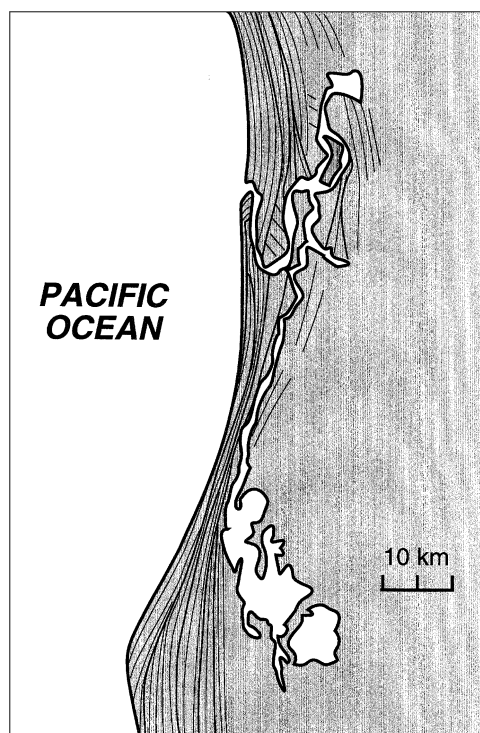
factors inhibit faunal colonization. An example is on flats exposed to very high tidal range (macrotidal flats). Here, the difference between high tide under neap and spring tidal conditions is sufficient to provide an upper flat that is only inundated during high-spring tides. This intermittent exposure precludes the development of a mature infauna, and stratification is preserved in the form of flaser, mixed, wavy, or lenticular bedding, depending on the proportion of sand and mud. Where mesotidal conditions (tidal range 2–4 m) obtain, the difference between high spring and neap tides is less and bioturbation prevails in nearly all of the flat sediment.

Most muddy tidal flats are crossed by narrow, dendritic *runoff channels* (tidal creeks), typically 1–2 m deep. Most runoff channels have an erosional and an accretionary bank. Sedimentation on the accretionary banks is very rapid and the sediment commonly displays laminae that reflect individual tides as well as neap and spring tidal cycles. Despite their meandering appearance, the rate of channel migration is slow, owing to the erosional resistance of the muddy substrate. Sediment that accumulates rapidly on accretionary banks is prone to slumping into the channel, where it is flushed out by the tidal currents. The limited channel migration produces muddy clinoforms or cross-strata, 1–2 m thick, characterized

by finely interlaminated sand, mud and organic detritus, and sporadic slump structures. Runoff channels on sandy tidal flats generally lack well-formed accretionary banks, and the associated distinctive facies.

Salt marshes are common on the upper part of many intertidal flats or marine delta plains (see Frey and Basan 1985). The substrate may be completely bioturbated or display faint rhythmic lamination. The organic content is typically high, and root or rhizome structures are typical. *Supratidal flats* lie higher still and are inundated only under combinations of high- astronomical and high-meteorological (river flood or storm surge) tides. Stratification, in the form of rhythmic alternations of silt and organic-rich mud, is typically crossed by numerous roots or rhizomes. Commonly the laminae form sets of several distinctive layers, probably deposited at successive high tides during a single meteorological event.

Bayhead deltas are common in estuarine bays. These may take the form of a small river delta, displaying a transition from muddy sediment in the most basinward, deeper parts to sandy sediment in the upper, river-dominated part. Fluvial or distributary channels cap the upper delta surface. The resulting deposit shows an upward coarsening. Evidence of tidal influences, such as tidal bundling or reversals in ripple-lamination or cross-bedding, can serve to distinguish this type of small delta from those produced by crevassing.



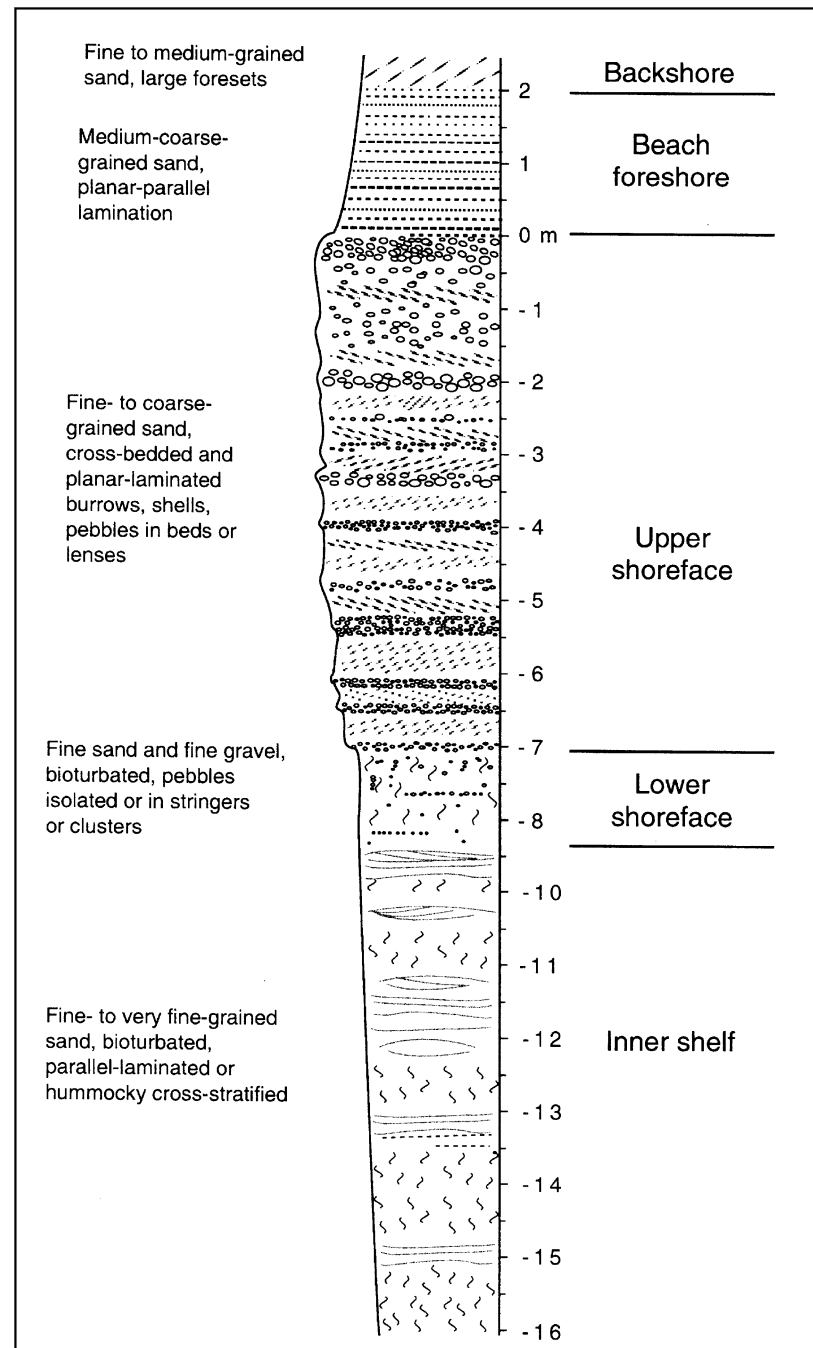
Coastal Sedimentary Facies, Fig. 5 Beach Ridges indicating progradation on the Nayarit coast of Mexico (After Curray et al. 1969). Each ridge marks a previous position of the coastline

In estuaries with a large tidal range, such as San Francisco Bay, the bay head delta comprises a complex of anastomosing channels without a clear deltaic form. In this setting, the bayhead delta consists of tidal channels that gradually transform landward into fluvial channels. The interface between salt and fresh water (head of the salt-water wedge) is typically very turbid and the site of rapid deposition from suspension. As with tide-dominated deltas on an open coast, the tidal influence on river flow may be expressed many kilometers upriver from the head of the salt-water wedge.

Washover fans are lobate accumulations of sand on the seaward side of lagoons, formed where storm waves breach a barrier and sweep sand into the adjacent lagoon (Leatherman et al. 1977). Bases are erosional, commonly marked by a shell lag. Internal structures include inversely graded planar lamination, ripple lamination, and climbing ripples, antidune stratification and, particularly at landward margins, landward-dipping cross-bedding. Tops of individual overwash accumulations can be bioturbated or contain rootlets.

Coastal Sedimentary Facies,

Fig. 6 Typical shoreface succession on a prograding pebbly, high-energy coast



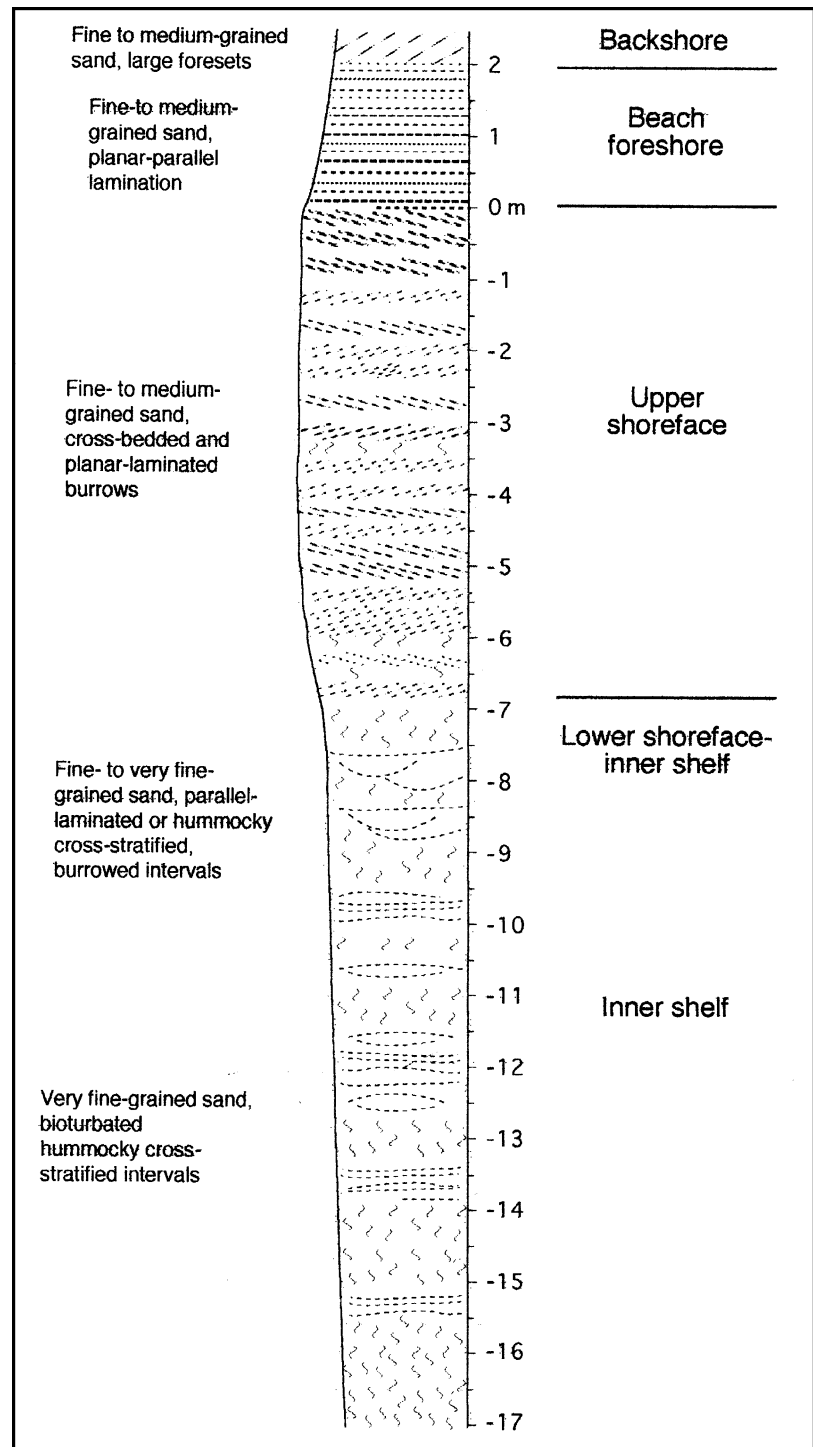
Sedimentary Facies of Inlet Areas

Tidal inlets form a fluid interface between bays and the open coast (see Boothroyd 1985). Inlets may range from simple channels a few meters deep and tens of meters wide to complexes of channels and shoals more than 10 km or more across wide at the mouths of large embayments. Individual

channels in this setting may be tens of meters deep and hundreds of meters wide. Inlet sediment is mostly sand, and occurs in a setting where tidal flow is likely to be at its highest. As a consequence, inlet channels are typically floored by two- and three-dimensional dunes. Individual channels may be dominated by either ebb or flood tide. Lag deposits of shells and or pebbles (where the adjacent shoreface is gravelly) are

Coastal Sedimentary Facies,

Fig. 7 Typical shoreface succession on a prograding finer-grained, non-pebbly, high-energy coast



generally present, and may contain species representative of both open coast and bay. Trough and tabular cross-bedding and abundant burrows characterizes the resulting deposit.

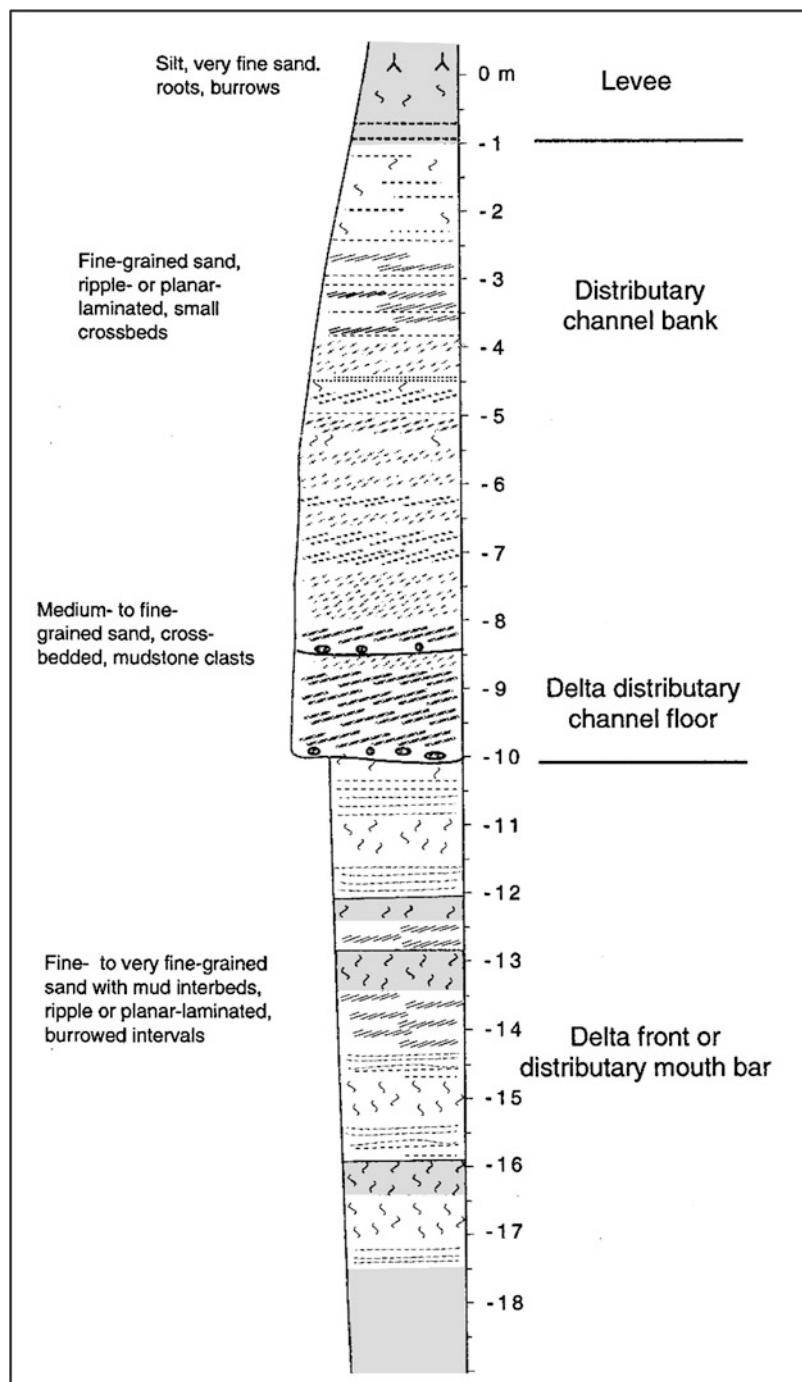
Small inlets tend to migrate laterally along the coast, particularly where wave-driven longshore sediment transport is significant. Larger inlets may be more stationary, but interior inlet channels may migrate rapidly. In Willapa Bay, Washington, for example, a primary inlet channel, 25 m deep and 600–700 m wide, has moved laterally more than a kilometer in the past 50 years. Inlet fades in such a setting are

likely to consist of a lower part composed of cross-bedded sand produced in the primary channel overlain by the deposits of smaller tidal channels and wave- and tide-influenced shoals. Channel migration can produce clino-forms, the thickness of which depends on the depth of the channel. The lateral extent of inlet deposits depends on the degree of inlet migration. Extensive migration will generate laterally extensive inlet facies within a coastal succession.

Lobate bodies of sand, *flood tidal deltas*, form sandy ramps on the bayward side of many inlets. Where the tidal inlet

Coastal Sedimentary Facies,

Fig. 8 Stratigraphic succession produced in the upper part of a prograding river-dominated delta



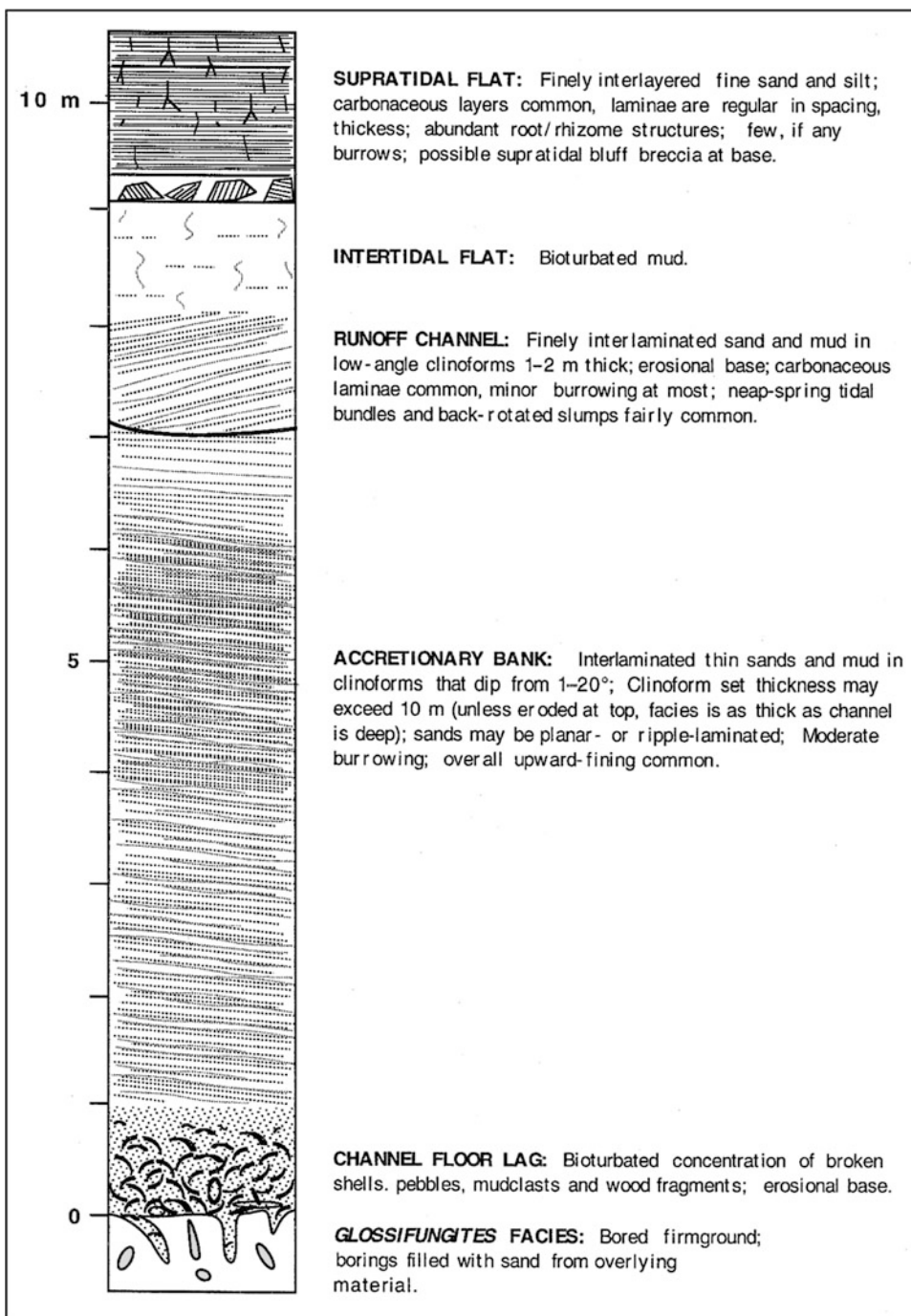
migrates, flood tidal deltas may compose laterally extensive sand bodies along the seaward side of a lagoon enclosed by a barrier island. Initially these deltas form as a set of lobes building into the bay, covered by straight to sinuous two-dimensional landward-facing dunes (Hine 1975). As they mature, tidal flow is concentrated into channels, ultimately producing a low, seaward-facing ramp dissected by broad flood-dominated tidal channels (Boothroyd 1985). The resulting deposit is dominated by landward-directed trough

and tabular cross-bedding and contains internal erosional surfaces capped by shell lags.

Ebb tidal deltas form on the seaward side of inlets, where they are subject to the influence of both waves and tides. Off large embayments, these deltas may be quite large, exceeding 10 km in width and extending 5 km or more into the adjacent sea. Off smaller inlets on a transgressive coast, the deltas may form the nucleus of shoreface attached ridges (McBride and Moslow 1991). Flood- and ebb-dominated tidal channels

Coastal Sedimentary Facies,

Fig. 9 Vertical facies succession produced by migrating muddy tidal channel system. Thickness of sequence depends on depth of tidal channel and degree of subsequent erosion of upper part of succession



cross the shoal constituting the delta: typically the ebb flow is concentrated in the deeper channels. The combination of flood tidal flow and wave swash can produce a generally landward orientation to bed-forms on the delta platform (Boothroyd 1985). Upward-fining storm beds with erosional bases are common, particularly in the shallower parts of an ebb tidal delta.

Preservation of Coastal Sedimentary facies and Resulting Facies Successions

Thick accumulations of coastal facies typically reflect a record of fluctuating sea level superimposed on a setting of overall tectonic subsidence. The shore response to sea-level change depends on the rate of sedimentation. In general, the influx of sediment at the shore promotes progradation, whereby the additional sediment forces the shore to shift basin-ward. Today, most prograding open coasts are characterized by a series of beach ridges, each marking an earlier position of the shore (Fig. 5). The resulting accumulation conforms to Walther's Law, which holds that a vertical arrangement of related facies conforms to their lateral distribution at the time of deposition. Progradational successions show an upward-shallowing, produced as shallower water facies encroach over their deeper water counterparts. On deltaic or other open coasts, the result typically generates an upward-coarsening interval. Progradation also occurs on a smaller scale where a bayhead delta builds into an estuary, or a crevasse delta builds into a deltaic bay. Figures 6, 7, 8 show examples of progradational successions.

Fining-up facies successions are generated within bays and on deltas by the lateral migration of tidal or delta distributary channels. In these cases, the shallower parts of the system

consist of intertidal flats or salt marshes, or, with distributary channels, levees. Figures 8 and 9 show examples of fining-up successions produced by migrating channels.

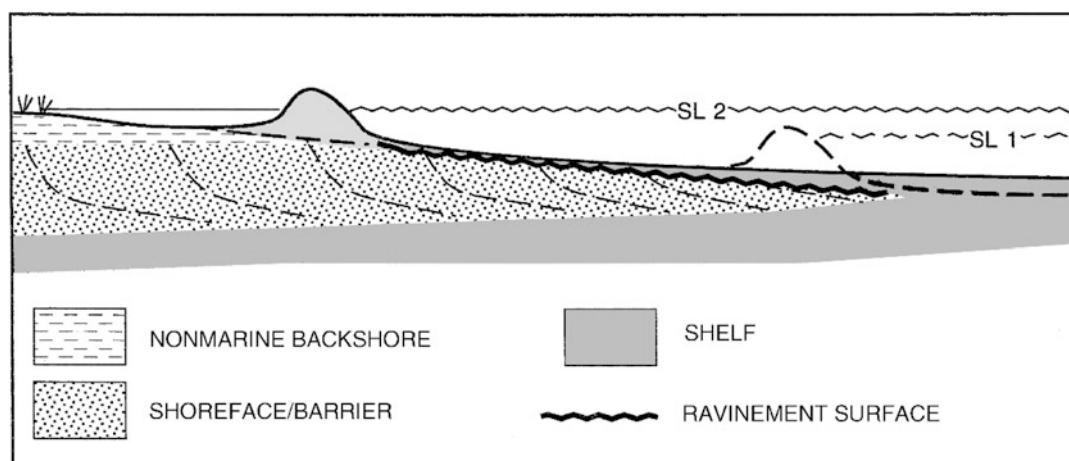
Although progradation is the most common circumstance of preservation of coastal sedimentary facies, not all facies are preserved. Those associated with features that stand in relief above other, more landward, features are likely to be destroyed as progradation progresses.

Breaker bars, for example, are common features of many upper shorefaces (Fig. 2). During progradation, the seaward migration of the bar troughs that lie to landward erases much, if not all, of the bar. Similarly, distributary mouth bars have a limited potential for preservation. During the seaward migration of a distributary mouth system, distributary mouth bars are likely to be eroded and cannibalized by the distributary channel that created them. If that channel avulses, however, the associated bar may be preserved as a feature with limited lateral extent.

Where relative sea level falls during progradation a "forced regression" ensues. Wave erosion associated with the readjustment to a new, lower base level can create a "regressive surface of marine erosion". The amount of sediment removed by this erosion depends on the rate and degree of base level fall and prevailing wave energy. Commonly the surface marks an abrupt change from shelf to shoreface deposits.

Progradation may occur during relative sea-level rise, if the rate of sedimentation is sufficiently high. But commonly, a rise in base level is associated with deposition in the feeder systems. With the resulting reduction of sediment contributed to the coast, marine transgression occurs, whereby the shoreline shifts landward.

During transgression, some facies are likely to be preserved and others lost. During lowstands that commonly



Coastal Sedimentary Facies, Fig. 10 Transgressive erosion caused by the lateral translation of the shoreface (see Bruun 1962). None of the transgressive coastal facies are preserved, and the transgression is

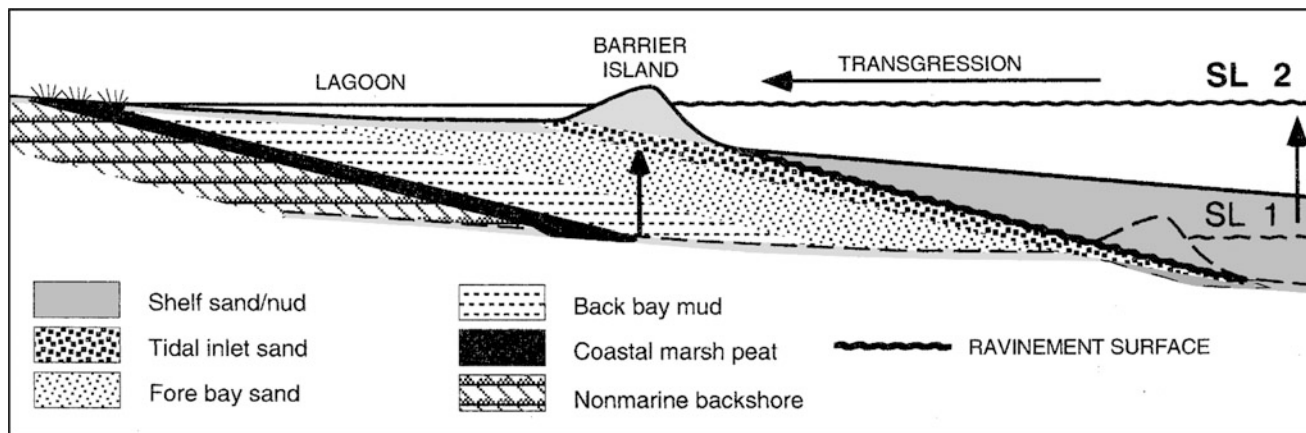
recorded by an erosional ("ravinement") surface overlain by deeper water typically shelf) deposition

precede transgression, coastal rivers incise valleys near the coastline. Estuarine facies deposited in these valleys at the lowstand or during the initial rise tend to be largely preserved through a transgression.

A different situation prevails on the open coast. The natural product of waves and coastal winds is a shoreface and beach capped by a foredune ridge. In most wave-dominated settings, enough sand exists that a small rise in relative sea level is met by an upbuilding of the shore to a topographically higher position. Where a sea-level rise

causes a reduction in sediment input to the coast, a basin forms behind the beach ridge and a barrier island is created. Most present-day coasts fronted by barrier islands are in a state of transgression.

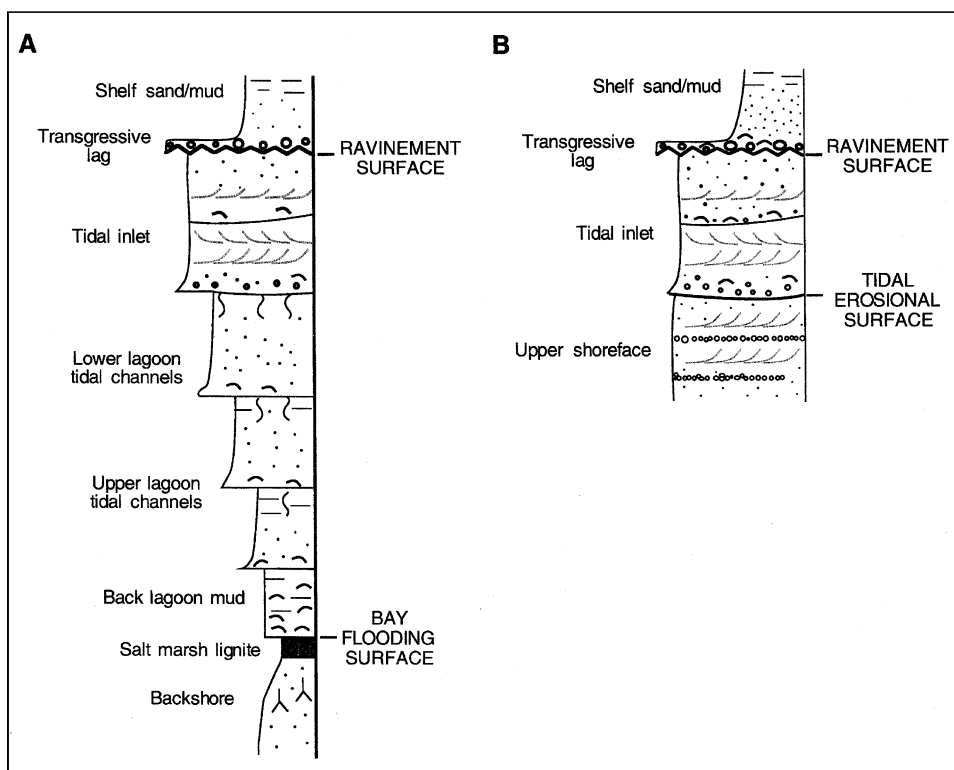
Barrier islands formed during transgression have a limited potential for preservation. The landward translation of a shoreface erases the deposits that lie landward from its base, leaving an erosional (*ravinement*) surface overlain by shelf facies (Fig. 10). Commonly this surface is the only preserved expression of a transgression.



Coastal Sedimentary Facies, Fig. 11 Transgressive succession produced in settings with rapid accommodation and adequate sedimentation. A lagoonal succession develops as the lagoonal facies are buried by

the landward-migrating barrier. It is capped by a “ravinement” surface overlain by shelf facies

Coastal Sedimentary Facies, Fig. 12 Vertical successions produced by preservation of transgressing lagoons. (a) High rate of accommodation. (b) Modest rate of accommodation



Under conditions of rapid accommodation and sufficient sediment supply, however, lagoonal facies are preserved (Fig. 11). These may occur in inverted succession, whereby the landward-most facies (lagoonal marshes and back-bay muds) are overlain by sandier deposits associated with the approach of the transgressing barrier and capped by tidal inlet facies just below the ravinement surface (Fig. 12a). The thickness and organization of these transgressive lagoonal deposits differs depending on the balance between accommodation and sedimentation. Although most are but a few meters thick, some can exceed 20 m. Most display a basal “bay-flooding surface” and a capping ravinement surface, although vigorous tidal currents and low rates of accommodation may in some places lead to the erosion of the lower part of the succession (Fig. 12b).

Cross-References

- [Barrier Island Landforms](#)
- [Bars](#)
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Coastal Soils

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Introduction

The occurrence of soil in coastal zones is not a simple matter to resolve. Summarized coastal zones form a global perspective that only briefly indicates potential occurrence of soil in relation to coastal landforms, maritime climates, drainage, vegetation, and time available for formation. Increased understanding of coastal soils is, however, only possible from the purview of specific examples that illustrate soil distribution patterns in terms of sequences based on soil-forming factors of topography (toposequences), time (chronosequences), climate change (climosequence), and biological factors (biosequence). The conceptual frameworks for the models were initially postulated by Jenny (1941) and subsequently enhanced by various other researchers (e.g., Butler 1959, 1967; Runge 1973; Huggett 1975). The value of these sequential paradigms is that they provide visualization of complex interrelationships in a framework that clarifies and elucidates coastal-soil transition inland from the shore.

The models are relatively uncomplicated in young (Holocene), more or less uniform parent materials such as those which may occur in small deltaic areas, mangrove swamps, coastal alluvium, and dunes. Complications arise along most coasts because coastal fringes reflect polygenetic development where exceedingly old parent materials and associated soils may occur juxtaposed to incipient pedogenesis. *Environmental complexity* in coastal areas where there is rise and fall of relative sea level, stillstands and diastems, increase and decrease of sediment supply, delta progradation and retrogradation, alluvial channel avulsion, development of deltaic superlobes, chenier formation, climatic change, and so forth, is manifested in weathering zones and soils that provide clues to evolutionary sequences as does the stratigraphy and nature of the included materials. Examples of depositional sequences are reported by Davis (1978), Milliman and Meade (1983), Liu and Walker (1989), Less and Lu (1992), Roberts and Coleman (1996), Saito et al. (2000), and Yatsko (2000), among others, for a variety of coastal settings. Dune systems, including large eolianites, which can be just as complicated as deltaic sequences, often provide stratigraphy with intercalated Paleosols (e.g., Fairbridge 1950; Cooper 1966; Denny and Owens 1979; Hearty and Kindler 1997). Pedogenic inertia and persistence (e.g., Huggett 1975; Finkl 1980) is another problem where older soils (which in turn may be initial parent material for soil forming under new

environmental conditions) may occur in geomorphically young coastal landscapes.

Thus, coastal soils *sensu stricto* are often regarded as those soils that are derived from marine or estuarine parent materials that have recently been exposed subaerially to pedogenesis viz. acid sulfate soils, etc. Others might be related to coastal dunes, beach ridges, chenier plains, salt marsh, or upland freshwater marsh. The soils mostly range in age from late Holocene to early Pleistocene, reflecting a chronosequence that is geologically very young. This restricted point of view excludes many coastal sectors where soils of great age may be found, especially in the intratropical zones where coastal erosion may expose intensely weathered deposits as coastal cliffs in laterite (e.g., Hays 1967; Petit 1985; Retallack 1990; Wang 2003). Coastal soils *sensu lato* include a very wide range of possibilities as almost any kind of soil may occur in coastal zones. Fortunately, the possibilities are not endless and there is some order to apparent chaos of haphazard occurrence. Some clues as to potentialities for soil distribution patterns, at myriametric scales, on coasts may be gleaned from the morphological classification of coasts on the basis of plate tectonics (e.g., Inman and Nordstrom 1971) where tectono-physiographic settings limit the scope of pedogenesis. Trailing continental margins provide large expanses of coastal plain sediments, as occurring along the eastern seaboard of the United States, south of Washington, DC, where there are thick weathering sequences composed of residual oxidized materials and soils developed in alluvial, estuarine, and marine deposits (e.g., Markewich et al. 1986). In the vicinity of major rivers, especially on the southern coastal plain of the United States, major floodplains and deltas give rise to organic-rich soils. The coastal plains of southern Florida give rise to Holocene calcareous soils on slightly higher elevations whereas, lower poorly drained sites contain a suite of organic and muck soils related to Everglades, sloughs, swales, and other karst depressional areas that became infilled with fine-grained materials mixed with organic materials following the Holocene transgression (e.g., Gleason et al. 1984; Finkl 1994, 1995; Lodge 1994). Leading continental margins, such as that occurring along the mountainous and (strike-slip) faulted North American west coast where an oceanic plate is subducted below an overriding continental plate, provide a completely different set of tectono-physiographic conditions for soil development. Uplifted marine terraces contain soils of different ages, depending on the rates of uplift and number of terraces along the coast. Tsunami deposits, which are preserved in protected areas, provide allochthonous arenaceous parent materials for soils (Clague et al. 1999) that date back to the penultimate great Cascadia earthquake 1,000 years ago. Coastal valleys in Oregon and Washington often retain interesting stratigraphy, where earthquake-induced subsidence leads to the burial of alluvial and marsh soil profiles and the

development of younger soils in the new parent materials above (e.g., Nelson and Kashima 1993). Pedogenic environments along these kinds of tectonic coasts where there are active seismotectonic regimes, as in Greece (Maroukian et al. 2000) and throughout much of the Mediterranean region and elsewhere, are much more dynamic than the quiescent trailing margins.

Finally, there is the ultimately confounding situation where former coastlines are now stranded long distances from present shores. Many of these rocky shores (e.g., Johnson 1992; Libbey and Johnson 1997) provide parent materials for subsequent soil development (neopedogenesis of soil materials), but these occurrences can hardly be considered “coastal” in that pedogenesis is now unrelated to coastal factors of soil formation, other than coastal-marine parent material. Coastal Paleosols are, however, usually associated with modern as well as ancient coastlines whether at present sea level, emergent, or submerged. Emergent examples are best known and have been extensively studied along with Quaternary geomorphology and stratigraphy, for example, in Papua New Guinea (e.g., Bloom and Yonekura 1985), Barbados (e.g., Radtke 1989; Schellmann and Radtke 2001), Bermuda (e.g., Hearty et al. 1992), Bahama Islands (e.g., Hearty and Kindler 1997), Point Peron in Western Australia (Fairbridge 1950), Chile and Argentina in South America (e.g., Rutter et al. 1989), and many other coastal areas too numerous to mention. Notable within the present purview of coastal soils, however, is the study by Schellmann and Radtke (2003) who employ morphologic, pedostratigraphic, and chronostratigraphic assessments of marine terraces along the Patagonian Atlantic coast. Many of these studies were used for reconstruction of local sea-level history, including evidence in southwestern Western Australia for a higher relative mean sea level (MSL) during the Holocene (Fairbridge 1961; Playford and Leech 1977), but as discussed by Searle and Woods (1986) evidence for the ± 3 m oscillations proposed by Fairbridge can be elusive. Still other ancient shorelines, instead of experiencing terrestrial sequestration from the sea, are found as drowned relicts. Examples are reported from Hawaii (Fletcher and Sherman 1995), but there is no mention of included drowned soils.

Deltaic plains, such as the Mississippi River delta deposits, provide similar examples where constructive (alluvial accumulation, build up and soil formation) and destructive cycles (e.g., subsidence and drowning) present complex pedogenic environments that respond to fluvialdeltaic processes and longer-term base-level changes (Roberts and Coleman 1996). Callaway et al. (1997), studying sedimentary accretion rates on low tidal-amplitude sites along the Gulf of Mexico, found evidence that contravened previous hypotheses concerning the relationship between tidal range and marsh vertical accretion rates. They found that there was little correlation between mineral and organic matter accumulation

rates, with average organic matter accumulation rates showing less variation compared to mineral matter accumulation. There thus may be limits to annual rates of organic matter accumulation. Sediment addition is, however, clearly required for maintenance of coastal marsh soils (e.g., DeLaune et al. 1987, 1990).

There are many different soil classifications in use today throughout the world (see discussions in Finkl 1981), but none feature-specific sections on coastal soils per se. Rather than being perceived as a disadvantage, this situation is advantageous because soils occurring in coastal environments are hierarchically related throughout comprehensive soil classification systems and placed within continental or international schemes that facilitate reference and comprehension. Prominent among the major soil classifications are efforts by American, Australian, Canadian, French, German, Brazilian, and Russian soil researchers but the system employed here is the one adopted by the Food and Agricultural Organization FAO of UNESCO. The FAO system (FAO-UNESCO-ISRIC 1988; FAO 1991) is employed because it is specifically designed as an international system and because it provides a comprehensive inventory of world soil resources in the form of electronic databanks and maps. Although this is the primary organizing or categorical basis for discussing coastal soils, classes in other systems are sometimes equally valid (Spaargaren 2000) and they are sometimes referred to for ease of reference and also to avoid unnecessary complications that would arise from correlation with other systems. Local soil classification schemes often reflect regional or cultural bias in their geography and therefore international efforts are encouraged (e.g., Finkl 1982a, b). As already appreciated by many coastal researchers, they must be ambidextrous when dealing with coastal soil morphology, genesis, and classification. Unless otherwise stated, classificatory units referred to here belong to the international FAO system of soil classification, but similar and related terminology is also employed from the US soil classification system as applied in *Soil Taxonomy* (Soil Survey Staff 1975) and keys (Soil Survey Staff 1992). The following descriptions are based on FAO major soil groups.

Histosols

The major soil groups of Histosols (from Gr. *Histos*, tissue) include a wide variety of peat and muck soils that range from moss peats of the boreal tundra, moss peats, reeds/sedge peats, forest peats of the temperate zone, and mangrove and swamp forest peats of the humid tropics (Spaargaren 1994; Inubushi et al. 2003). Histosols are unlike all other soils in that they are formed in and by “organic soil material” with physical, chemical, and mechanical properties that differ strongly from those of mineral soil materials. They develop in

conditions where organic materials are produced by an adapted (climax) vegetation, and where biochemical decomposition of plant debris is retarded by low temperatures, persistent waterlogging, extreme acidity, oligotrophy and/or the presence of high levels of electrolytes from organic toxins. Organic soil material contains more than 12% organic carbon by weight (20% OM). The diagnostic horizons for Histosols are the histic and folic horizons. Other diagnostic horizons used to separate soil units in Histosols are termed sulfuric, sulfidic, and salic. The combination of specific environmental conditions, composition of the organic soil material, and degree of decomposition produces six soil units that include: Gelic Histosols (permafrost occurs within 200 cm of the soil surface), Thionic Histosols (there is a sulfidic or sulfuric horizon starting within 125 cm of the surface), Salic Histosols (there is a Salic horizon starting within 50 cm of the surface), Folic Histosols (folic horizon present), Fibric Histosols (more than two-thirds of the organic soil material consists of recognizable plant material), and Haplic Histosols (other units).

Histosols form where the rate of accumulation of plant debris exceeds the rate of its decay (e.g., Nyman et al. 1993). In practice, Histosols occur where microbial decomposition of plant debris is impaired by: cold temperatures, excessive wetness, salts or other toxins, and severe aridity or oligotrophy. Histosols under the permanent influence of groundwater, unless artificially drained (e.g., “low moor peats”) occur in low-lying positions in fluvial, lacustrine, and marine landscapes, mainly in temperate regions but also to limited extents in the tropics. Other soils occurring in the same environment include Fluvisols, Gleysols and, in coastal regions, Solonchaks (e.g., adjacent to coastal mangrove peats). In lacustrine landforms and in coastal marsh embayments, Histosols may be associated with Vertisols, such as that occurring on the Texas Gulf Coastal Plain (e.g., Stiles et al. 2003). Oligotrophy and prolonged wetness are primarily accountable for the low decay of organic debris. In the wet tropics (mainly the Indonesian region surrounding the Sunda Flat), their formation is conditioned by high rates of organic matter production from the climax forest vegetation. Change in use of tropical peatlands can affect the decay rates of organic matter and production of methane and nitrous oxide gases viz. conversion from secondary forest peatland to paddy fields, as reported for Kalimantan in Indonesia (Inubushi et al. 2003). In addition to landuse change as a forcing function for pedogenic succession, similar migration of soil properties are associated with tidal inundation along transgressive coastal areas such as around the Chesapeake Bay on the eastern US coastal plain. Hussein and Rabenhorst (2001) report, for example, pedogenic transformation of Ultisols to Alfisols and eventually to Histosols in response to sea-level rise, based on changes in sodium adsorption ratios and exchangeable sodium percentages. Salinization and alkalization

processes as a function of tidal inundation frequency across low-lying coastal landscapes produce lateral changes in pedogenic properties that ultimately relate to different profile characteristics and classification of soils. Lateral pedological linkages exist with a variety of other soil groups, including Podzols, Fluvisols, Gleysols, Cambisol, and Regosols.

The global extent of Histosols is about 275 million ha (FAO 1991) (Table 1), roughly half of which are located in the Arctic zone, onethird in temperate lowlands, and one-sixth in tropical lowlands (Spaargaren 1994). An example of the latter case is the subtropical Florida Everglades, which constitute the largest single body of Histosols in the world (Stephens 1956, cited in Amador and Jones 1995). A large area of cold region organic soils occurs on the southern and southwestern subArctic margins of Hudson Bay (Manitoba, Ontario, Quebec) and around Great Bear Lake (Northwest Territories) in Canada. Here, Cryic Fibrisols (with frozen subsurface) occupy about 171,715 km² of coastal plain (Clayton et al. 1977).

In lower latitude coastal areas, continually rising sea level has caused brackish or saline waters to engulf drowned river valleys (rias) or to extend over formerly upland soils. This has led to the formation of coastal marsh Histosols in several geomorphic settings (Darmody and Foss 1979). As sea levels continue to rise, marsh margins are pushed landward and the organic materials continue to thicken so that the older and deeper Histosols generally exist nearer the open water (e.g., Nyman et al. 1993; Rabenhorst 1997). Although distant from high-latitude glacial activity, coastal Histosols at lower latitudes were impacted by glacio-eustatism (e.g., Rabenhorst and Swanson 2000). During the glacial maximum (~20,000 year BP) when large quantities of water were bound in glacial ice, eustatic sea level was at least 150 m below present MSL. Deglaciation cycles caused sea level to rise at such a rapid rate (~10–20 mm a⁻¹) that initially vegetation could not colonize tidal regions. Approximately 3,000–5,000 years ago, sea-level rise slowed so that marsh vegetation became established and organic parent materials began to accumulate (Bloom and Stuvier 1963; Redfield 1972). As sea level rose, organic materials accumulated in Histosols so that coastal marshes and mangroves generally accreted at about the same rate. Peat accretion in coastal areas presently ranges from 3 to 8 mm a⁻¹, which is much higher than in non-coastal regions. The highest rates of sea-level rise may, however, be too great for marsh systems to maintain and some areas suffer marsh loss and degradation (e.g., Huiskes 1990; Nyman et al. 1993; Boesch et al. 1994) where Histosols are eroded, degraded (e.g., Mendelssohn and McKee 1988) or drowned. Loss may also occur in response to subsidence (primarily as microbial mineralization of soil organic matter), such as in areas of the Florida Everglades where drainage of coastal wetlands (Everglades marsh) (Finkl 1995) has exposed organic soils subaerially (Stevens et al. 1984) and

Coastal Soils, Table 1 Main kinds of soils occurring in coastal areas, based on a world reference base for soil resources, indication the different types of soil cover in relation to landscape features and associated linkages with related soils

Soil group (FAO) ^a	Area (×10 ⁶ ha) ^b	Major subdivisions ^c	Associated soils ^d	Lateral linkages or intergrades ^e	Landscape position, climate, drainage, topography, or parent materials
Histosols	275	Gelic, Thionic, Salic, Follic, Fibric, Haplic Histosols	Fluvisols, Gleysols, Solonchaks, Vertisols	Podzols, Fluvisols, Gleysols, Cambisols, Regosols	Low-lying positions in fluvial, lacustrine, and marine landscapes. Solonchaks often adjacent to coastal mangrove peats
Anthrosols	NA	Hydragric, Irragic, Cumulic, Hortic Anthrosols	Border most of the major soil groups where soils have been strongly influenced by man	Includes coastal archaeological sites	Old cultivates sites, excluding paddy soils, on coastal plains, river valleys, deltas, etc.
Leptosols	1,655	Lithic, Cryic, Skletic, Rendic, Mollic, Umbric, Dystric, Eutric Leptosols	Regosol, Cambisol, Podzol, Calcisol, Luvisol, Histosol, Anthrosol	Wide range of intergrades	Tropics to polar tundra, coastal lowlands to mountains
Cryosols	1,800	Histic, Thixotropic Cryosols	Histosols, Gleysols, Stagnosols, Cambisols	Gelic or cryic units of Histosols, Gleysols, Podzols, Planosols, Glossisols, Stagnosols, Cambisols	Arctic coastal plains of NE Eurasia, Alaska, Yukon, and NW territories of Canada; islands of Siberian, Beringian, and North American sectors of the Arctic Ocean; coastal Greenland
Fluvisols	350	Thionic, Salic, Vertic, Mollic, Calcaric, Umbric, Dystric, Eutric	Cambisols, Regosols, Arenosols, Leptosols, Gleysols, Solonchaks	Arenosols, Cambisols, Fluvic Gleysols, Salic Fluvisols (intergrade to Solonchaks)	Large deltas and fans viz., Ganges, Mekong, Mississippi, Niger, Po, Rhine; major river floodplains; coastal barriers; tidal flats
Solonchaks	260–340	Gleyic, Stagnic, Mollic, Gypsic, Calcic, Sodid, Haplic Solonchaks	Chlorida and sulfate soils	Salic Fluvisols, Thionic Fluvisol-Solonchak intergrades	Coastal salt lakes, lagoons, pans; chloride facies-soils of marine origin under mangrove; neutral chlorido-sulphate soils-of marine origin but enriched by sedimentary gypsum deposited by rivers; acid sulphate soils-soils of marine origin under mangrove as in Gambia, Senegal, Guinee Bissau
Gleysols	720	Cryic, Thionic, Plinthic, Arenic, Mollic, Umbric, Fluvic, Calcic, Haplic Gleysols	Leptosols, Vertisols, Fluvisols, Solonchaks, Stagnosols, Histosols, Regosols	Histi-Mollic, Histi-Umbric Gleysols, Fluvic Gleysols (intergrades to Fluvisols)	Sandy to loamy (sometimes clayey) textural classes of parent material on alluvial, fluvial, fluvio-glacial deposits in riverine or coastal areas. Common in depressional and organic-rich coastal areas; for example, Arctic coastal plains, coastal Alaska; Florida Everglades
Podzols	485	Gelic, Gleyic, Stagnic, Humic, Duric, Umbric, Cambic, Haplic	Histosols, Gleysols, Cryosols, Cambisols, Ferralsols, Planosols, Glossisols	Podzol-Histosol-Gleysol sequences, Cambisol-Podzol sequence	Sandy deposits affected by illuviation and cheluviation, boreal climatic zone, coniferous vegetal cover, temperate and tropical wet climatic areas. For example, NE and NW coastal Canada, Scandinavia, NW coastal Russia; coastal plain of Florida; southwestern Australia; coastal Indonesia
Sesquisols	60	Petric, Aeris, Albic, Stagnic, Humic, Eutric, Haplic Sesquisols	Ferralsols, Alisols, Acrisols, Lixisols	Found in association with Gleysols and Stagnosols in areas conditioned by hydromorphy; with Ferralsols, Alisols, Acrisols, and Lixisols in better drained positions in the landscape	Usually ancient landsurfaces truncated by modern coastal erosion or in close proximity to the sea; older coastal plains; for example, western Africa, western India, Mekong catchment,

(continued)

Coastal Soils, Table 1 (continued)

Soil group (FAO) ^a	Area ($\times 10^6$ ha) ^b	Major subdivisions ^c	Associated soils ^d	Lateral linkages or intergrades ^e	Landscape position, climate, drainage, topography, or parent materials
					northern and southwestern Australia, eastern Amazon region
Ferralsols	750	Humic, Geric, Gibbisc, Lixic, Rhodic, Eutric, Plinthic, Gleyic, Haplic Ferralsols	Cambisols, Acrisols, Nitisols, Gleysols, Planosols, Arenosols, Sesquisols	Intergrades to Cambisols, Acrisols, Nitisols, Gleysols, Planosols, Arenosols, Sesquisols	Mid- to end-Tertiary peneplains; for example, Brazilian Shield, French Guyana, Senegal, Madagascar, northern and SW Australia; extreme tropical and subtropical weathering
Planosols	130	Gelic, Vertic, Histic, Mollic, Umbric, Dystric, Eutric Planosols	Vertisols, Acrisols, Luvisols	Stagnic Solonetz, Albic Sesquisols, Stagnic Gleysols; related soils with abrupt textural change but insufficiently wet to show reduction by surface water would fall into Luvisols, Lixisols, Alisols, Acrisols	Low-lying, nearly level fluvial or marine terraces, depressional lands on coastal plains, including wetland rice paddies; most extensive in climes with a strongly seasonal variation in rainfall that results in intermittent saturation and reduction of upper soil horizons
Solonetz	135	Gleyic, Stagnic, Salic, Albic, Mollic, Gypsic, Calcic, Haplic Solonetz	Linked through sodic units to Vertisols, Solonchaks, Gleysols, Calcisols	Salic Solonetz, Sodic Solonchaks; through sodic properties in Vertisols, Solonchaks, Gleysols, and Calcisols; with Gleysols and Stagnosols through the Gleyic and Stagnic Solonetz soil unit	Solonetz landscapes are governed by microrelief, waterlogging at the surface, and salinity in the profile; lake terraces, river deltas, depressions along liman coasts; Rio de la Plata deltaic plain in Argentina; coastal South Australia and coastal southeastern Western Australia
Calcisols	800	Petric, Luvic, Sodic, Cambic, and Haplic Calcisols	Vertisols, Solonchaks, Gleysols, Solonetz, Luvisols	Calcic Gleysols, Calcic Gypsicisols, Calcic Luvisol	Arid coastal regions viz. Baja, California; Patagonia and Argentine Pampa; coastal Atacama; Namibia; coastal northern Africa; Nullarbor Plain in Western Australia and coastal plain in South Australia
Alisols	>100	Plinthic, Gleyic, Humic, Vertic, Ferric, Chromic, Luvic, Haplic Alisols	Ferralsols, Nitisols, Acrisols, Lixisols, Vertisols, Cambisols	Vertic Alisols, Alic Nitisols, Alic Acrisols; facies of other highly weathered soil units in intertropical regions	Coastal plain in southern USA; parts of intertropical coastal India, Africa, and northern Australia
Acrisols	>900	Plinthic, Gleyic, Humic, Arenic, Albic, Ferric, Haplic	Lixisols, Luvisols, Alisols, Nitisols, Glossisols	Planosols, Lixic Ferralsols, Arenosols, Regosols, Cambisols	Ancient shield landscapes, in tropical regions, that are truncated by coastal plains or marine erosion; coastal plain is southeastern USA, Caribbean Central America coastal lowlands, Orinoco deltaic plain, western Africa, Southeastern Asia, Indonesia
Luvisols	650	Gleyic, Albic, Vertic, Calcic, Ferric, Chromic, Dystric, and Haplic	Alisols, Lixisols, Acrisols, Nitisols, Glossisols, Solonetz	Luvic Chernozems, Luvic Gypsicisols, Luvic Phaeozems, Luvic Calcisols, Luvic Arenosols	Humid to subhumid temperate regions of western Europe (Belgium, The Netherlands, coastal Germany), parts of the Mediterranean region, southern Australia (extreme southwestern Western Australia, central South Australia), eastern Indian coastal regions, Sri Lanka
Lixisols	435	Plinthic, Gleyic, Humic, Arenic, Albic, Ferric, Haplic	Nitisols, Alisols, Acrisols, Luvisols, Glossisols	Ferralsols, Regosols, Cambisols, Luvisols	On stable shield landscapes in tropical regions; in seasonally dry tropical, subtropical, and warm temperate regions, on Pleistocene and older surfaces (including Alluvial fans); coastal savanna regions

(continued)

Coastal Soils, Table 1 (continued)

Soil group (FAO) ^a	Area ($\times 10^6$ ha) ^b	Major subdivisions ^c	Associated soils ^d	Lateral linkages or intergrades ^e	Landscape position, climate, drainage, topography, or parent materials
Cambisols	1,500	Gelic, Gleyic, Vertic, Fluvisol, Mollic, Calcaric, Ferralic, Dystric, Chromic, Eutric Cambisols	Luvisols, Ahsols, Acrisols, Lixisols, Nitisols, Ferralsols	Regosols, Leptosols, Gleysols, Fluvisols, Acrisols, Sesquisols	Islands (Puerto Rico, Madagascar, New Zealand, Papua-New Guinea); Atlantic Spain and Portugal; Mediterranean Spain, France, western Italy, Turkey; southeastern Australia
Arenosols	900	Leptic, Albic Protic, Gypsic, Calcaric, Luvic, Ferralic, Cambic, Haplic	Regosols, Leptosols, Cambisols, Gleysols, Fluvisols	Acrisols, Lixisols, Ferralsol, Planosols, Histosols, Anthrosols, Solonchaks, Regosols, Leptosols, Calcisols	Dry tropical and temperate interior regions, but some coastal areas in western Africa, western Madagascar, and northwestern Australia
Regosols	260	Gelic, Anthropic, Tephric, Gypsic, Calcaric, Dystric, Eutric	Arenosols, Leptosols, Gypsisols, Ferralsols, Cryosols	Found in all landscapes of the world; intergrades and extragrades may develop into many other soils, depending on the most important soil forming factor(s)	Shallow, weakly developed mineral soils in the initial stages of soil formation; all coastal landscapes

Notes: Compiled from Spaargaren (1994)

^aBased on the FAO World Reference Base (WRB) major soil groups, compiled and revised from the FAO/UNESCO International Reference Base for Soil Classification and the Legend of the Soil Map of the World (FAO-UNESCO-ISRIC 1988)

^bTotal area (in millions of hectares) is based on estimated WRB worldwide occurrence of the soil group; coastal occurrences are a smaller subset of unknown areal extent

^cMajor subdivisions that are proposed to separate the soil groups on the basis of diagnostic properties. Not all of the soil units listed occur in coastal areas as major soils, but they are listed for completeness, where they may be found in specialized environmental situations

^dOther soil groups that occur in the same general area or environment; soil groups were not differentiated on the basis of climate in order to keep the number of units manageable. Differences in topography, drainage, and other diagnostic properties and materials result in a range of different soil groups that occur in close geographic proximity to each other

^eLateral linkages are commonly intergrades that merge soil units in time and space. These units also reflect the spatial integration of soil properties in the coastal landscape where differences in the factors of soil formation (i.e., topography or relief, time, drainage or hydrology, organics, parent or initial materials) may lead to rapid lateral changes in soil units. Relief, drainage (or lack thereof, salt spray, or influence of tidal waters), parent materials (terrestrial versus marine sediments), and time (especially in regard to young soils forming in marine sediments) are particularly relevant to coastal soil distribution patterns

their surface elevations may decrease by as much as -2.5 cm a^{-1} .

The Florida Everglades is the largest continuous sedge moor outside of the Pleistocene glaciated area of the United States (Gleason et al. 1984). The expansive coastal wetlands and freshwater marsh of southern Florida are a result of the very slow relative rise of sea level during the past 3,200 years (average rate of 4 cm per century) (Spackman et al. 1966; Wanless et al. 1994). A large number of coastal sediment bodies in southern Florida were initiated or stabilized between 3,300 and 3,000 year BP and again approximately between 2,300 and 2,500 year BP (Wanless et al. 1994). The origin and development of those coastal and shallow-marine carbonate and clastic sediment bodies is dependent on the interactive roles and timing of sea-level changes, climatic fluctuations, storm influences, vegetative colonization, preexisting topography, and provision of detrital (input or recycled) siliciclastic and carbonate sediment. Of these, the rate of sea-level rise was the primary control on the clastic, carbonate, and organic sediment bodies of southern Florida.

As one of the largest marshes in the world, the Everglades peat deposits are commonly differentiated into subtypes

based on compositional variation, thickness, and degree of weathering. Water is required for the formation of peat soil because it restricts atmospheric oxygen from the soil; without oxygen, microorganisms cannot decompose dead marsh plants as fast as they accumulate. Ever-thicker deposits of peat are deposited until an elevation is reached where the surface dries enough that oxygen-dependent decay (the opposing aerobic process), or fire, prevents accumulation (Lodge 1994). Freshwater peats of the Everglades merge southwards with mangrove peats on the tidal plain fronting Florida Bay. Occupying more than 400 ha of marshland, the Everglades Peat forms from degraded saw grass in long-hydroperiod habits. It is the most abundant peat occurring in the Everglades Basin. The Loxahatchee Peat, covering about 29,000 ha, occurs in deeper marsh areas and contains less saw grass remains and more evidence of slough vegetation such as the roots, rhizomes, leaves, and seeds of water lilies. Very poorly drained organic soils of the Everglades usually have a surface layer of black muck (sapric material-organic material that is too finely divided for identification of plant remains) that overlies black or reddish-brown muck subsoils that in turn rest on sand, limestone bedrock, or other materials

(McCollum et al. 1978). The term *muck* refers to highly disintegrated peat soils that are composed of more mineral than organic matter (see discussion in Brown et al. 1990). *Marl* is the product of periphyton. During the dry season, the organic material (dead algae) in the periphyton mass oxidizes, leaving the calcium carbonate particles as a light-colored soil. Marl is descriptively called *calclitic* mud and is the main soil of the short-hydroperiod wet prairie habitats near the margins of the southern Everglades where bedrock lies close to the surface (Lodge 1994). Marl and peat soils are opposites, requiring aerobic and anaerobic soil conditions, respectively. Peat cannot accumulate in shorter hydroperiod marshes where marl persists, and acid conditions within a peat soil dissolve marl and prevent its accumulation. However, there are large areas in the northern Everglades where a thick layer of marl underlies peat. This occurrence is evidence that the hydroperiod increased substantially in those areas as the Everglades ecosystem evolved. Much of the Everglades region is flooded during the summer rainy season, which is a natural process essential to the health of this wetland. Many areas of urban development, especially on the coastal plain of southeast Florida, occur on former Everglades (wetlands that were drained as part of reclamation programs) (Finkl 1995) where Histosols have been stripped away and building sites “demucked” and filled with dredged materials. The region is, however, prone to returning to its natural wetland condition during high rainfall events making localized sectors of the region susceptible to flooding (Finkl 2000).

Acid Sulfate Soils

Acid sulfate soils occupy an area of some 24 million ha worldwide where the topsoil is severely acid or will become so if drained (Ritsema et al. 2000). An equal area of these soils may be thinly covered by peat and nonsulfidic alluvium (van Mensvoort and Dent 1997). Acid sulfate soils are typically associated with mangrove forests that once covered 75% of the coastlines in tropic and subtropical countries (Quarto and Cissna 1997). Estimates of the extent and distribution of acid sulfate soils suffer, however, from the lack of field surveys, few reliable laboratory data, and from variable definition. More significant than their areal extent is their location where they are concentrated in otherwise densely settled coastal and floodplains, mostly in the tropics, where development pressures are intense with few suitable alternatives for expansion of farming, or urban or industrial development. Two-thirds of the known extent is in Vietnam (e.g., Mekong River Delta), Thailand, Indonesia, Malaysia, western Africa (Senegal, Guinea Bissau), northern Australia, and on coastal plains in southern China (e.g., Pearl River Delta). By far, the largest use of acid sulfate soils is for rice cultivation viz. the Balanta system in western Africa (van Ghent and Ukkerman

1993) where land is ridged annually to a height commensurate with the expected freshwater flood of the rainy season. The ridging and annual turning speedup removal of soluble toxins from the surface soil, but the system is declining because of the heavy labor demand and low financial returns.

Acid sulfate soils exhibit enormous spatial variability that is tied to the dynamic estuarine, deltaic, and flood plain environments in which they occur. Figure 1 shows typical colonization by mangroves into carbonate intertidal flats that will eventually become sites for accumulation of organic debris derived from the mangroves. Organic-rich muds will eventually develop and be accompanied by pedogenesis. Significant spatial variability occurs at myriametric scales from pyrite-plugged pores to the fine-grained soil matrix to infield, local, and regional soil patterns (Dent 1986). Remote sensing is widely used in tropical, tidal environments because surveys are arduous and difficult due to mosquitoes and crocodiles. Acid sulfate soils also exhibit significant temporal variability that is associated with their defining characteristics of acidity and related toxicities. Almost uniquely, acid sulfate soils export their problems in drainage and floodwaters.

Evident in acid sulfate soils are patterns related to sedimentary history viz. the common distinction between highly sulfidic, unripe clays or peat in backswamps and ripper, coarser textured soils, or even calcareous soils on levees and creek fillings. Soil patterns are not always visible at the surface because present landforms and vegetation may be only partly and indirectly related to the environment of sulfide accumulation. The area may, for example, have been buried by peat or nonsulfidic alluvium, fresh waters may have succeeded brackish, and new plant communities have replaced prior ones.

Sulfidic soils that oxidize and generate H_2SO_4 when drained are referred to as *potential acid sulfate soils* (Wang and Luo 2002). If they have been drained and are generating H_2SO_4 , they are called *raw acid sulfate soils*, but if they have passed through the acid-generating phase but remain severely acid, they are referred to as *ripe acid sulfate soils*. Sulfidic materials include marine and estuarine sands and clays, gyttja in brackish lakes and lagoons, and peats that originally formed in fresh-water but which have been inundated subsequently by brackish water. The common factors are: (1) supply of organic matter, (2) severely reducing conditions caused by continuous water logging, and (3) a supply of SO_4^{2-} , usually from tidewater that is reduced to sulfides by bacteria decomposing organic matter, and (4) a supply of Fe from the sediment for the accumulation of iron sulfides which make up the bulk of reduced S compounds. These conditions are fulfilled in tidal swamps and salt marshes where thick deposits of sulfidic clay have accumulated in concert with Holocene sea-level rise (e.g., Pons and van Breeman 1982; Dent and Pons 1995). As sedimentation raises the soil surface



Coastal Soils, Fig. 1 Initiation of mangrove foothold on intertidal marine platform on Eva's Cay, west (bank or leeward side) of Salt Pond on Long Island land, Bahamas. Carbonate muds and micritic sediments show initial phases of biochemical weathering as mangroves colonize the area. Accumulation of organic debris derived from the mangroves eventually results in dark coloration of chemically reduced muds. Mangrove soils often show light (thin bands) and dark (thick bands) lamination in the profile because pedogenesis is interrupted by

storm waves that deposit new unweathered materials on top of pedogenically altered materials. This mangrove mud is minimal as the stand is only just taking a foothold; older, well-developed mangrove forests such as that occurring in the world's largest estuarine delta (2,600 km²) along the Indian coast southeast of Calcutta in West Bengal as Sunderbans (beautiful forests) and many other tropical coastal locations feature thick organic-rich profiles.

above MSL, topsoil accumulates under better drainage as a nearly ripe, mottled layer that contains little or no sulfide. Sulfidic clays may be buried by peat or nonsulfidic alluvium where freshwater conditions succeed brackish water.

Reclamation and drainage of sulfidic soils causes dramatic changes. When insufficient carbonate is present to neutralize the H₂SO₄ generated by oxidation of sulfides, extreme acidity develops within weeks or months. Raw acid sulfate soils, characterized by a pH <3.5, contain jarosite but frequently display a black subsoil as some of the SO₄²⁻ generated by drainage is reduced to FeS deeper in the profile. Experiments on acid sulfate soils in the southern coastal regions of Guangdong, China, for example, show that the sulfur content of acid sulfate soils is 2–6 times higher than that of coastal Solonchaks and 3–9 times higher than that of paddy soil (Wang and Luo 2002). The potential acidity is 2–3 times higher than the actual acidity and the amount of lime required to correct the potential acid ranges from 143.5 t ha⁻¹ to 2,687.5 t ha⁻¹.

Acid sulfate soils cause on-site problems and also in adjacent areas with the latter attracting attention (Dent 1986). On-site problems include Al toxicity in drained soils, Fe and H₂S toxicity in flooded soils, salinity, and nutrient deficiencies. Engineering problems include: (1) corrosion of steel and concrete, (2) uneven subsidence, low bearing strength, and fissuring leading to excessive permeability of unripe soils, (3) blockage of drains and filters by

ochre, and (4) difficulties of establishing vegetation cover on earthworks and restored land. Off-site problems stem from drainage effluents, earthworks, excavations, and mines. The acid drainage waters carry Al released by acid weathering of soil minerals and heavy metals that are released by oxidation of sulfide minerals. Episodic release of toxic drainage waters may occur, for example, at the onset of the wet season after a period of low water table during which oxidation has taken place.

Gleysols

Gleysols (from Russian local name *gley*, mucky soil mass) or soils with gleyic properties are permanently wet and reduced in the subsoil (e.g., Schlichting 1973), and periodically to permanently wet in the topsoil (Spaargaren 1994). Mainly found in periodically poorly drained depressions, valleys, and low-lying coastal areas, the solum is either mottled (in case of temporary aeration) or has colors reflecting reduction. These redoximorphic features form under the influence of poorly to welldrained groundwater regimes. Typical gleyic soils are classified as Gleys or Groundwatergleys in many European systems of soil classification (e.g., Austria, Germany, France, Switzerland, United Kingdom), although in the German classification the periodically flooded soils of coastal regions are separated

as *Marschböden* from normal Gleysols. Gleysols are found in nearly all climates, from perhumid to arid conditions, and cover an area of almost 720 million ha (FAO 1991) (Table 1). In regions with permafrost (e.g., Arctic Canada, coastal plain of northern Siberia), Cryic Gleysols occur in association with Cryosols and Gelic Histosols. In the lowlands of the temperate latitudes, they are formed in alluvial sediments associated with Fluvisols near riverbeds in coastal areas, and with Histosols. In the humid tropics, they are found in valleys associated with Acrisols, Lixisols, Nitisols, Alisols, or Ferralsols.

Gleysols (Eutric Gleysols, FAO) are a dominant soil of Canada, with major areas occurring in the subArctic and Hudson Bay Lowland (Clayton et al. 1977). Most Gleysols occur on nearly level to undulating topography, particularly where associated with lacustrine, alluvial, or marine deposits that are moderately calcareous and loamy in texture. Cryic Gleysols occur throughout the entire Arctic region of northern Canada, including the Arctic islands extending from the Labrador coast to the Yukon-Alaska border. About 37,698 km² or 0.4% of Canada is mapped as dominantly Cryic Gleysols; these areas occur in the Northwest Territories and Yukon. They are found within the Mackenzie Delta and the Arctic Coastal Plain. Gleysols are often spatially associated with Podzols and Luvisols.

Anthrosols

These soils result from human activities that produce profound modification or burial of the original soil horizons. In the 1988 Revised Legend of the Soil Map of the World (FAO-UNESCO-ISRIC 1988), these are "... soils in which human activities have resulted in profound modification or burial of the original soil horizons, through removal or disturbance of surface horizons, cuts and fills, secular additions of organic materials, long-continued irrigation, etc. ..." Implicit in this general definition is the recognition of a distinct set of anthropedogenic processes that are associated with characteristic anthric diagnostic horizons. Anthropogenic soil materials that result from anthropogeomorphic processes are better considered as "non-soil," unless they are subjected to a sustained period of pedogenesis (Spaargaren 1994). Many different soil classification systems from around the world make provisions for these kinds of soils. Continuous cultivation of paddy soils often produces a hydric horizon that also includes redoximorphic and illuviation features in the subsoil. Many Anthrosols occur in coastal deltaic environments, especially in Southeast Asia and Indonesia. The Mississippi Delta region in the United States contains extensive areas of Anthrosols that are now subject to drowning by relative sea-level rise and contribute to the land-loss problem in coastal Louisiana (Boesch et al. 1994).

Leptosols-Entosols, Regosols, Lithosols, etc.

The name Leptosols (from Gr. *leptos*, thin) is used in the FAO *World Reference Base for Soil Resources* (Spaargaren 1994) for shallow soils overlying hard rock or highly calcareous material, and soils that have, by weight, less than 10% fine earth material. Leptosols occupy more than 1,655 million ha from the tropics to polar tundra (Table 1). Different names have been used in reference to these kinds of shallow, rocky soils. The term *Lithosol* was applied by Thorp and Smith (1949) to denote azonal soils having an incomplete solum or no clearly expressed soil morphology and consisting of freshly or incompletely weathered mass of hard rock or hard rock fragments. *Rankers* were defined as soils that have a dark-colored surface horizon, rich in organic matter with a low base status that is overlying unconsolidated siliceous material exclusive of alluvial deposits or hard rock. The name *Rendzina*, as first used by Sibirtzev (1901), was originally assigned to intrazonal soil on calcareous rocks. *Regosols* were first defined by Thorp and Smith (1949) as an azonal group of soils consisting of deep unconsolidated rock (soft mineral deposits) in which few or no clearly expressed soil characteristics developed.

Leptosols represent the initial phases of soil formation and as such, they are of great significance because they are the forerunners of the young or weakly developed soils of other units. These soils occur in landscapes where deposition or erosion, resistant bedrock, inert parent material, saturation, or human activity was limited normal pedological development. These soils commonly form on mountains, sand dunes, flood plains, and riverine lunettes, and coastal plains. In northern Canada, for example, Cryic Orthic Regosols in Arctic Lowlands commonly occupy gently undulating coastal plains developed on stony or sandy glacial till and marine deposits and occupy upwards of 1,305,423 km² (Clayton et al. 1977), but not all of this area is coastal. Parent material lithology is a factor that may inhibit pedogenesis where resistant bedrock, for example, increases runoff and erosion, and thereby decreases weathering. These soils also form in sandy parent materials that have an abundance of rock fragments. When parent materials consist of quartzitic sands without coarse fragments, such as in many sand dunes, profile development is also retarded.

Coastal dunes are built of beach sand blown onshore. Calcareous sand may give rise to calcareous eolianite, as on the coast of western Victoria (Australia), southwestern Western Australia, Bermuda, and the Bahamas. Quartz sand dunes are common in eastern Australia and grew particularly large during the last Ice Age, when sea levels were low and wide sandy flats were exposed (Ollier and Pain 1996; Short 2003). The high dunes of Queensland such as Fraser Island are the highest dunes in the world. Leptosols are also associated with beach ridges (Tanner 1995) that are formed as beaches with

broad curved shapes and lengths often of many tens of kilometers, as seen in this aerial photograph of a pocket beach in the Bahama Islands where the beach ridges mirror successive storm deposit build up (Fig. 2). Cheniers on the Yangtze Delta (China) are also tens of kilometers in length (Yan et al. 1989), up to 500 m wide, and similar to bight coast cheniers (Otvos and Price 1979) along the Louisiana-Texas coast. As wind blows sand from the beach to make ridges, these accumulations eventually become stabilized to support soil formation. Beach ridges often join together to make a chenier or beach ridge plain (cf. Fig. 2) where the ridges usually consist of coarser materials than the intervening swales. The more inland ridges are older than nearshore ones and composition of the ridges may be quite variable. A recent study by Rink and Forrest (2004) regarding the accretion history of beach ridges on Cape Canaveral and Merritt Island, FL, estimated ridges to be $4,000 \pm 150$ years old on the peninsula and about 43,000 years old on the more landward Merritt Island. Average ridge spacing was 109 ± 20 m, giving an average ridge accumulation rate of 80 ± 8 years per ridge on the cusped foreland. The ridge accumulation rate is related to storm frequency (tropical storms or hurricanes); soil formation proceeded on the ridges subsequent to deposition. On a coastal plain in southeastern Australia, for example, the inland beach ridges are quartzitic in composition and date back to the Miocene, whereas the nearshore sets are calcareous and formed throughout the Quaternary to the late Holocene (Cook et al. 1977). Many Tertiary marine beaches are now found high and dry in Western Australia, at places like Eneabba, Yoganup, and Minninup (Lissman and Oxenford 1975) that are now some distance inland from the present shore of the Indian Ocean. More recent beach ridge systems

are described for the inner Rottneest Shelf coast near Perth in southwestern Australia by Searle et al. (1988) where the main geomorphic components in the area are shore-parallel Pleistocene eolianite ridges and their associated intervening depressions, and the Holocene units in basins, slopes, banks-nearshore platforms, beaches, and beach-dune complexes.

Anaerobiosis may slow pedogenesis where there are persistent high water tables that prevent leaching of soil constituents, inhibit redox cycles, or preclude subsurface horizon development. Poorly drained floodplains and coastal marshes host these kinds of soils.

Cryosols

Cryo-hydromorphic soils occur in Arctic, subArctic, and boreal regions under cold continental, (sub)humid, or semi-arid climatic conditions. They support a characteristic vegetation cover of dominantly taiga, forest-tundra, or tundra with a ground cover of lichen moss. These soils have permanently frozen subsoil, often at a shallow depth of <1 m, and occupy enormous areas (1,800 million ha) typically on Arctic coastal plains (Table 1). Some show a significant accumulation of organ matter (dominantly weakly decomposed tissues of oligotrophic plants) at the surface, others are mainly mineral (or organo-mineral) in nature. They are saturated with water during the entire period of thawing, but lack features associated with reduction or redistribution of iron. The more mineral soils often show the peculiar phenomenon of thixotropy. Thixotropic properties are defined as the ability of relatively compact fine-textured soil materials to transform under

Coastal Soils, Fig. 2 Beach ridge plain near the southern tip of Eleuthera Island, Bahamas. The succession of calcareous beach ridges represents a developmental sequence of deposition, often associated with phases of storminess, where younger dunes and soils developed on them occur closer to the shore and older dunes and associated soils occur farther inland. Stages of soil development are usually referred to as chronosequences, where greater profile differentiation occurs on the older dunes. Minimal profile developed associated with weak phases of pedogenesis produce Leptosols (Lithosols), Arenosols, and Cambisols on the seawardmost ridges. Other types of calcimorphic soils occur inland in association with longer times for pedogenesis to precede.



pressure into an apparent liquid and unstable state when saturated with loosely bound water (quicksand-like property). When pressure weakens, the material gradually regains its initial compactness. Cryosols have been previously described in Russia as Peaty Frozen, Taiga Frozen, Gleyic Frozen, and Destructive Taiga Frozen soils and more recently as Homogeneous (or Peaty-Duff) and Thixotropic Cryozems. In Canada they are known as Organic and Turbic Cryosols. In *Soil Taxonomy* (Soil Survey Staff 1975, Soil Survey Staff 1992), they correspond with soils having a pergelic soil temperature regime.

Fluvisols

The central concept of this group relates to soil developed in fluvial (alluvial) sediments. Fluvisols were previously recognized by a variety of names including but not limited to swamp soils, Lithomorphie soils, Brown soils, Groundwater Gley soils, and Vaaggronden (“Vague soils,” Netherlands). Whatever the name, these soils are deposited in aqueous sedimentary environments viz. (1) inland fluvial and lacustrine environments, (2) marine environments, and (3) coastal saltings or brackish marsh environments, of which deltas are a special case (Dreissen and Dudal 1989). Fluvisols occupy significant areas on deltas and fans, coastal barriers, and tidal flats (350 million ha) (Table 1). On aggrading coastlines, combinations of sand or shingle spits and offshore bars allow dunes to accumulate, giving shelter to a nearshore zone where

silts and organic muds accumulate in low-energy or quiescent environments. Tidal inundation brings additional supplies of new suspended sediment that flocculates in the water column but which eventually settles to the bottom to help accretion of marsh soils. Aggradation of marsh surfaces takes place relatively slowly and is often associated with strictly zoned vegetation according to the length of time the marsh is inundated. Creeks give tidal waters access to the marsh at flood tide and drain the marsh at ebb tide. Deltaic systems develop where rivers debouching in the sea bring significant amounts of material, providing coastal conditions are satisfactory for sediment accumulation, as seen here in Fig. 3 along the Netherlands coast. In the intertropical regions, particularly, the conditions are such that reducing conditions may be developed in which the sulfates in the seawater are reduced to pyrites. Sediments containing pyrites may become parent materials for acid sulfate soils.

Fluvisols show clear sedimentary layering or stratification in a major part of the profile. Soil formation may have affected the upper part of the profile to produce dark-colored surface horizons that are high in organic matter and which form in comparatively little time (Fig. 4). Also, the material in the upper part of the profile below the surface horizon may have lost evidence of sedimentary layering due to weathering and have acquired the color and structure of stronger developed soils. These juvenile soils tend to show little evidence of weathering and soil formation below 25 cm depth, except possible gleying. Permanent or seasonal saturation with water, causing recurrent anaerobic conditions and low



Coastal Soils, Fig. 3 Organic-rich soil developed in fluvial-deltaic sediments flanking the IJsselmeer. These soils, reclaimed by diking and drainage, are susceptible to flooding and find use as pasture lands and reserves. Many soils developed in parent materials derived from marine

sediments become extremely acid when drained due to oxidation of sulfur compounds. The low pH values of acid sulfate soils preclude many uses.



Coastal Soils, Fig. 4 Core from a polder soil in The Netherlands. Depending on the nature of sediment grain-size distributions in the original marine sediments, soil texture can range from sand to clay. The flights making up this continuous core show minimal effects of pedogenesis due to the high groundwater table and reducing conditions. The upper part of the soil profile may show redoximorphic properties (e.g., mottles) due to fluctuating water tables and alternation of reducing and oxidizing conditions. The length of the hydroperiod in the upper part of the profile largely determines surface soil characteristics.

biological activity, tends to preserve the original stratified nature of the original deposits. Consequently, the more important linkages of Fluvisols are with other weakly developed soils viz. Cambisols, Regosols, Arenosols, Leptosols, Gleysols, and Solonchaks.

Solonchaks

Solonchaks (from Russian *sol*, salt, and *chak*) are widespread in arid and semiarid regions, occurring widely (covering 260–340 million ha) in many parts of Asia, Australia, South and North America, China, and the Middle East (Table 1). Salts responsible for salinity have many origins in parent materials that are related to marine deposition; volcanic, hydrothermal, and eolian processes; and lithologic properties of initial materials. Mangrove deforestation contributes to

salinization of coastal soils, erosion, land subsidence, and release of carbon dioxide to the atmosphere. The presence of these salts, the amount of osmotic pressure of the soil solution, or the toxicity of a given ion leads to special landscapes, either occupied by salt-tolerant (halophyte) vegetation or characterized by the complete absence of vegetation (salt lakes, salt lagoons, salt pans, etc.), depending on the degree of salinity. Two chemical criteria are used to define these soils: (1) the solubility product of the accumulated salts or of salts that may form and (2) the ion concentration in the soil solution. To be considered salt-affected, the soils must contain an important quantity of salts that are more soluble than gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$: $\log K_s$ is -4.85 at 25°C), thus also more soluble than calcium carbonate, jarosite, and iron sulfide. For this reason, gypsiferous soils, calcareous soils, and soils with jarosite are excluded from the Solonchaks and have a home in the Gypsisols, Calcisols, and Thionic Fluvisols, respectively (Spaargaren 1994). These soils must also have a total salt concentration of the soil extract, expressed as the electrical conductivity (EC) within a depth of 125 cm of the surface. This minimum value, measured in a saturated paste extract (or in a 1:1 extract for very sandy soils) must be: (1) $>15 \text{ dS m}^{-1}$ (or mS cm^{-1}) at 25°C , if the pH is less than 8.5 (neutral salts) or (2) more than 8 dS m^{-1} if the pH is more than 8.5 (for alkaline carbonate soils) or less than 3.5 (for acid sulfate soils other than Thionic Fluvisols).

Chloride soils are of marine origin but influenced by the potential acidity of the soils under mangrove. The pH is very acid to acid (3.5–5) in reaction, calcium is almost lacking and sulfates, entirely derived from seawater, are present in only very small quantities compared to chloride. Sodium is dominant in the soil solution ($\text{CL} \gg \text{SO}_4 > \text{HCO}_3$; $\text{Na} \gg \text{Ca}$). White salt deposits (NaCl , CaCl_2 , MgCl_2) may occur at the soil surface as crusts or efflorescence, which seasonally can turn into brown powdery surface layers. Examples are noteworthy in Gambia and in the Casamance region of Senegal.

Neutral chloride-sulfate soils are mostly of marine origin, but can be enriched by sedimentary gypsum deposited by rivers. When the pH is near neutral, sulfates and chlorides occur, and carbonates, sodium, calcium, and magnesium are in variable proportions and exceed bicarbonates present in variable quantities. Salt deposits are mostly NaCl , CaCl_2 , and MgCl_2 associated with gypsum in the profiles. These soils are widespread and examples can be found in the alluvial plains, chotts, and sebkhas of North Africa, the Mexican Lagunas or playas, along the Euphrates River in Syria and Iraq, and in salt flats elsewhere.

Acid sulfate soils of marine origin are formed in potentially acid environments of mangrove vegetation (mangals) where acidification is promoted by dry periods in some tropical regions. When the environment becomes extremely acid ($\text{pH} < 3.5$), the result is acidolysis of clay and liberation of aluminum, iron, and magnesium; there is little calcium and chloride in these systems. The lack of bases and very low pH

prevents the formation of jarosite. Examples occur in Gambia, the Casamance region of Senegal, and in Guinea Bissau. In reduced condition, these soils classify as Thionic Fluvisols as they have a sulfidic horizon; in oxidized and acidified states, these soils may classify as Salic Fluvisols if the EC is sufficiently high.

Carbonate soils include alkaline bicarbonate-sulfate soils and strongly alkaline carbonate soils. They form in aerobic environments with the production of sulfates from sulfides present in the original marine muds. The liberated acidity is neutralized by an excess of sodium and calcium carbonates in the biogeochemical system. Variations of these soils are found in the polders of Tchad, New Caledonia, the Balkan region, and Australia. The Solonchaks occupy the lower parts of landscapes where runoff, seepage, and shallow groundwater can accumulate; they are separated from most other soils by having a high salt content at or near the ground surface.

Podzols

The term *Podzol*, meaning “white earth” is derived from the Russian terms *pod* (beneath) and *zol* (ash). These soils are widely known as acid ashy-gray sands over dark sandy loams (Buol et al. 1980). These two contrasting albic and spodic horizons, with an abrupt boundary between them, make these soils among the most dramatic, eye-catching, and photogenic soils in the world (Fig. 5). The setting in which such a soil is produced requires the combination of soil-formation factors that yield necessary conditions where there is accumulation of iron, aluminum, and/or organic matter in a subsoil horizon. Podzols thus occupy large areas where there are sandy parent materials from the tropics to the boreal climatic zone (485 million ha, Table 1). Many kinds of vegetation, including, grasses, can yield certain organic compounds that enhance podzolization, but litter from a certain few species of plants in particular foster such accumulations viz. the hemlock tree (*Tsuga canadensis*) in forests of northern latitudes, the kauri tree (*Agathis australis*) of New Zealand, and heath (*Calluna vulgaris* and *Erica* sp.) in northern Europe. Podzols additionally occur under a wide variety of trees (e.g., *Picea*, *Pinus*, *Larix*, *Thuya*, *Populus*, *Quercus*, *Betula*) and understory plants in moist climatic zones that are cool (e.g., Humid Meso- and Microthermal climates). Humic Podzols (Cryohumods in *Soil Taxonomy*) occur as dominant soils in Canada, for example, in the Maritime Provinces (e.g., Cape Breton and Newfoundland), in the coastal forests of the British Columbia mainland, and on coastal islands (Clayton et al. 1977). Extensive areas of these soils occur on the humid subtropical coastal plain of the southeastern United States. Impressive are the 9 m thick Bh horizons in Spodosols on the North Carolina Coastal Plain (Buol et al. 1980). Ages of these soils are extremely variable, some Spodosols being able to



Coastal Soils, Fig. 5 Spodosol (Podzol) developed in siliceous, Pleistoceneage Pamlico Sands overlying a portion of the limestone Atlantic Coastal Ridge along the subtropical southeast coast of Florida. Here, prominent albic tongues (eluvial zone) extend at least 1.5 m into the organic- and iron-rich spodic horizon (illuvial zone). These coastal Spodosols are extremely photogenic due to the extreme biochemical differentiation within the profile, dramatic contrasts in soil color, abrupt wavy horizon boundaries, and thick profiles.

form relatively quickly. Burges and Drover (1953), for example, reported evidence that 200 years were required to leach calcite from beach sand in New South Wales, Australia, but 2,000 years to produce an iron-Podzol, and 3,000 years to produce an iron-humus Podzol with pH as low as 4.5. An iron-Podzol in coastal California has been estimated at about 1 million years (Jenny et al. 1969). On coastal dunes in cooler climates, Podzols may develop within a few centuries, but in warmer or less humid climates Podzols require several millennia or longer periods to form (Sevink 1991). Dune soils are easily leached because of their high permeability and low water-storage capacity. Although sometimes high in carbonates, the acid buffering capacity of the non-calcareous fraction (e.g., quartz) is generally very low. As a consequence,

acidification proceeds rapidly with concurrent changes in soil-forming processes and profile characteristics. Rates of leaching and acidification are high in climates with a large precipitation surplus and low decomposition rates (because of low temperatures). In arid climates, leaching will be strongly reduced and mobile components such as carbonates and gypsum may accumulate within the solum. Accumulations of secondary carbonates (e.g., caliche, kunkar) are frequently cited in studies of Holocene drier systems in drier coastal zones.

Podzolization is thus a bundle of processes that bring about translocation, under the influence of the hydrogen ion, organic matter, iron, and aluminum (and a small amount of phosphorus) from the upper part of the mineral solum to the lower part (Buol et al. 1980). Cheluviation, the dominant process of soil formation of Podzols, involves the combination of two processes: (1) weathering of primary minerals (especially phyllosilicates) by organic complexing acids and (2) translocation of organic matter both in the weathering of primary minerals and in the translocation of Al and Fe complexes. These processes lead to special morphological and analytical characteristics that are used to identify a soil material as Podzolic. Cheluviation affects large areas of soils in the boreal zone. The process is also active in humid regions, especially in the temperate zone but also in the equatorial realm where many examples of Podzols have been described.

Podzols are soils characterized by the presence of an illuvial spodic horizon. In this horizon, amorphous compounds have accumulated consisting of organic matter and aluminum, with or without iron or other cations. The process of translocation and accumulation, known as cheluviation, is usually shown by the occurrence of an albic (bleached) horizon underlain by a spodic horizon. The illuviation of organic compounds can often be demonstrated by the presence of thick cracked organic coatings on the skeleton grains within the spodic horizon. In coarse sandy materials, Podzol morphology is well expressed and strong contrasts occur between eluvial (albic) and illuvial (spodic) horizons (cf. Fig. 5). In most Podzols, where clay percentage is often less than 10%, coarse texture ranges from sand to sandy loam. Accordingly, water retention capacity is low, less than 50 mm and, although Podzols develop in humid climates, they often show moisture stress. Podzols are very acid soils, with a pH ranging from 3.5 to 4.5 in surface horizons; the pH may increase up to 5.5 in lower horizons. The cation exchange capacity (CEC) is mostly due to the organic compounds present and base saturation is always very low. Organic matter has a high C/N ratio especially in the surface horizons ($C/N = 25$ or more) and in the spodic ($C/N = 20$ or more), indicative of low biological activity and slow degradation of organic matter.

The limits of the Podzols are determined by the minimum expression of the spodic horizon and the minimal contrast between the eluvial and illuvial horizons. Soils showing

evidence of illuvial organo Al/Fe complexes but lacking sufficient amounts of it to qualify for Podzols, form intergrades with Cambisols, Arenosols, and Gleysols. Podzols and intergrades are typical of a soil mantle on sandy coastal plains with poor quartzitic materials affected by a shallow water table.

Sesquisols, Ferralsols, Alisols, Acrisols, Lixisols

Sesquisols are soils either containing at shallow depth a layer indurated by iron (petroplinthite), or at some depth mottled material that irreversibly hardens after repeated drying and wetting (plinthite). *Plinthite* is an Fe-rich, humus-poor mixture of clay with quartz and other diluents that occurs as redoximorphic concentrations in platy, polygonal, or reticulate patterns (Shaw et al. 1997). These kinds of soils occur mainly in the tropics but examples are common to old landscapes in (sub)tropical and temperate regions. Those with a shallow petroplinthite horizon are known as (high level) laterites, ironstone soils, or Sols ferrugineux tropicaux a cuirasse. They have widespread occurrence in western Africa, especially in the Sudano-Sahelian region where they cap structural tablelands; they are also common in central-southern India, the upper Mekong catchment, northern Australia, and the eastern part of the Amazon region (Table 1).

The Sesquisols with plinthite are known as Plinthosols, groundwater laterite, sols gris lateritiques, or Plinthaquox (Soil Survey Staff 1992). They are found in extensively flat terrains with poor internal drainage, such as the late Pleistocene or early Holocene sedimentary basins of eastern and central Amazonia and the central Congo basin, but also on coastal plains such as in southwestern Western Australia (e.g., Swan Coastal Plain, Ridge Hill Shelf, Donnybrook Sunkland). Well-drained soils with loose ironstone concretions (sesquiskeletal material) are frequent nearly everywhere in the tropics and subtropics, in many landscape situations. The material is the result of former plinthite formation, subsequent hardening, and transport or re-weathering.

The central concept of Sesquisols is one of a soil affected, at present or in the past, by groundwater or stagnating surface water in which iron has been segregated to such an extent that a mottled (redoximorphic) layer has been formed which irreversibly hardens when exposed to air and sunshine. Included in the concept are those soils that have a hardened layer at shallow depth. The most important characteristic of Sesquisols therefore is the presence of plinthite and petroplinthite. Plinthite is an iron-rich, humus-poor mixture of kaolinitic clay with quartz and other diluents. It commonly occurs as red mottles in platy, polygonal, or reticulate patterns, and changes irreversibly on exposure to a hardpan or to irregular aggregates on exposure and repeated wetting and drying with free access of oxygen. In a perennial wet soil,

plinthite is usually firm but can be cut with a spade. A plinthic horizon is at least 15 cm thick in which plinthite takes up 10% or more by volume. Petroplinthite is a continuous layer of indurated material in which iron oxides are an important cement and in which organic matter is absent, or present only in traces. The Fe_2O_3 content is generally greater than 30%. The continuous ferruginous duricrust layer may be either massive, reticulate, interconnected platy, or columnar pattern that encloses non-indurated material. The indurated layer may be fractured, but then the average lateral distances between fractures is 10 cm or more and the fractures themselves do not occupy more than 20% by volume within the layer.

Sesquisols dominantly occur in intertropical regions and have linkages with Ferralsols, Alisols, Acrisols, and Lixisols. Sesquisols may occur in distinctly different landscape positions. Petric Sesquisols occupy the higher landscape positions, often as a result of landscape inversion due to lowering of the erosion base (McFarlane 1976). They now form tablelands and are usually freely drained. The other Sesquisols are found mainly in depressions or other areas with impeded drainage. On the Coastal Plain of Georgia (USA), for example, most soils have sandy surface horizons with high infiltration rates; many, however, have slowly permeable subsoil horizons with high concentrations of plinthite (Shaw et al. 1997) where only 10% platy plinthite is necessary to perch water (Daniels et al. 1978). These soils have developed from Miocene aged sediments and are classified in *Soil Taxonomy* in fine-loamy, siliceous, thermic families of Plinthaquic, Aquic, Arenic Plinthic, Plinthic, and Typic Kandiodults. In Georgia, more than 1.5 million ha of soils containing plinthite have been mapped. Sesquisols are found in association with Gleysols and Stagnosols in areas conditioned by hydromorphy, and with Ferralsols, Alisols, Acrisols, and Lixisols, which occupy the better-drained positions in the landscape.

Tropical weathering and pedogenesis produces Ferralsols that are characterized by a colloidal fraction dominated by low-activity clays (1:1 alumino-silicate minerals) and sesquioxides. These highly weathered soils lack rock fragments with weatherable minerals and are characterized by the presence of low-activity clays and a uniform morphology that lacks distinct horizonation. If there is sufficient iron in the original material, these soils are reddish in color and contain a weak pedal structure; there are few marks of other soil-forming processes such as clay accumulation through translocation. Secondary accumulation of stable minerals such as gibbsite or iron oxyhydrates may be present in concretionary forms (e.g., pisolites) as part of the fine earth fraction. Typical soils are situated on old geomorphic surfaces, which formed through erosion and deposition, and occur today as relict landscapes over large areas (750 million ha, Table 1). Having formed in transported and reworked colluvial materials, these

strongly weathered soils have little relationship with underlying geological strata. As a result of transport and deposition, the soils may be stratified (e.g., McFarlane 1976; Finkl 1980, 1984). In extreme stages of weathering in a free leaching environment, there is relative accumulation of sesquioxides and small amounts of 1:1 alumino-silicate minerals (e.g., kaolinite) as well as resistant minerals such as anatase and rutile. The sesquioxide minerals include goethite, hematite, and gibbsite; the alumino-silicate minerals are generally degraded kaolinites and some muscovite. In this weathering environment, neogenesis of alumino-silicates is rare with the possible exception of aluminum chlorite.

Occurring on geomorphically old surfaces, these soils are associated with Cambisols in areas with rock outcrops or where rock comes near to the surface. On the stable surface they occur together with Acrisols, which seem often to be related to the presence of more acidic parent materials (e.g., gneiss). On more basic rocks they occur associated with Nitisols, with no apparent relation to underlying rocks or topographic positions. Near valleys, Ferralsols merge into Gleysols.

Alisols (from *L. alumen*, alum), occurring in humid (sub) tropical and warm temperate regions, are acid soils with a dense layer of illuviated clay in the subsoil; they cover more than 100 million ha worldwide (Table 1). The intense weathering processes in these soils degrade 2:2 and 2:1 clay minerals and release large amounts of aluminum to create a very acid environment. Alisols correlate partly with Aquults, Humults, and Udults in *Soil Taxonomy* (Soil Survey Staff 1975, 1992), the Ferralsols or sols ferralsitiques tres lessive in French systems (e.g., AFES 1992; CPCS 1967), and Red Yellow Podzolic soils with a high clay activity in the Brazilian soil classification.

Acrisols (from *L. acris*, very acid) are characterized by a subsurface accumulation of active clays, a distinct clay increase with depth, and a base saturation (by 1 M NH_4OAc) of less than 50%. The soils have been named Red-Yellow Podzolic soils, Podzólicos vermelho-amarelo distróficos a argila de atividade baixa, sols ferrallitiques fortement ou moyennement désaturés (CPCS 1967), Red-Yellow Earths, Latosols, and oxic subgroups of Alfisols and Ultisols. The later have been redefined as kand- and kanhapl- great groups in *Soil Taxonomy* (Soil Survey Staff 1975, 1992). These soils are common in tropical, subtropical, and warm climatic regions, on Pleistocene and older surfaces that take up more than 900 million ha worldwide (Table 1).

The dominant characteristic of Luvisols (from *L. luere*, to wash) is textural differentiation in the profile that forms a surface horizon depleted in clay and an accumulation of clay in the subsurface argic horizon. These soils are further characterized by moderate to high activity clays and low aluminum saturation. These soils are known as sols lessivés in France, Parabraunerde in Germany, pseudo-Podzolic soils

in Russia, Gray-Brown Podzolic soils in earlier US terminology or as Alfisols in *Soil Taxonomy*.

Lixisols (from L. *lix*, lye) are characterized by a subsurface accumulation of low-activity clays (cation exchange capacity is less than 24 cmol(+)/kg) and moderate to high base saturation. They show a distinct increase in clay content with depth. These soils have been named Red-Yellow Podzolic soils, Podzólicos vermelho-amarelo eutróficos a argila de atividade baixa, sols ferrugineux tropicaux lessives, and sols ferrallitiques faiblement désaturés appauvris (CPCS 1967), Red and Yellow Earths, Latosols, and oxic subgroups in *Soil Taxonomy* (Soil Survey Staff 1992). Limits and linkages of Lixisols, Acrisols, Alisols, and Luvisols are entirely based on analytical properties. Therefore, in areas where these kinds of soils have CECs close to this value, they will merge into each other.

These soils, which are relict (out of pedogenic phase with the present environment) and not related to contemporary soil-forming processes in coastal zones, are nevertheless important pedogenic and geomorphic features along many coasts; they occur about 435 million ha worldwide (Table 1). Although commonly associated with truncated landsurfaces that produce coastal cliffs such as those seen on marine sediments along the eastern boundary of the Guiana Shield in northeastern South America (Valeton 1981); the coastal Trivandrum district in southern Kerala (India) (Karunakaran and Sinha Roy 1981); Long Reef Beach near Sydney, New South Wales (Retallack 1990); Darwin, Northern Territory (Australia) (McNally et al. 2000); and as lateritic spurs and benches along the Darling Fault Scarp in southwestern Australia (e.g., McArthur and Bettenay 1960), these weathering profiles also occur as reefs on seaward sloping surfaces that extend below present MSL (Short 2003) in the Northern Territory and into the Gulf of Carpentaria (Hays 1967). Interestingly, these Tertiary and older weathering profiles extend from northern Australia to the southern boundary of Papua New Guinea but are now drowned by the eastern Arafura Sea and Torres Strait. Depending on the thickness of the weathering profile and its intersection with present MSL, a variety of materials may occur along the shore with beaches or marine terraces developed in saprolite, pallid zones, mottled materials, or ferruginous duricrust.

In the case of the Darwin porcellanite (Darwin Member of the Bathurst Island Formation), a siliceous duricrust that was produced by leaching and replacement of preexisting rocks, the degree of alteration is less than that found in silcrete (McNally et al. 2000). Pisolithic ferricrete (lateritic or iron-stone “pea gravel”) occurs in close association with the porcellanitic duricrust throughout the Darwin area. A karstlike relationship between the two types of duricrust (porcellanite versus ferricrete) is exposed along road cuts in Darwin, as described by McFarlane et al. (1995) where the ferricrete fills cavities in the porcellanite. This ferricrete is pedogenetically unrelated to the old lateritic pallid zone that

developed in the Cretaceous sulfide-bearing marine muds following the retreat of the early Cretaceous sea. Porcellanization around Darwin is mostly confined to lower Cretaceous rocks that weathered to form the pallid zone of a laterite profile 50–70 million years ago. This deep weathering profile was later silicified in a number of cycles that were most intense during the Miocene (Milnes and Thierry 1992). Both silicification and ferruginization continue to the present day, as evidenced by cliff-base 1940s junk that has been cemented together in a kind of “bottlecrete” by iron-charged spring waters. Silicified and ferruginous “beachrock” occurs at Fanny Bay (Darwin) between the present high and low tide lines. McNally et al. (2000) report that these beachrocks occur in a groundwater discharge zone, indicating that some duricrusts have developed since the postglacial sea-level stabilization around 6,000 BP. Marine erosion along the Darwin coast cut marine benches and eroded cliffs into deep mantles of chemical weathering that retain lateritic and siliceous Paleosols. Slumps of pallid zone materials at the cliff base provide residual lithogenic quartz and secondary (epidiagenetic) agglomerated porcellanized particles as beach materials. Parts of the beachface are silicified and ferruginized to form an unusual kind of “beachrock.” Younger coastal soils have developed in the older deep weathering profiles that are exposed along the shore. Complicated polygenetic laterite profiles are also reported from Hong Kong (e.g., Wang 2003) where pedogenic and ground-water origins represent paleosurfaces that require prolonged subaerial exposure, quiescent tectonism, and a certain level of climatic stability.

Planosols

Planosols (from L. *planus*, flat) have a bleached, light-colored eluvial horizon that abruptly passes into dense subsoil with significantly higher clay content. These soils typically occur in seasonally or periodically wet, level areas, often above normal flood levels of nearby rivers or estuaries. Planosols occur on nearly level river or marine terraces in Southeast Asia from Bangladesh to Vietnam, southern Brazil, southern and eastern Africa, and southern Australia. On a global scale, Planosols occupy about 130 million ha (Table 1). Melanization, the addition of dark-colored organic matter, is the process by which a dark surface horizon extends down into the profile. This bundle of specific processes includes: (1) extension of roots from grasses in the profile, (2) partial decay of organic materials in the solum, (3) reworking of the soil and organic matter (bioturbation) by earthworms, ants, rodents, with the development of krotovinas (follod burrows), (4) eluviation and illuviation of organic colloids along with some mineral colloids, and (5) formation of resistant “ligno-protein” residues that give a dark color to these soils (Buol et al. 1980).

Cambisols

Cambisols (from *L. cambiare*, to change) are moderately developed soils that are characterized by slight to moderate weathering of parent materials. They do not contain appreciable amounts of illuviated clay, organic matter, aluminum and/or iron compounds. These soils are particularly common in temperate and boreal regions that were influenced by ice during recent glacial periods. The poorly developed profiles result because soil parent materials are young and because soil formation is slow under low temperatures (or even permafrost) of high latitudes in northern Russia and Canada. Cambisols also develop in pre-weathered, old parent materials where they occur in association with highly developed soils such as Acrisols and Ferralsols. The largest continuous surface of Cambisols in warm regions is found on young alluvial plains, terraces, and deltas of the Ganges-Brahmaputra river system. Cambisols are also quite common in arid climates, where they are closely associated with Calcisols. Cambisols cover an enormous area of about 1.5 billion ha worldwide (Table 1) and constitute the third largest major soil grouping in the FAO Revised Legend (FAO 1991). The minimum degree of soil development present in Cambisols separates them from Regosols. Leptosols are not permitted to have a cambic horizon and the shallowness of Leptosols separates them from Cambisols. When the cambic horizon reaches its maximum expression, Cambisols border soils defined by an argic horizon (e.g., Luvisols, Alisols,

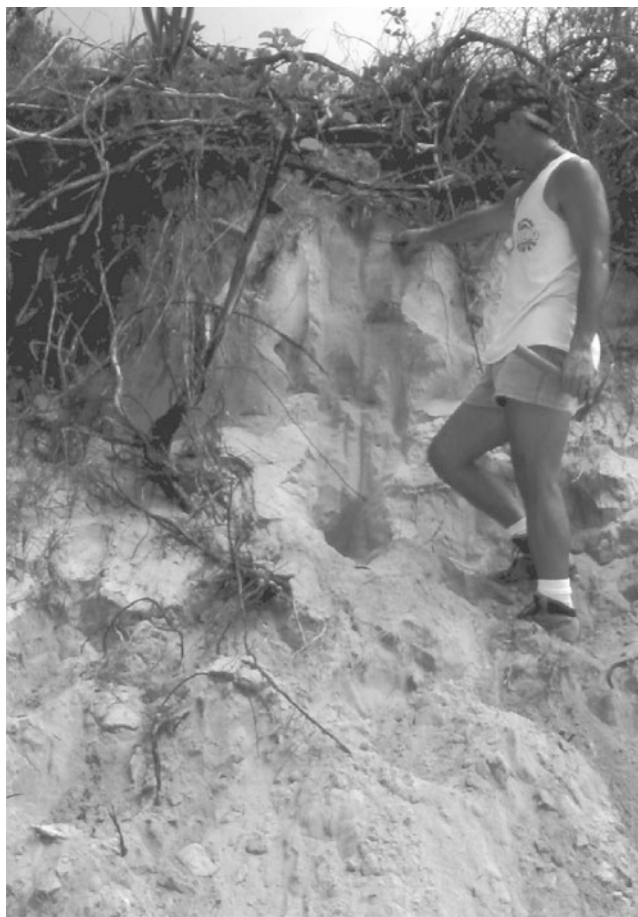
Acrisols, Lixisols, Nitisols), a spodic horizon (Podzols), or ferralic horizon (Ferralsols). Definition of these horizons is such that the cambic characteristics are excluded.

Arenosols

Arenosols (from *L. arena*, sand) are sandy soils with slight to moderate profile development. These soils are widely distributed and form one of the more extensive major soil groups in the world. FAO (1991) estimates Arenosols to cover about 900 million ha (Table 1), or 7% of the land surfaces; if shifting sand and active dunes are included, the coverage is about 10%. Sandy coastal plains and coastal dune areas are, of course, of smaller extent but ecologically very important. Although Arenosols occur predominantly in arid and semi-arid regions, they are typical azonal soils that occur in the widest possible range of climates from very arid to very humid and from hot to cold. Arenosols occur predominantly on eolian sands, either dunes (Fig. 6) or flats, but have also formed on marine, littoral, and lacustrine sands of beach ridges, lagoons, deltas, lakes, etc. Other names for Arenosols include Psamments (Soil Survey Staff 1975, 1992), sols minéraux bruts (France), siliceous, earthy and calcareous sands, and various Podzolic soils (Australia), red and yellow sands (Brazil), and raw sands (Britain and Germany), among others. Arenosols thus have linkages with most other major soil groups. Only a few groups, such as Vertisols, have no

Coastal Soils, Fig. 6 Primary coastal dune on the Atlantic coast of southern Eleuthera Island, Bahamas. These Holocene carbonate dune sands are stabilized by vegetation and anchored to a rocky headland (photo center). Beachrock outcropping on the beachface helps to protect the dune foot from erosion.





Coastal Soils, Fig. 7 Exposure of fine-grained calcareous dune sands on the landward side of dunes. The imprint of pedogenesis is minimal and basically associated with accumulation of organic matter on the dune surface. Translocation of organic compounds down into the profile is limited to the upper half-meter or so.

linkages. The limits are determined by either textural requirements or the occurrence of certain diagnostic horizons within specified depths. Of most importance in their development are factors that limit soil horizon development in wetlands, alluvial lands, sand lands (Fig. 7), rocky areas, and various unconsolidated materials. The spatial distribution of Arenosols indicates that many factors are involved, which operate in various combinations to limit profile development: (1) dry but hot to warm to cold climates and microclimates that parse the amount and duration of water movement in the soils (Arctic, subArctic, Antarctic, and temperate and inter-tropical desert zones provide these conditions), (2) mass wasting and solifluction remove surficial material faster than most pedogenic horizons can form, (3) cumulization adds new material to the surface of the soil faster than new material can be assimilated into a pedogenic horizon, (4) immobilization of the soil plasma in inert materials, in carbonate-rich flocculated materials, and in some highly siliceous sediments to inhibit profile differentiation by illuviation, (5) exceptional

resistance to weathering (pedologic inertia) of some initial materials, (6) infertility and toxicity of some initial materials to plant growth that inhibits biogenetic differentiation (e.g., on serpentine barrens), (7) saturation with water or even submergence for long periods, (8) short exposure of initial material to active factors of soil formation (e.g., marine flats newly exposed by uplift), and (9) a recent drastic change in the biotic factor that initiates formation of a new pedogenic regime in an old one, which in turn serves as initial material (Buol et al. 1980).

Regosols

Regosols (from Gr. *rhegos*, blanket) are well-drained, medium-textured, deep mineral soils that are derived from unconsolidated materials and which have been traditionally separated from shallow soils (Lithosols, Leptosols, etc.) and from those of sandy or coarser texture (e.g., Arenosols). A Regosol does not have assemblages of diagnostic horizons, properties, or materials that are definitive for any other group of soils. Most properties of Regosols are associated with the materials themselves and not with genetically developed soil features. Conceptually, Regosols are the initial state for pedogenesis representing recently deposited or recently exposed, earthy materials on a ground surface; these weakly developed soils contain poorly defined horizons. Some soil materials, which represent a late stage of weathering with few distinguishing characteristics, are recognized as Regosols. For example, thin ferrallitic materials that cannot go into Cambisols and may be classified with Regosols (Spaargaren 1994). Although there are examples of initial stages of soil development in all landscapes throughout the world, the areal extent of Regosols is often limited and these soils are often inclusions in other soil units. Although of generally limited extent on a global scale (Table 1), Regosols can be important soils in specialized environments (e.g., tundra), such as Arctic plains in northern Canada where Cryic Regosols account for about 23.5% of Canada's land area (Clayton et al. 1977), extending from Labrador and Baffin Island on the east to the Mackenzie Delta in the west. Nearly 3,000 km² of Cumulic Regosols (Dystric Fluvisols in *Soil Taxonomy*) occur in tidal marshes adjacent to the Bay of Fundy in the Maritime Provinces as well as on coastal floodplains adjacent to the Yukon and Mackenzie deltaic river systems. In the terrain, Regosols are mostly associated with degrading or eroding areas, while other soils occur on aggrading, depositional, or stable areas. As time passes and soil formation proceeds, Regosols may develop into many other soils depending on the most important soil forming factor(s). Sand dunes on the central Delmarva Peninsula of Maryland and Delaware, for example, contain weathering zones that are represented by Regosols (Entisols in *Soil Taxonomy*) with profiles that are only a meter

or two in thickness (Denny and Owens 1979). Basal horizons in the dunes constitute an ancient weathering profile, truncated at the top and overlain by a younger deposit of sand. These so-called two-cycle dunes thus contain an older basal dune that is Wisconsinan in age and the early episode of weathering may have been early Holocene, a comparatively warm Hypsithermal interval. The surface sand may be no more than a few hundred years old, the result of deforestation and cultivation in colonial times (Denny and Owens 1979). The Sharon Coastal Plain in Israel is predominantly comprised by sand dune fields and eolianite ridges. Cliffs along this plain contain three major sand accumulation periods between 65 and 50 ka, from 7 to 5 ka, and from 5 to 0.2 ka (Frechen et al. 2002), that in turn seem to correlate with periods of sapropel formation in the Mediterranean Sea and so with periods of strongly increased African monsoon activity. The uppermost eolianite (calcareous sandstone) seems to correlate with a sapropel formed at about 7.8 ± 4.0 ka BP. Major phases of soil formation occurred between 35 and 25 ka resulting in a Rhodoxeralf soil (Grumusolic Dark Brown soil) and between 15 and 12 ka forming a Rhodoxeralf (hamra) soil. Frechen et al. (2002) identify at least seven weak pedogenic phases resulting in weakly developed horizons of Regosol-type soils that are intercalated in the eolianite sequences. The oldest eolianite layer contains kurkur (caliche) and seems to be about 55 ka bp in age.

Coastal Plain Soils

As a general rule, soils of younger landscapes are less differentiated (i.e., they contain simple sequences of horizonation and plasmic soil materials are unorganized) than those of older surfaces where weathering profiles tend to contain highly organized soil materials with complex horizonation (e.g., Butler 1959; Huggett 1975; Buol et al. 1980; Gerrard 1981; Ollier and Pain 1996). Because considerable time is needed to form differentiated soils where soil horizons and materials are highly organized, it is possible to sequentially arrange soils in chronological order with soils developed in young deposits placed before those formed later. Soils arranged in this way represent a sequence of development, referred to as a *chronosequence* (Jenny 1941). If prior conditions of weathering continue, the young soils in time develop characteristics seen in the oldest (Gerrard 1981; Ward and McArthur 1983). This is a simplistic interpretation of uniform and uninterrupted development, but other more complicated (interrupted) sequences are possible, such as apparent trends due to secular changes in climate, drainage, or biota. Investigation of elemental mass balance trends, based on element translocation patterns in response to climate forcing, show promise for estimating the weathering flux under drier or wetter mean annual precipitation regimes (e.g., Stiles et al.

2003). Optical dating of young coastal dunes (e.g., Ballarini et al. 2003) provides a tool for reconstructing coastal evolution on a timescale of decades to a few hundred years. Quartz optically stimulated luminescence (OSL) dating is based on the fact that OSL dates of less than 10 years are very well zeroed prior to deposition and burial, as demonstrated by Ballarini et al. (2003) for an accretionary coastal segment near Texel, The Netherlands. Other luminescence sediment dating methods, more amenable to timescales of a very few thousand years, have found application in the study of the activation and stabilization of coastal dunes on a North Carolina barrier island (Berger et al. 2003). Here, the quartz SAR (single-aliquot regenerative-dose method) is preferred for photonic dating in this region because of the relative abundance of quartz. These kinds of dating methods provide tight controls for estimates of pedogenic phases and the estimation of time intervals required to produce specific horizonation sequences in coastal dunes.

Radiocarbon dating is limited by the availability of datable materials and the technique can be applied only to younger soils (<50,000 year BP) that range from latest Pleistocene to Holocene in age (e.g., Vogel 1980; Bowman 1990). The prospect of dating and correlating coastal and bottom deposits as well as terrestrial-marine weathering sequences, as seen in coastal soils, is best achieved along the littoral where sedimentation and subsequent soil development are closely associated with changes in sea level (e.g., Karpytchev 1993). Lake-level histories as well can sometimes be gleaned from radiocarbon dates for coastal dune fields, as demonstrated by Arbogast and Loope (1999) for Holocene coastal dunes on the shores of Lake Michigan. When marine deposits are correlated with terrestrial weathering sequences by stratigraphic linkage at the coast, pedogenesis can be related to sea-level change and erosion-deposition cycles (Kukla 1977). If dates related to sea-level changes can be assigned to coastal landforms, then the chronology can be extended inland by using local markers (Ward and McArthur 1983). At the coast, abundant shells can supplement wood and peat as a basis for radiocarbon dating (Mook and van de Plassche 1986).

If there is uncrystallized coral present, then dates back to 200,000 years can be obtained from uranium decay methods, and on volcanic shores potassium/argon dating provides useful information. For coasts where these methods are not applicable, the principal means of dating refers to the correlation of stranded shorelines. If the former shores in one locality cannot be dated directly, it is sometimes possible to relate them to the shores in other localities where dates exist, assuming that changes in sea level are instantaneous and global (Ward and McArthur 1983).

Because the earth's crust is in dynamic equilibrium with the asthenosphere and tectonic adjustments are always taking place, no coastal landmass can be regarded as completely stable, no matter how free it is from seismic activity or

neotectonism. Few coasts are so mobile, however, that they contain no part of the eustatic record. Landforms in the coastal zone are complex and not all are only the product of terrestrial processes or marine action.

The apparent simplicity of most coastal plains is suggested by their lack of relief, which disguises these plains from their true nature as terminal geomorphic surfaces of complex transgressive-regressive depositional units laid down during glacioeustatic fluctuations. Bounded at the surface by terrestrial and marine erosional, transportational, and depositional landforms, the stratigraphic units (including paleosols) are limited below by unconformities of terrestrial and marine origin. Sedimentary facies between these surfaces reflect coastal plain (alluvial, palustrine, and eolian), estuarine, and marine open-shelf environments (e.g., Curray 1978; Davis 1978). Their evaluation requires a complete stratigraphic examination, as offered by Haq (1995) in summaries of the principles and concepts of sequence stratigraphy (the study of rock relationships within a chronostratigraphic framework wherein the succession of rocks is cyclic and is composed of genetically related stratal units viz. sequences and systems tracts).

Ideally, two main sedimentary sequences occur within a transgressive-regressive unit (e.g., Blum et al. 2001). The terrestrial sequence, which develops as sea level regresses seaward through lowering, results from extension of alluvial erosion and sedimentation, and of soil formation, into the coastal zone. The marine sequence forms during subsequent landward transgressions or rise of sea level when marsh, estuarine, beach, dune, and deltaic sediments are introduced. The relative thickness, significance, and spatial extent of these sequences depend on the dynamics of coastal processes and on sediment volumes brought to the shore by rivers. When major rivers deposit large volumes of alluvium at the shore, the marine contribution to the sedimentary pile is limited. Coastal landforms thus result predominantly from alluvial distribution.

Coastal strand plains with no alluvial contribution, such as those forming in the Holocene landscape of Gippsland, Victoria, in Australia (Ward 1977), lie at the other extreme of the scale. Relative placement of terrestrial and marine sediments here is determined by gradients at the coast, and by the history of eustatic fluctuations relative to the land. If, for example, sea level rises above a previous high stand, the earlier deposits will be truncated, buried, or partially submerged. Conversely, a eustatic fall will leave the earlier deposits stranded as a terrace.

Almost all the soils on the Swan Coastal Plain, southwestern Australia, are formed by material deposited by rivers (fluvial, alluvial soils) and wind (eolian soils). The Yilgarn Block (part of the West Australian craton), east of the Darling Scarp bordering the Swan Coastal Plain, was epeirogenically uplifted as part of cymatogenic movement and warping about 40–50 million years ago. This tectonic uplift increased the

differential between base level and the plateau surface causing erosion of the craton by rivers and streams. Detritus from deep chemical weathering (e.g., laterite profiles) or saprolite was either deposited onto the Swan Coastal Plain or carried to the sea (Fairbridge and Finkl 1980). The eroded material formed new parent materials for soil development on the Plain.

Broadly there are thus two major groups of soils on the Swan Coastal Plain. The first is a series of dune systems near the coast formed as a result of deposits from the sea, and including material originally derived from erosion of the Yilgarn block mixed with marine sediments. Once formed, the dunes can be eroded by strong winds. The sand is mostly eroded from the dunes nearest the coast, and is redeposited on the dunes further inland. The second major soil types are a series of soils formed by deposits directly eroded from the Yilgarn block and which comprise soils on an alluvial (Pinjarra) plain, occurring between the dunes and the scarp. There is a third, very narrow strip of soil, called the Ridge Hill Shelf, next to the scarp, formed from material eroded from the scarp.

Although different conditions prevail in different locations, some general principles emerge from regional comparisons. It is impossible to review the development of all coasts, but some cases illustrate the salient principles that are involved.

Soils of Coastal Dune Systems

Several distinct features characterize coastal sand dune systems (e.g., Cooper 1966; Nordstrom et al. 1990). The dunes themselves often show a regular succession from the more active and unstable foredunes at the back of the beach to older, more stable vegetated dunes inland (cf. Figs. 6 and 7). The other major element is the system of slacks or swales (low-lying, narrow coastal wetlands) that occur between the main dune or beach-ridge trends (cf. Fig. 2). Dune slacks are low-lying areas within dune systems that are seasonally flooded and where nutrient levels are low. They occur primarily on the larger dune systems in the United Kingdom, for example, especially in the west and north where the wetter climate favors their development when compared with the generally warmer and/or drier dune systems of continental Europe. The range of plant communities found in slacks is considerable and depends on the structure of the dune system, the successional stage of the dune slack, the chemical composition of the dune sand, and the prevailing climatic conditions. When a coastal dune forms, dune sand accumulates seaward if there is an abundant sediment supply or the dune grows to its maximum height and then moves landwards (Gerrard 1981). If the dunes are carbonaceous, leaching progresses with increasing age and when coupled with an increase in organic matter, there is a transition in soil reaction

from alkaline to acid conditions. The rate of leaching is initially rapid but declines as the amount of carbonate, rather than the amount of rainfall, becomes a limiting factor. The rate of leaching in the early stages also depends on the nature of the shell fragments and matrix grain size, being slower if the fragments are large. The process is temporarily halted if material is added to the dune system by wind action because leached and organic-rich layers are buried.

Differences in soil moisture between the younger and older dunes, and between dunes and swales direct soil formation in regard to oxidation of organic matter (Gerrard 1981). Young swales are initiated with a considerable advantage because nutrients and other minerals accumulate in the wetlands, a situation that is ultimately reflected in the soil profiles and pH values. The rate of increase of organic content depends on the initial lime content of the dunes. On lime-deficient dunes, early colonization by vegetation takes place with a concomitant increase in the rate of litter accumulation. As more acid conditions prevail, litter breakdown is inhibited and organic matter build up is promoted.

Swan Coastal Plain, Southwestern Australia

The tectonically stable limestone coastline of southwestern Australia is well suited to preservation signals of former sea levels. The eustatic record is limited by the time frame when lithified dunes formed in middle to late Pleistocene times. Fairbridge (1961) interpreted the field evidence in calcarenite sequences and Paleosols as fluctuation of sea-level against the

backdrop of a stable landmass, incorporating the curve in a proposed global standard. The sequence of former sea level stands was based on wave cut benches, submerged and stranded beach cobble deposits, karst, and marine bands associated with the Pleistocene limestone.

The Swan Coastal Plain in southwestern Western Australia presents a complex chronosequence of mid-Pleistocene to Holocene dune soils, the oldest interior sets (Bassendean dunes of siliceous sands with Podzol morphology) merging with a series of laterite-covered spurs of the Ridge Hill Shelf that form the foothills of the Darling Scarp (McArthur and Bettenay 1979). The Bassendean Dune System is the oldest of the dune systems of marine origin and today consists of a series of low hills varying from 20 m to almost flat, which rise to 110 m above sea level. They were formed as a belt of coastal dunes and associated shoreline deposits that accumulated mostly during a period of high sea level, about 115,000 years ago during the Riss-Würm interglacial period (McArthur and Bettenay 1979). Secondary or epidiagenetic calcrete commonly occurs in an interdunal area (between the Spearwood and Quindalup dunes), which is marine or estuarine in origin, as a cap on top of fossiliferous limestone that in turn is covered by a shallow sandy soil. The Swan Coastal Plain contains alluvium and dune sands (McArthur and Bettenay 1960). The fluvial deposits, displayed as coalesced alluvial fans, are of different ages and limited to the west by sequences of coastal dunes. The oldest Bassendean dune sequence is siliceous and Podzols have developed on them. The next Spearwood sequence has a core of eolianite beneath the weakly podzolized siliceous sand (Fig. 8).

Coastal Soils, Fig. 8 Exposure of the Spearwood Dune System on the Swan Coastal Plain, southwestern Western Australia, near Perth. This dune system was formed during the Würm I interstadial about 80,000 years ago. The surface soils are leached and the carbonate has been precipitated below to form layers and columns of hard, compact limestone deposited from solution by percolating waters. Infilled solution pipes are clearly evident in these calcareous dunes that show multiple phases of solution and precipitation, erosion and deposition, and soil formation. Many solution pipes contain a thin rind of travertine in addition to being infilled with Terra Rosa type soil materials.



The Quindalup dunes are calcareous and sequentially the youngest. Properties of the dune soils change in an orderly fashion with age. Assuming that the oldest was originally calcareous, there is progressive loss of carbonate, a lowering of the pH, and formation of Podzols. Between the two younger dune sequences is a zone that is marine in origin.

In the southern extension of the Swan Coastal Plain, the Quindalup dune systems extends along the coast and are backed by soils of the Stirling Swale, a marsh environment, and soils developed on older dunes farther inland (McArthur and Bettenay 1956). Alluvial and swamp soils occur still inland on the coastal plain. This sequence of dune systems, often calcareous near the coast and siliceous on the landward side of the coastal plain, separated by intervening swales is typical of many coastal plains around the world. Alluvial soils truncate these sequences as rivers make their way to the coast and swamps often prevail in older degraded dune systems.

Gippsland Area, Victoria, Australia

The coastal zone south of the Diving Range in Victoria is separated from the sea by a narrow sand barrier (Ward 1977). Generally characterized as extensive lowland with broad, low-lying sand plains and swamps around lagoons, 14 marine terraces elsewhere rise in shallow steps to the 130 m contour. Dune sands form low ridges on the marine terraces and alluvial terraces beside the main rivers terminate at former shorelines defined by stranded marine terraces, except for the youngest one that passes beneath present sea level. Three groups of soils are defined whether the parent sediment is eolian sand, alluvium, or stranded marine terraces. The soils on the eolian sands show progressive development of Spodosol characteristics when viewed in order of age. In the other groups, the most striking feature is the progressive development with increasing age of a contrast in texture between solum and subsoil. Changes in clay mineral suites and chemical composition accompany these changes in morphology.

Southeast South Australia, Australia

The most widespread soils of the Kingston-Mt. Gambier region in southeast South Australia are derived mainly from calcareous beach sands (Podzols and Terra Rossa profiles) or from estuarine and lacustrine deposits (Groundwater Rendzina and Solodized Solonetz profiles) (Blackburn et al. 1965). The calcareous beach sands occur in about 30 stranded Quaternary beach ridges that comprise a soil chronosequence with the oldest soils occurring more than 80 km inland from the present coast. Variation in morphological properties of the main kinds of soils shows gradual transition in age sequence

that is represented by depth of red soil, thickness of kunkar, and occurrence of ferruginous concretions (iron-rich pisolites). The time span for soil formation for some of the older soils appears to be on the order of several hundred thousand years (Blackburn et al. 1965). Most of the ridges are consolidated, immobile structures on which soil profiles show little evidence of truncation. In some places there is evidence of truncation in the form of secondary siliceous sand dunes and exposed kunkar. Most of the older landsurfaces are represented by non-truncated soil profiles of sand Podzolics.

Soils of Polar Coastal Plains

Cold soils occupy roughly one-third of the world's land area (Rieger 1983). Maritime and continental climatic types are recognized, but each has varying degrees of expression. In maritime types, precipitation is high and uniformly distributed throughout the year, and temperature differences between winter and summer are minimal. Many sediments and soils in Polar regions have a complex history, particularly as they relate to cyclic deposits, glacial activity, wind action, solifluction, as well as cryogenic processes (Tedrow 1977; Allard 2001). The coastal plain on the northern tip of Alaska is flat and the drainage pattern is poorly integrated. Barrow is a low-relief emergent region where lakes, marshes, or former lake basins occupy 50–75% of the region. The remainder of the coastal plain consists of initial surface residuals, areas that do not show evidence of thaw-lake activity. Black (1964) reports that the complexity of the surficial materials near Barrow makes it difficult to subdivide the geologic deposits. Beaches, bars, distributary deposits, and deltaic deposits are interspersed with shallow-water marine sediments and lagoonal, lacustrine, and fluvial deposits. The extremely low ground and air temperatures subdue chemical and biological activity that inhibits pedogenesis. Studying weathering in northeastern Greenland, Washburn (1969, 1973) defines the major weathering processes in terms of oxidation, formation of desert varnish, case hardening, rillenstein formation, carbonate coating, salt wedging, granular disintegration, differential weathering, exfoliation, cavernous weathering, frost cracking, and frost wedging. More recently, Rieger (1983) pointed out the morphological impacts of freeze-thaw cycles, hydrostatic mounding and fissuring, and cryoturbation (viz. turbation accompanied by icewedge formation), on soil formation in cold regions. Study of cryoturbated Paleosols associated with past ice-wedge activity on late-Holocene sandy fluvial terraces in northern Québec (e.g., Allard 2001), for example, corresponds with paleoclimatic information obtained from existing Arctic-wide and regional proxy records. The Little Ice Age stands out as a period of intense ice-wedge activity, followed by a warm thawing-out interval during the first half of the twentieth century. From AD 1946 to

1991, climatic cooling reactivated ice-wedge formation (Allard 2001). In some situations, especially in the main coastal tundra zone, some of the physical and chemical weathering processes are inseparable from incipient soil-forming processes (Tedrow 1977). Due to the adverse and specialized conditions in these cold lands, researchers recommend different kinds of soil classification systems, which are reviewed in Tedrow (1977). Modern soil classification systems tend to incorporate soil on cold coastal plains as part of more comprehensive approaches, rather than have separate systems for specialized soil environments. In the Canadian system (Clayton et al. 1977), for example, those cold region coastal soils that show weak profile development fall into the Podzolic (Humo-Ferric Podzol) great group. Mostly Spodic Cryopsamments (in *Soil Taxonomy*) and the Brunisolic order (Cambisols in FAO/UNESCO; Inceptisols, excluding Aquepts in *Soil Taxonomy*) provide for the more normal soils. The Regosolic order (equates to the concept of Entisol, excluding Aquepts, in *Soil Taxonomy*) provides for various soils with little or no profile development. Gleysolic orders (Aquepts, Aquepts, and some Fluvaquepts in *Soil Taxonomy*) include the gamut of poorly drained mineral soils with an organic layer up to 30 cm thick. Within the Canadian system, *cry* is attached to the names of 10 subgroups. The names Cryic Regosol and Cryic Gleysol indicate similarity to Regosol and Gleysol but also to those soils that are underlain by permafrost within a depth of 100 cm from the surface. The major genetic soils (excluding Ranker) of the tundra zone can be arranged in catenary form as described by Tedrow (1977) viz. Arctic Brown, Upland Tundra, Meadow Tundra, and Bog soil. The Arctic Brown represents zonal soils whereas the Tundra soil is indicative of hydromorphic conditions. Organic soils occur in low-lying, poorly drained coastal areas referred to as bogs, swamps, marshes, or muskegs. Since many coastal landscapes have been free of glacial ice within the Holocene (past 10,000 years) there are many depressions in the landscape, water-impoundment areas, and flat plains with sluggish surface drainage. Recently deglaciated landscapes tend to exhibit poorly developed surface drainage patterns that result in water-saturated environments and the formation of Bog soils. Russian investigators commonly refer to the presence of organic deposits in sectors that have “sculpture-accumulative” relief (Tedrow 1977). Where coastlines are elevated, the organic deposits are usually intermixed with windblown sand. South of Barrow in the Meade River basin, surficial deposits have a sandy character with numerous blowouts, sand dunes, and vegetative mats that are buried by blowing sand. The coastal plain west of the Colville Delta contains geologic materials that incorporate a marine facies with dunes and loess. Because floodplains and sandbars are exposed to the wind action in summer, soils in the vicinity of coastal rivers and deltas contain admixtures of windblown sediments. The sand and silt is blown inland by onshore

winds and deposited along the banks. Soil properties are characteristic for a xeric, willow-vegetated dune system along the river but where the dunes merge with tundra inland, the terrain is poorly drained and wetland soils develop.

Interesting organic-rich soils of cold regions are the ornithogenic soils of coastal continental Antarctica (Casey Station, Wilkes Land) where penguin guano plays an important role in nutrient cycles in the ecosystem (Beyer et al. 1997). During the breeding season, seabirds bring huge amounts of organic matter from the sea to the land and this organic matter has accumulated to considerable thickness and consists of a friable layer of droppings, feathers, shells, bones, and dead penguins (Ugolini 1972; Tedrow 1977). On Signy Island (maritime Antarctica), 5% of the total snow-free surface is made up of ornithogenic soils from penguin guano. The chemical properties of the ornithogenic soils changes rapidly after the abandonment of the rookeries by the penguins, mainly because of CO_2^- evolution and N and P release in the sea (Tatur 1989). Other processes include the formation of oxalic acids in the first few centimeters and simultaneous concentration of recalcitrant soil compounds such as chitin, urates, and phosphate minerals. Leaching of iron and organic P fractions and the accumulation into the subsoil in maritime Antarctica produces a spodic Bh dark brownish (5 YR 3/1.5) horizon under a pale gray (10 YR 5/2) AE horizon, which are typical of Podzols in temperate regions (Beyer et al. 1997).

Coastal Barriers and Soils as Transitional Biophysical Systems

Barrier systems are widespread throughout coastal areas (Jelgersma et al. 1970; Leatherman 1979) and form the parent materials of soils. The barrier usually forms offshore, and a lagoon or backbarrier flat develops between the main barrier and the mainland; barrier-lagoon systems usually show transgressive, stationary, and regressive facies assemblages as seen along the coast of China (Li and Wang 1991) and elsewhere. Distinctive soil suites similar in age develop on the barrier and the backbarrier flats (Daniels and Hammer 1992). Some barriers form on the mainland, so only the soils seaward and those in the geomorphically equivalent estuaries or lagoons are the same age as the barrier. Goodwin (1987) describing coastal soil-landform relationships, reports that on the Coastal Plain of the Carolinas in the United States, soils on the landward Talbot Plain are older than the soils on the seaward Arapahoe Ridge or the Pamlico Plain. The Arapahoe Ridge and its association with the Talbot Plain is an example of why stratigraphic relations must be proven because the form alone often leads to incorrect conclusions.

Barrier systems may be wide, such as those in South Carolina, or very narrow as the Arapahoe Ridge in North

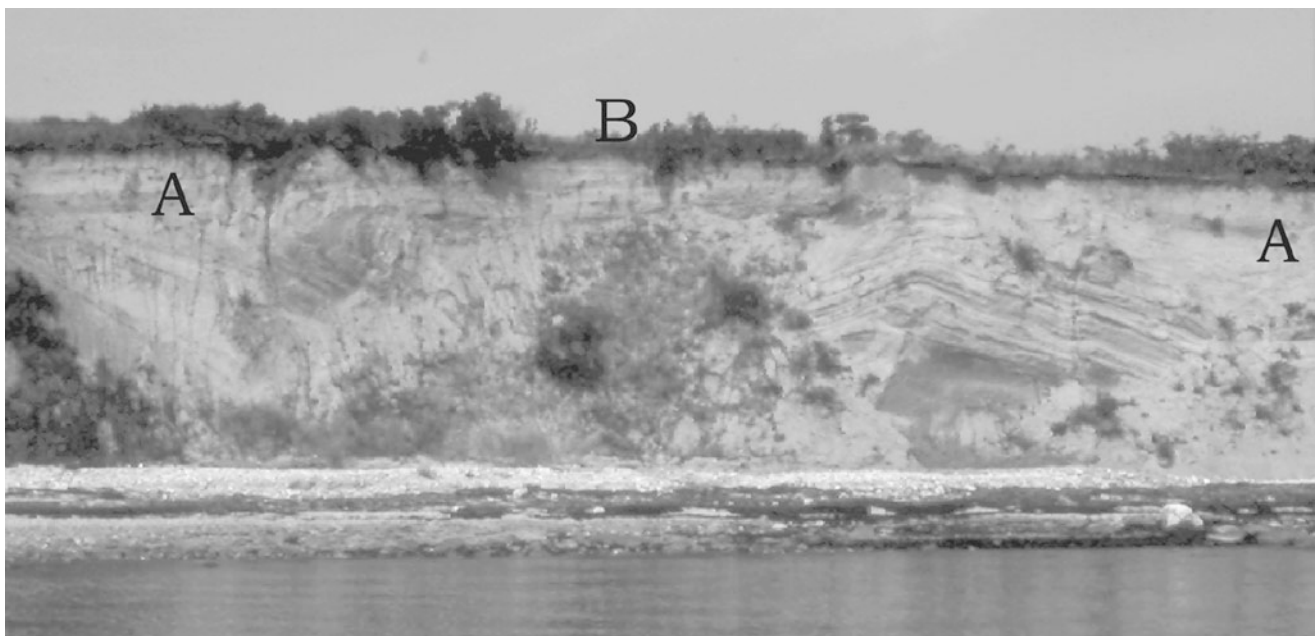
Carolina (Goodwin 1987) or the Atlantic Coastal Ridge in southeastern Florida (Finkl 1994). These systems may be very smooth, broad flats with little local relief, or irregular as in prograding beach ridge systems (e.g., Tanner 1995; Rink and Forrest 2004). Most barrier systems are a series of barrier and backbarrier flats where the seaward soils tend to develop on sandy parent materials.

Soils formed in the barrier sediments typically are 80–95% sand, with fine and medium sands dominating (Markewich et al. 1986). Spodic or humate horizons are common in the sandy barriers; in the Netherlands, peat and gyttia interfinger with older dune systems and on top of the Older Dunes there is a strongly developed Podzolic soil (Jelgersma et al. 1970). The Younger Dunes began to form in the twelfth century and are weakly podzolized. The spodic horizons may be very few centimeters thick beneath dryer sites, or 6 or 7 m thick beneath broad flats with deep-water movement. Most soils on the Arapahoe Ridge have spodic horizons, as do soils on the beach ridge complex. The soils on the beach ridge are complex sands. The A1 and spodic horizons on the ridge are often discontinuous and thin. The soils in the swales have thick organic rich A horizons and distinct spodic horizons. Soils and vegetation change abruptly in response to differing hydrological gradients across the ridge and swale sequence.

Soils on the backbarrier flats can have a wide range of textures. Many backbarrier systems are extensive, have a nearly level surface, and have finer textures than the adjacent barrier.

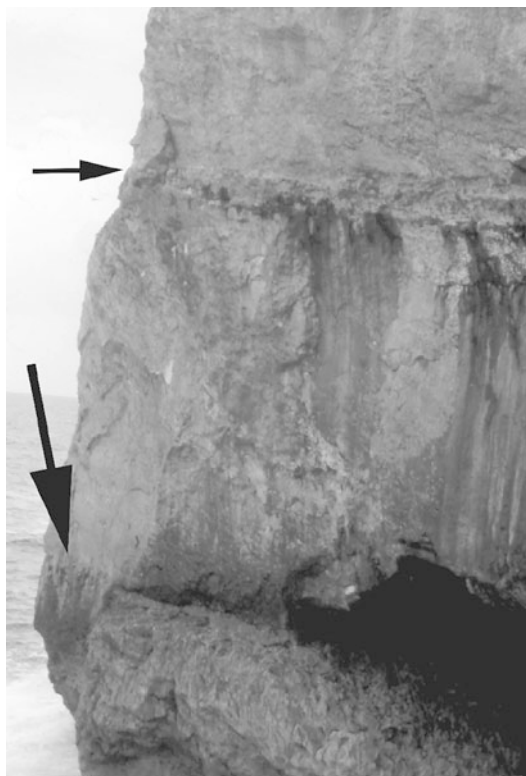
Coastal Paleosols

Paleosols are soils that formed on landscapes of the past (Yaalon 1971). They are identified on the basis of soil-forming processes, which produce the soil profile morphology, that no longer operate due to changes in climate (Fig. 9), local environmental conditions, or because of burial by younger (terrestrial or marine) sediments (cf. Fig. 9), lava, or inundation (drowning) by the sea. Ruhe (1965) recognized three main types of Paleosols: buried, exhumed, and relict. *Buried Paleosols* form at the landsurface but are subsequently covered by sediment or other materials, a process that separates and removes the soil from the contemporary soil-forming zone (Fig. 10). The depth of burial or integrity of the covering layer (e.g., lava) must suspend the soil-forming processes and yet be rapid-enough to prevent development of a Cumulic soil profile. Common examples occur in coastal dune fields, on overwashed barrier flats or spits, or on chenier plains where the rate of episodic shifting sands is slow-enough for soil formation to proceed and the new material



Coastal Soils, Fig. 9 Weakly developed phases of pedogenesis in the disturbed upper part of a glacial till associated with the Ronkonkoma Moraine on Long Island, New York, USA. Two phases of soil formation (layers A and B), separated by intervening unweathered reworked sediments, occur near the upper part of the moraine, which advanced about 21,000 years ago as part of the latest Wisconsin glaciation cycle. Soil layer (A) is a buried soil or paleosol. Soil layer (B) is a contemporary

surface soil. This most recent phase of soil formation is represented by an organic-rich accumulation at the surface and melanization of underlying materials. The translocation of organic compounds and oxidation of iron imparts a darker color to the subsoil. View is from Long Island Sound toward an erosional exposure that provides morainic sediments and weathered materials to the littoral system.



Coastal Soils, Fig. 10 Paleosol bands (arrows) within a calcarenite sequence on the Atlantic coast of northern Eleuthera Island, Bahamas. These weak phases of soil formation took place during stable phases with the Wisconsin glacial cycle. Reactivation of dune formation imitated additional cycles of deposition followed by stabilization and soil formation. The dark color of the soil bands is due to oxidative weathering of iron-rich minerals that accumulated in the profile by eolian accretion. Note that the seaward portion of the dunes and intercalated soils have been eroded, leaving soil layers and dune stratification projecting outward to where the dunes once stood.

to be incorporated into the soil profile. Epidiagenesis is common and resulting changes in original soil properties and horizons (formed prior to burial) must be considered in interpretation of Paleosol composition. *Exhumed Paleosols* occur at the land surface when a buried Paleosol is uncovered by erosion of the blanketing material. These soils may be out-of-phase with existing environmental conditions, as they occur at the time of reexposure. Exhumed Paleosols are commonly found in calcarenite sequences or active (including periodically reactivated) dune environments. *Relict Paleosols* formed on preexisting landscapes but were never buried all the while environmental conditions changed to a different regime, leaving behind weathering products that characterized prior landscape conditions. Common examples occur in a variety of landscapes, but those associated with ancient tropical and subtropical land surfaces are exemplars viz. *Sequisols*, *Ferralsols*, *Alisols*, *Acrisols*, *Luvisols*, and *Lixisols*.

A common example of buried coastal Paleosols occurs in horizontally bedded sequences, as in tropical southern Florida where there are five distinct marine units punctuated by episodes of subaerial weathering (Enos and Perkins 1977). Exposure to the atmosphere and chemical weathering of the exposed rock by dissolution resulted in the formation of discontinuous bands of dense caliche-type crusts (pedogenic calcrete as also reported in the Bahamas by Brown 1986), Paleosols, freshwater limestone, and laminated crusts (Enos and Perkins 1977; Beach 1982). Identification of subaerial exposure surfaces is an accepted means of identifying boundaries (unconformities) between different stratigraphic subdivisions of the Quaternary units. Interestingly, some of these weathering surfaces are radioactive and referred to as Secondary Depositional Crusts (SDC) (Krupa 1999). These SDCs are typically associated with increased natural gamma radiation immediately below subaerial exposure surfaces. These SDCs seem to be related to the paleo groundwater interface (Krupa 1999; Harvey et al. 2002), which was formed by residual downward percolation of rainfall through the vadose zone. The SDCs exhibit elemental abundances (e.g., Ca, Mg, Al, Sr, Fe, U, Th, and K) above the natural background level for the area.

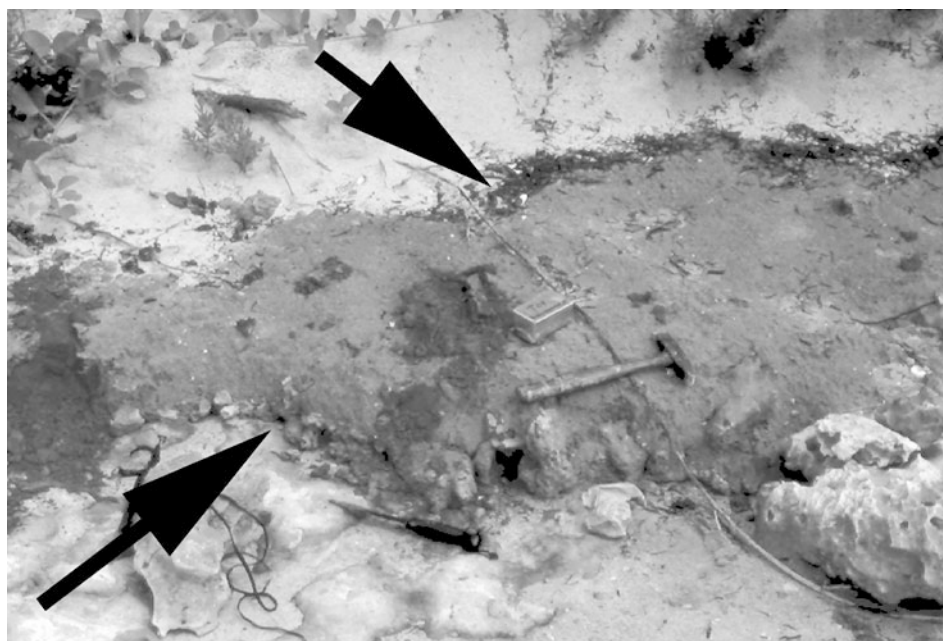
The Bermuda record is a complex mosaic of carbonate eolianites, shoreline-marine limestones, calcarenite *protosols* ("accretionary Paleosols"), and Terra Rossa Paleosols (Vacher and Hearty 1989; Herwitz et al. 1996). Sayles (1931) was among the first to recognize that alternation of marine limestones, eolianites, and Paleosols in local sections provides a record of Pleistocene sea-level fluctuations. Bretz (1960), drawing diverse observations into a comprehensive history, established the correlation that eolian and marine limestones were deposited during Pleistocene interglacials when the shallow platform (−20 m) surrounding Bermuda was submerged and that Terra Rossa Paleosols formed during glacial-age platform emergences. Solutional depressions often formed in the eolianites subsequent to dune stabilization and weathering to produce solution holes or pipes. These depressions were sometimes infilled with fine-grained sediment (Fe-rich dust) that weathered to red soils or Terra Rossa (Herwitz and Muhs 1995) (cf. Fig. 8). The recurrent stratigraphic pattern, as reported by Vacher and Hearty (1989), features an erosional coastline marine deposit overlain by a regressive eolianite. Where the interglacial complex is entirely exposed above present sea level, high-stand erosional coastline deposits can be traced down-dip into laterally extensive beach deposits of a depositional coastline. The beach deposits thus grade upward and laterally into eolianite, which in turn extends landward over the marine deposit and oversteps the erosional coastline deposits.

Bahamian paleosols, by way of another example of buried and exhumed Paleosols (Figs. 11 and 12), are predominantly

Coastal Soils, Fig. 11 Buried paleosol in the supratidal zone on the northeastern Atlantic coast of Eleuthera Island, Bahamas. This buried Paleosol is now partially exposed along the landward margin of an abrasion platform, indicating active erosion of dune cover sands.



Coastal Soils, Fig. 12 Close-up shot giving details of the buried soil layer seen in panoramic view in Fig. 11. The abrupt stratigraphic boundaries above and below the soil layers (arrows mark boundaries) and contrasting nature of all three materials verifies the independence of the soil layer, now partially truncated leaving just the stump of a former profile. The hammer and Kubiena box (for collection of undisturbed, oriented samples) are for scale.



calcium carbonate and contain minor amounts of mineral-insoluble residue (quartz, plagioclase, and clay minerals) (Boardman et al. 1995). There are at least four megascopically distinct Paleosols: laminated crusts, homogeneous crusts, breccia/conglomerate, and homogeneous matrix. All four types of Bahamian Paleosols and all mineralogic varieties of paleosols are found on stratigraphically equivalent rock units (Boardman et al. 1995). The number and character of Paleosol-bound parasequences on New Providence Island (Nassau), Bahamas, suggest that at least

five interglacial cycles are distinguishable. Whole-rock amino acid racemization (AAR) ratios confirm the stratigraphic sequence and indicate that isotope Stages 1 through 11 are likely included among eight depositional packages (Hearty and Kindler 1997). The Bahamian stratigraphic and pedogenic sequences, not unlike other calcarenite environments elsewhere, are complicated by the intricate stacking of similar materials in complex geomorphological settings that respond to fluctuating climates and sea levels (e.g., Carew and Mylroie 1994).

Similar late Pleistocene- Holocene coastal dune sequences and buried soils are described for the southeast Queensland coast by Ward and Grimes (1987), who suggest that Holocene sea level soon after 3,900 radio-carbon years BP was lower than present.

In addition to calcareous precipitates, other kinds of indurated materials related to ferruginous crusts and intensely weatherized materials are also common on coastal plains. Delaney (1966), for example, describes interesting mixed and contra-genetic sequences of caliche and laterite (Fe-duricrust) on the Rio Grande do Sul of southeast Brazil and northern Uruguay. Here, Quaternary units consist of a perched blanket of sand (the Itapoã Formation), arkosic sands and gravels (Graxaim Formation), the Serra de Tapes Laterite (a soil- stratigraphic unit), and a marine quartzose sand (Chuí Formation). The relict Serra de Tapes Laterite is not forming today and must have formed during a climate with high temperature and moderately high precipitation, conditions that do not exist on the coastal plain today. Pleistocene caliche was formed in a climate almost opposite to that of the Serra de Tapes Laterite.

Similar kinds of relationships are found in cold regions where ice-permafrost sequences on the Siberian coastal plain, with large polygonal ice wedges, represent excellent archives for paleoenvironmental reconstruction. Such deposits contain numerous well-preserved records (ground ice, Paleosols, peat beds, fossils), which permit characterization of environmental conditions during a clearly defined period of the past 60 ka (Schirrmeister et al. 2002). Lake plains along the eastern shore of Lake Michigan provide analogous setups where buried soils occur in dune fields with cyclical development after relatively long (>100 year) periods of low sediment supply (Loope and Arbogast 2002). Because soil formation takes place during stable landscape conditions, the presence of buried soils facilitated construction of a general hypothesis for late Holocene interaction of foredune and perched dune models on this lake plain, as it relates to multicentury lake-level fluctuations. It seems, on the basis of this study of 32 dune fields and included Paleosols, that the modern landscape originated after the postglacial peak in lake level (i.e., after the Nipissing transgression, about 5,500–5,000 cal year BP) (Loope and Arbogast 2002).

The aforementioned descriptions feature Quaternary soils in contemporary coastal landscapes, but there are numerous other Paleosolic developments along ancient shores where old weathering sequences persist in the form of deep weathering profiles of the lateritic or bauxitic type in senile landscapes. Salients among these examples are found in the formation of lateritic bauxites in Cretaceous and Tertiary coastal plains. Bauxites commonly form elongate belts hundreds of kilometers long, parallel to lower Tertiary shorelines in India and South America (Valeton 1983). Typical sediment

associations are found in India, Africa, South and North America, that are characterized by: (1) red beds rich in detrital and dissolved material of reworked laterites, (2) lacustrine sediments and hypersaline precipitates, (3) lignites intercalated with marine clays, layers of siderite, pyrite, maracasite, and jarosite, and (4) marine chemical sediments rich in oolitic iron ores or glauconite. The geographic relationship between in situ lateritic bauxites and the shoreline at the time of bauxite formation are interesting, if not somewhat controversial. Occurrences of bauxites residing parallel to ancient coastlines is documented for the karst bauxites of the Mediterranean (Southern France, Greece, Yugoslavia). Most of these types of lateritic bauxites, which are soft at the time of formation, belong to the Aquox subgroup of the group of Oxisols in *Soil Taxonomy*. The Fe-rich parts eventually form hard ferricretes whereas the Al-rich parts form hard alucretes (Goudie 1973).

Alternating sequences of landscape stability and instability, as related to phases of soil formation in shield areas, are comprehended in the *cratonic regime* (Fairbridge and Finkl 1980). This model recognizes a thalassocratic-biostatic condition (a long-term stable phase) during which prolonged deep chemical weathering takes place. In contrast, the epeirocratic-rhexistatic condition is a shorter-term unstable phase when there is general ecological stress and erosion of weathering profiles. The dual-phased, long-term-short-term regime was developed from experience on the West Australian craton (Yilgarn Block) but is applicable to other cratonic margins around the world. The significance to coastal soils relates to occurrences of laterites, bauxites, and associated materials along contemporary coastlines and ancient shores. Phases of soil formation are related to landscape stability and high sea levels; erosional phases where there is stripping of weathered materials from the land surface is associated with low sea levels. This sequence of events along cratonic margins produces a wide range of distinctly contrasting coastal Paleosols that include and which are derived from residual weathering that juxtaposes Quaternary coastal plain soils with relict Tertiary and older weathering profiles.

Epeirocratic-rhexistatic phases on the Yilgarn Block (West Australian craton) have, for example, resulted in the stripping of regolith materials from the craton surface to produce a series of etchplains that represent degrees of soil removal (Finkl and Churchward 1973; Finkl 1979). The least amount of stripping is associated with incipient etchplains, which occur on the seaward-most margins of the craton, along the north-south trending Indian Ocean coastal boundary and along the east-west trending Southern Ocean coastal fringe. Buried etchplains occur on the southern margin of the craton where deep chemical weathering profiles are intersected by present sea level and overrun by coastal dune systems.

Conclusion

Coastal soils are, in a certain sense, almost enigmatic because most soils occurring in coastal zones have complex histories and are related to environmental conditions that no longer predominate. This fact is perhaps not surprising because coasts themselves are extremely complex, resulting from combinations of land movements and fluctuations of sea level. Some coasts thus have histories of ingression where relative sea-level rise drowns preexisting landscapes (Kelletat 1995). The accompanying soils become drowned or are buried by younger coastal sediments and new phases of soil formation. In other cases, the seabed becomes emergent by sea-level fall or neotectonic uplift providing new parent materials for soil formation. In other situations, the land-sea boundary remains more or less stable long enough for soils to develop in coastal deposits such as mangrove muds. Whatever the mechanisms involved, there exists a diverse range of soils in coastal zones. Some are clearly “coastal” in the sense that they are closely related to present shoreline materials and conditions; they are therefore younger Holocene in age. Other soils occurring in coastal zones are not strictly “coastal” because they can form independently of maritime conditions. The complex array of coastal zone soils is not easily comprehended without careful study and some background in pedological concepts, principles, practices, and field experience. Complex terminologies of different soil classification systems also tend to confound all but the most intent coastal researchers who make the effort to ferret out the details. Coastal soils are an interesting topic that offers scope for studying coastal change or evolution as recorded in the pedological record.

It is seen that coastal zones contain the youngest soils on earth as well as some of the oldest. Newly exposed seafloor muds form parent materials for acid sulfate soils and Solonchaks, recently stabilized coastal dunes host youngest Holocene Regosols whereas some older dunes feature strongly differentiated Pleistocene Podzols. Holocene Gleysols and Histosols form in juxtaposed swales and slacks where drainage is poor. Planosols and Solonetz occur on marine terraces and deltaic regions along with Histosols, whereas Cryosols and Leptosols characterize cold region coastal lowlands. Developed on older coastal plains in the low to middle humid latitudes there is a range of pedogenic formation that includes Podzols, Planosols, Aliosols, Luvisols, and Arenosols and Calcisols in drier realms. Deltaic plains often feature Fluvisols, Planosols, Solonetz, and Histosols. Ancient surfaces truncated by coastal erosion, especially in intertropical regions, show Sesquisols, Ferralsols, Alisols, Acrisols, and Lixisols. Last but not least there are Anthrosols that characterize pedogenic regimes in coastal areas where there has been a long record of human habitation and cultivation of field crops. This description of coastal soils may seem like a long

digression, but it is actually a light vignette because the situation is far more complex than what has been alluded to here.

This overview nevertheless provides a fair indication of the range and complexity of pedogenesis in coastal zones as well as indicating the close interrelationships between coastal processes, landforms, vegetation, hydrology, maritime climates, sediments, and the imprint of biochemical weathering on coastal landscapes and seascapes. More importantly, however, is the fact that coastal soils provide much biophysical information that is useful in attempts to unravel some of the major events in coastal evolution, especially phases of soil formation that correspond to sea-level stillstands and landscape stability.

Cross-References

- ▶ Alluvial-Plain Coasts
- ▶ Barrier Island Landforms
- ▶ Beach Ridges
- ▶ Beachrock
- ▶ Bogs
- ▶ Changing Sea Levels
- ▶ Cheniers
- ▶ Coastal Sedimentary Facies
- ▶ Deltaic Ecology
- ▶ Dune Ridges
- ▶ Hydrology of the Coastal Zone
- ▶ Mangroves, Ecology
- ▶ Mangroves, Geomorphology
- ▶ Muddy Coasts
- ▶ Peat
- ▶ Salt Marsh
- ▶ Sandy Coasts
- ▶ Submerged Coasts
- ▶ Vegetated Coasts
- ▶ Weathering in the Coastal Zone
- ▶ Wetlands

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Coastal Subsidence

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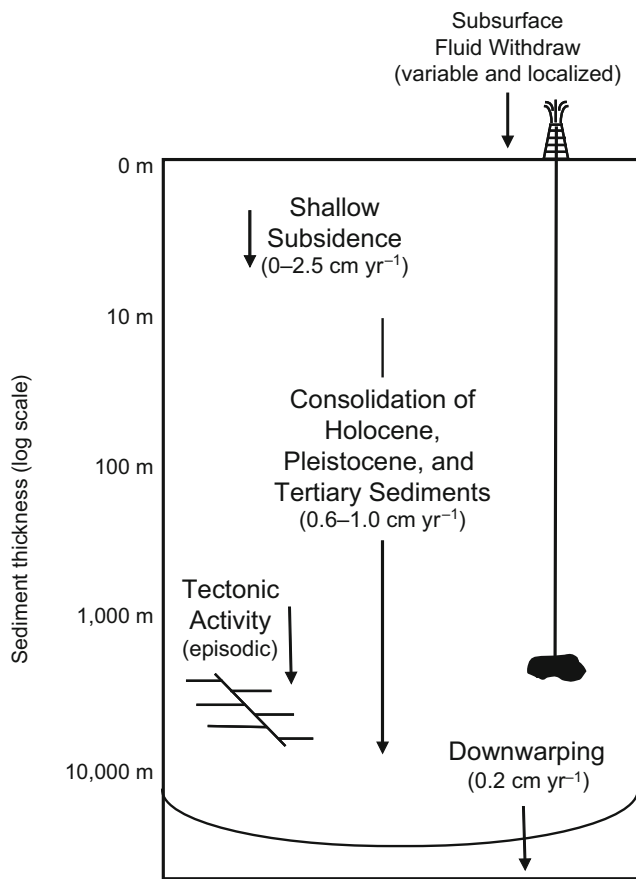
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Introduction and Terminology

Coastal subsidence, which can be described as the downward displacement of the land relative to sea level, often occurs in deltaic regions associated with riverine and estuarine sedimentation. These regions are commonly sites of thick deposits of Tertiary, Pleistocene, and

Holocene sediments, exceeding 12,000 m in the Mississippi River delta, for example (Penland et al. 1988). Coastal subsidence in these depositional zones are a function of five processes, each described in detail below: (1) downwarping, (2) tectonic activity, (3) consolidation of Tertiary, Pleistocene, and Holocene deposits, (4) shallow subsidence, and (5) underground water and gas extraction (Fig. 1). When only natural processes are considered, 1 + 2 + 3 define deep subsidence. Shallow and deep subsidence combined equal total subsidence (Cahoon et al. 1995).

Eustatic sea-level rise (the global sea-level rise due to changes in the volumes of glaciers and ice caps, and due to density-temperature relationships) can also be viewed as a “relative” land sinking or subsidence process. On a global scale, eustatic sea levels have risen approximately 125 m



Coastal Subsidence, Fig. 1 Processes contributing to total subsidence in coastal regions. Comments and numbers in parentheses indicate the relative magnitude of each process in the Mississippi River delta, a region where coastal subsidence has been intensely studied (modified and redrawn from Penland et al. 1988)

since the last glacial maximum 18,000 yr. BP. Current rates of eustatic sea-level rise are estimated at 0.15 cm yr^{-1} (Church et al. 2001). The term “relative sea-level rise” is used to describe deep subsidence plus eustatic sea-level rise because it is often difficult to differentiate deep subsidence rates from eustatic sea-level rise along the world’s coasts.

Downwarping

The weight of thick sediment sequences deposited during the Quaternary by major rivers that have discharged considerable sediment load, such as the Amazon, Yangtze, Mississippi, Yellow, and the Ganges/Brahmaputra has caused downwarping of the underlying crust (Fig. 2) (Coleman 1982; Chen et al. 2000). For example, in the Mississippi River delta, the combined weight of Cenozoic deposits accumulating in the geosyncline off the river mouth has resulted in the slow downwarping the Mesozoic basement. However, this process, estimated to be occurring at an average rate of

0.2 cm yr^{-1} , accounts for only a small fraction of total subsidence in the region (Penland et al. 1988).

Tectonic Activity

The tectonic-fault framework that exists in coastal regions can sometimes control the geomorphic configuration of the coast-delta extension (Coleman 1982; Chen and Stanley 1995) and the intensity of the fault motion plays an important role in influencing the overall subsidence rate (Stanley, 1988; Chen and Stanley 1995). On a large-scale, fault tilting can trigger the diversion of large rivers from one region of a coast to another. Consequently, the estuarine depocenter marked by the thickest sediment sequences, and the most active subsidence, switches laterally. The subsidence triggered by tectonism is a long-term and episodic, geologic processes that is difficult to distinguish from the subsidence caused by the long-term compaction of Cenozoic sediments. A regional estimate of tectonic subsidence along the coast can be made by evaluating subsurface data including petrologic, litho- and chronostratigraphic descriptions from a series of boreholes through sedimentary sections.

Consolidation of Tertiary, Pleistocene, and Holocene Deposits

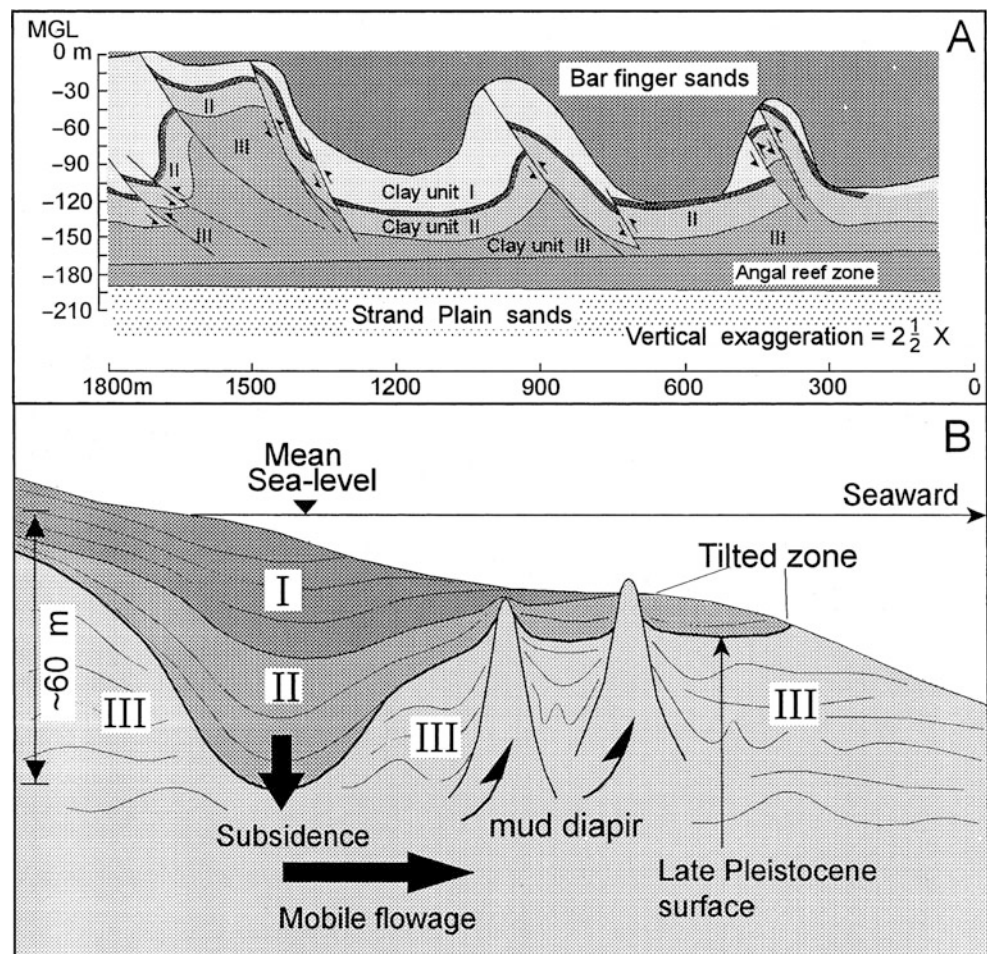
The consolidation of Tertiary and Pleistocene marine and riverine deposits can be a significant component of total subsidence. However, it is the consolidation of Recent sediments that is considered the primary cause of background subsidence in coastal regions. Large amounts of riverine sediments deposited over the past 15,000 years have accumulated along coastal depocenters resulting in high rates of subsidence, regionally. This consolidation, or compaction, is caused by the dewatering of sediments (primary consolidation), the rearrangement of sediment structure (secondary consolidation), and the decomposition of organic matter. In some parts of the Mississippi River delta, subsidence rates due to consolidation exceeds 0.6 cm yr^{-1} (Penland et al. 1988).

Shallow Subsidence

The compaction in the upper 5–10 m of the sediment contributes a shallow component that is often overlooked when calculating total subsidence in coastal regions (Cahoon et al. 1995). Shallow subsidence is the result of primary consolidation and the decomposition of organic matter. It is an especially important process in coastal systems under stress (from flooding or salt, for example) where below ground plant structures, such as roots and rhizomes, die and decompose,

Coastal Subsidence,

Fig. 2 Diagram showing the gravitation subsidence both in the Mississippi and Yangtze estuarine depocenters. (A) The Mississippi model indicating the more seaward displacement than the downward, resulting in tremendous reverse mud slump in front of the river mouth region (modified after Coleman 1982), and (B) The Yangtze model indicating downward and seaward sediment displacement



leading to rapid subsurface collapse. In the Mississippi River delta, for example, high rates of relative sea-level rise in the coastal marshes have lead to increasingly long periods of flooding, plant mortality, and rates of shallow subsidence that exceed 2.5 cm yr^{-1} .

Underground Water and Gas Extraction

In some coastal regions, large amounts of fresh groundwater have been extracted along the coast for industrial, agricultural, and domestic use. Rates of subsidence due to this process can be alarmingly high, although the effects are usually localized. For example, subsidence rates within the Po delta of Italy and the Yangtze delta of China have, in the some past years, exceeded 30 cm yr^{-1} , due largely to groundwater withdrawal (Cencini 1998; Chen 1998). The reinjection of salt water into the substrata, to replace extracted fluids, has been successfully used to slow down or even halt, but not reverse, extraction-induced subsidence. Hydrocarbon extraction along the coast has the same potential for causing land subsidence, but has been less studied.

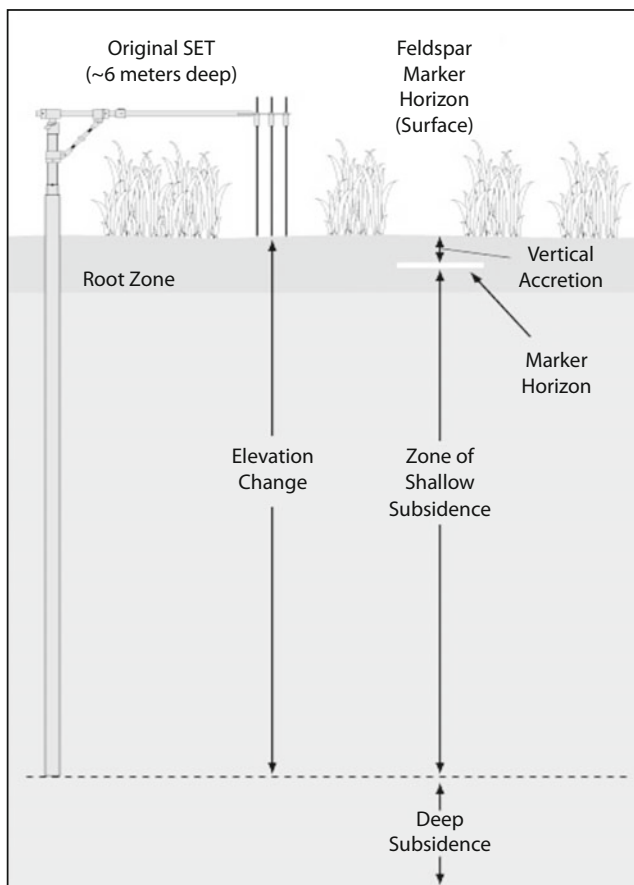
Methods of Measurement

Estimates of deep subsidence are usually based on long-term records from tide gauges that are mounted on stable piers, bridges, and pilings that extend through the shallow subsidence zone (and thus do not include shallow subsidence). A tidal gauge record spanning at least 18.6 years is required to factor out variations due to the moon's nodal cycle. Typically, mean annual or monthly water levels are regressed against time to yield a rate of relative sea-level rise (Emery and Aubrey 1991). To estimate the deep subsidence component of relative sea-level rise, the eustatic component (currently 0.15 cm yr^{-1}) is then subtracted. Eustatic sea-level rise is derived from the analysis of tide gauge data from coasts worldwide that are assumed to be geologically stable (Penland et al. 1988).

In addition, a radiocarbon-dated stratigraphy can help to calculate the net subsidence rate along the coast (Stanley 1988). Stratigraphic analysis of sediment borings can serve to define the subsurface configuration of Recent deposits. Isopachous maps of the Holocene sediment sections can also reveal thickened sediment sequences along the coast.

Moreover, high-resolution seismic profiles can provide evidence of Recent subsidence. Deformed late Quaternary strata, mud slumps, diapirs, tilted bedding, and gas-deformed structures in Holocene sequence imply remobilization of underconsolidated sediment within and adjacent to the estuarine depocenter (Chen et al. 2000). Deformation types and their specific position relative to the depocenter record the displacement of sediments as a result of compaction due to sediment overburden.

The tidal gauge technique for estimating subsidence (described above), and most estimates of subsidence reported in the literature, measure only deep subsidence, even though rates of shallow subsidence can greatly exceed the rates of either of these processes in some marshes (Cahoon et al. 1995). A recently developed field technique that uses a sediment marker horizon (such as feldspar clay) in conjunction with a surface elevation table (Fig. 3), an instrument that measures changes in elevation relative to a shallow subsurface datum, has made it possible to measure shallow subsidence (Cahoon et al. 1995).



Coastal Subsidence, Fig. 3 A surface elevation table (SET), shown in use with a feldspar marker horizon, used to estimate rates of shallow subsidence in coastal systems (Figure courtesy of Jim Lynch, and Don Cahoon, United States Geological Survey)

Subsidence and Its Implications

In stable or prograding coastal systems, relative sea-level rise is usually balanced by the accretion of fluvial or reworked marine sediments. However, extensive hydrologic alterations such as dams and levees have sharply reduced the amounts of fluvial sediments transported to coastal regions. As a result, saltwater intrusion, flooding, and coastal erosion has become much more prevalent in the past 100 years. In the Mississippi River delta plain, for example, land loss rates exceed $65 \text{ km}^2 \text{ yr}^{-1}$, due largely to a subsidence/accretion imbalance. To add to this problem, the eustatic sea-level rise component of relative sea-level rise is expected to accelerate over the next 100 years due to global warming (Church et al. 2001). Therefore, accurate measurements of coastal subsidence are vital for planning coastal protection in densely populated and low-lying region vulnerable to marine inundation.

Cross-References

- [Changing Sea Levels](#)
- [Coastal Sedimentary Facies](#)
- [Dams, Effect on Coasts](#)
- [Deltas](#)
- [Eustasy](#)
- [Isostasy](#)
- [Sea-Level and Climate Change](#)
- [Sedimentary Basins](#)
- [Submerged Coasts](#)
- [Submerging Coasts](#)
- [Tide Gauges](#)

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Coastal Temperature Trends

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Definition

Coastal water temperature trends – sea surface temperature values per unit time – are an indicator of global change.

Introduction

Few long-term records of coastal water temperatures are extant, but from many tide gauge sites of the U.S. Coast and Geodetic Survey (C&GS – now incorporated into the National Ocean Service of the National Oceanic and Atmospheric Administration, NOAA), there are handwritten observations covering much of the twentieth century. In the 1990s, NOAA switched to digital thermistor sea surface temperature measurements. Herein the digitized handwritten records are concatenated to the thermistor records, and century-scale SST trends are calculated.

Prior to the development of acoustic tide gauges and digital data-logging and transmission, trained tidal observers would visit C&GS primary tide gauge sites, typically on a daily basis. Their duties included comparisons between the marigram and the tide staff, measuring sea surface temperature and density and performing routine maintenance. The sea surface temperature (SST) values were recorded on standard forms, and mean monthly and annual values were calculated. These data forms were then archived and largely ignored. Only twenty or so C&GS tide gauge station temperature records of significant multidecadal length have been recovered and are analyzed in detail by Maul et al. (2001).

Tidal observers determined SST by lowering a bucket from the pier or other structure supporting the tide gauge, retrieving a sample of water, and measuring the temperature with a standard mercury-in-glass thermometer. Seawater density was also measured with a hydrometer (USC and GS

1929). In typical Coast and Geodetic Survey tradition, all instruments were numbered and routinely calibrated, with emphasis on data quality dating back to 1807 when President Thomas Jefferson founded the agency.

In the acoustic tide gauge, the thermistor is placed at the intake of the stilling well (vertical pipe that houses the acoustic hardware). Thus the temperature measured at high tide might be 10 m below the sea surface and at low tide 1 or 2 m below. If the water at the tide gauge site is vertically well-mixed, the SST and intake temperatures should be comparable. The thermistor data are archived every 6 min in concert with the water level measurement. Annual mean SSTs are calculated by averaging all 8760 hourly thermistor observations for a given year (8784 h in leap years). The thermistor data are available from NOAA at <https://tidesandcurrents.noaa.gov>.

Discussion

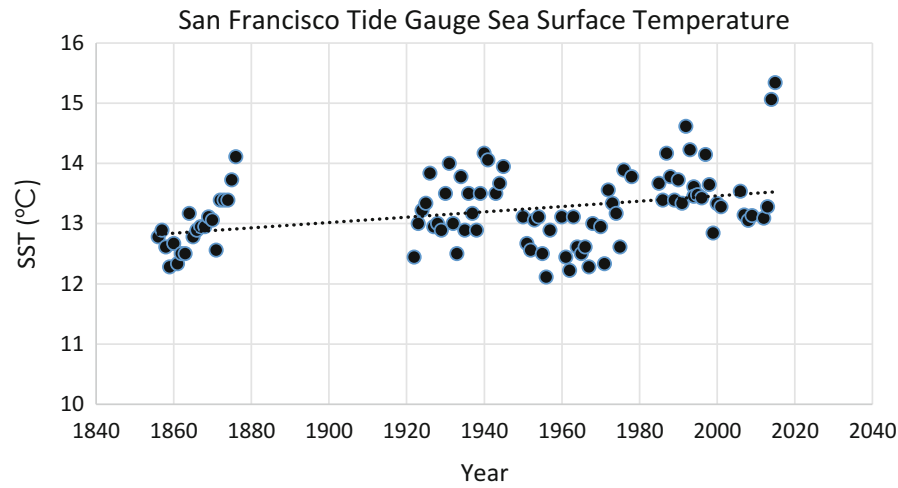
The longest record is from San Francisco, California. The historical C&GS SST data (1856–1994) are concatenated to the thermistor data (1995–2015) and are shown in Fig. 1. Typical of these data sets are gaps of various length. Some gaps are 1 or 2 years, others are decades long. The straight dotted line in Fig. 1 is the linear trend over the entire 1856–2015 record. Trends in annual mean SST are calculated by standard linear least-squares methods where an equation of the form $T = a \cdot t + b$ (where T is temperature, a is temperature trend ($\partial T / \partial t$) in °C per century, t is time in years, and b is the linear intercept) is fitted to the data.

Table 1 is the summary statistics of the longer SST records. The first column lists the sites, chosen to represent biogeographic regimes around the coastal United States with century-scale data. Columns 2, 3, and 4 are the years, number of years excluding gaps (N), and the trends $\pm$ the standard error of the trend of the historical C&GS data, respectively. Columns 5, 6, and 7 are the years, N , and trend statistics of the combined data. The thermistor data are only two decades long, a record far too short for trend analysis separately.

In general, concatenating the thermistor temperatures to the C&GS temperatures causes the trend to increase. Whether this is due to the change in measuring technology or coastal warming is uncertain. Lengthening the record on the other hand seems to decrease the uncertainty estimate of most trends. Baltimore, MD, and Galveston, TX, uncertainty both decrease from ± 0.5 °C per century to ± 0.3 °C per century. For the most part, the added thermistor data seems to blend well with the older thermometer data – there being no statistically significant difference in the mean temperatures in the 1980–2010 timeframe.

There is a nonsignificant degree of variation from locale to locale. The largest SST trend is at Baltimore, MD. This

Coastal Temperature Trends, Fig. 1 Time series plot of SST San Francisco, California. Note the missing data, which is typical of many such geophysical records, and the interannual variability



Coastal Temperature Trends, Table 1 Sea surface temperature (SST) trends from NOAA tide gauge sites

Station	Years	N	Trend ^a	decs	N	Trend ^a
Boston	1921–1994	73	0.5 ± 0.3	1921–2014	87	0.8 ± 0.3
Baltimore	1914–1993	65	0.9 ± 0.5	1914–2015	80	1.3 ± 0.3
Charleston	1922–1992	70	-0.1 ± 0.3	1922–2014	79	0.8 ± 0.3
Key West	1926–1994	36	0.0 ± 0.3	1926–2015	47	0.4 ± 0.2
Galveston	1922–1992	35	-0.9 ± 0.5	1922–2015	49	-0.2 ± 0.3
Los Angeles	1923–1991	40	0.8 ± 0.4	1924–2016	61	0.2 ± 0.4
San Francisco	1856–1994	79	0.4 ± 0.1	1856–2015	95	0.4 ± 0.1
Seattle	1923–1993	41	0.0 ± 0.4	1923–2015	61	0.0 ± 0.2
Seward	1926–1993	29	0.1 ± 0.4	1926–2013	44	0.8 ± 0.3

^a°C per century

unusually large trend, $a = 1.3 \pm 0.3$ °C/century, may be caused by the geography of Baltimore Harbor, by a general warming of Chesapeake Bay, and/or by urbanization of the marine environment. Nearby Boston ($a = 0.8 \pm 0.3$ °C/century) also shows significant warming as does Charleston, SC ($a = 0.8 \pm 0.3$ °C/century). Note however how the trend at Charleston has markedly changed after concatenating the thermistor data. The average Charleston temperature of the thermistor data (2001–2014) is 21.0 ± 0.6 °C, and the average of the juxtaposed 1980–1993 C&GS temperatures is 20.1 ± 0.5 °C. Although the difference between 21.0 ± 0.6 and 20.1 ± 0.5 is not statistically significant at the 95% confidence level, 2001–2014 is notably warmer. In the mid-1990s, the Atlantic Multidecadal Oscillation switched to a warm phase, which in part, may be the warming measured at these three tide gauges.

The Gulf of Mexico region, on the other hand, shows little statistically significant warming or cooling. At Key West, FL, the southeastern region’s most oceanic site, $a = 0.4 \pm 0.2$ °C/century, a slight warming trend. Galveston, TX, on the other hand, shows a small cooling trend of -0.2 ± 0.3 °C/century. As with Charleston, concatenating the thermistor data has markedly changed the trend at both Key West and at

Galveston. There is no evidence of acceleration in SST trend in these or any other station in this entry (i.e., fitting exponential or polynomial equations to these data does not explain additional variance r^2).

Along the US west coast, the SST trends among the sites are most similar: Los Angeles, CA ($a = 0.2 \pm 0.4$ °C/century); San Francisco, CA ($a = 0.4 \pm 0.1$ °C/century); and Seattle, WA ($a = 0.0 \pm 0.2$ °C/century). There is an important difference between Seattle, WA, and Seward, AK ($a = 0.8 \pm 0.3$ °C/century). Highly populated Seattle is well within Puget Sound with no large rivers to flush the estuary, and Seward (population 2,500) is at the head of Resurrection Bay – an oceanic environment surrounded by national forests. The zero SST trend at Seattle is not unusual among the west coast stations, but serves to illustrate the variability in seawater temperature trends from place to place. Seward is more influenced by the Pacific Decadal Oscillation, and care must be exercised when least-squares fitting a linear equation to oscillatory data as a trend will be calculated depending on the phase of the oscillation. West coast sites farther south are similarly influenced by ENSO (El Niño–Southern Oscillation) events and Atlantic sites by the Atlantic Multidecadal Oscillation.

Conclusion

This entry demonstrates the complex nature of water temperature change in the coastal United States using an inhomogeneous dataset. As with sea level change and other indicators of climate change, SST is seen to be site specific and that it is poor scientific practice to extrapolate to other regions of Earth. It also shows the challenge and value of data archeology (Maul et al. 2001). Many climatic records are not yet digitized and may be deteriorating or are already lost. While these data may not improve clarity in the climate change debate, they offer an important independent opportunity to learn more about the evolving nature of our planet's geophysical history.

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Coastal Upwelling and Downwelling

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Definition

Coastal upwelling and downwelling refer to the upward and downward transports of water mass, respectively. These vertical fluxes are the results of mass balance and are caused by the horizontal water motions of divergence and convergence. They occur in various scales of sizes and durations. Coastal upwelling is more noticeable because it brings cooler subsurface water upward to the surface in the case of a thermally stratified water column and is accompanied with primary productivity and abundance of fish resources. The phenomenon has considerable economic significance because it affects fisheries, weather, and oceanic currents in many parts of the world. This has prompted much interest in research on upwelling systems around the world. Some executed in the 1970s, including those under the umbrella of Coastal Upwelling Ecosystems Analysis (CUEA), were compiled through an International Symposium on Coastal Upwelling (Richards 1981). The upwelled water often forms a plume with fronts that can be identified by remote sensing technologies.

Upwelling and downwelling scales vary from localized episodes to large ocean areas. Upwelling speeds are very low and vary from 1 to 10 m/day, but upwelling regions cover thousands of square kilometers.

The following sections describe the physics and types of coastal upwelling and downwelling systems, major places of their occurrence, environmental consequences, mathematical models of upwelling motion resulting from longshore wind stress, and measurements.

Physics and Types

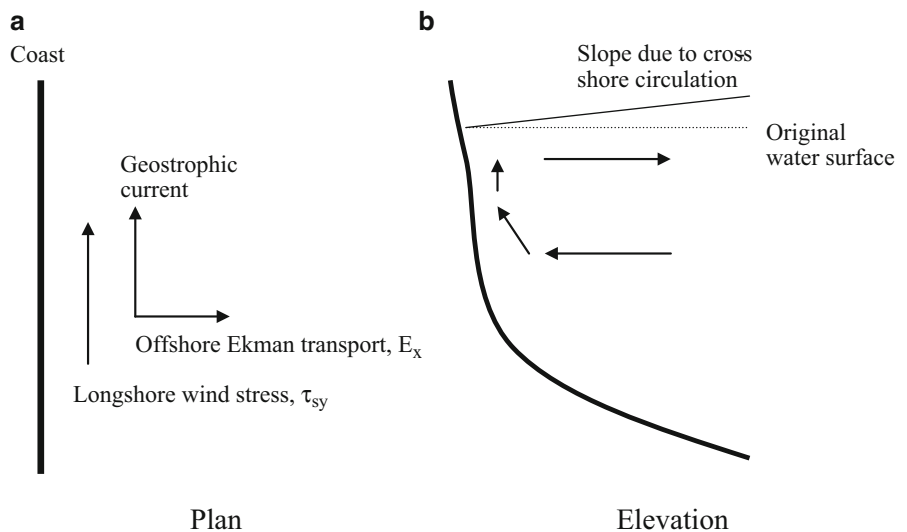
The upwelling and downwelling processes can be discussed under four fundamentally different initiation mechanisms.

Longshore Wind

The most important episodes of coastal upwelling and downwelling, in terms of size, duration, and economic importance, occur due to longshore wind. When wind blows parallel to a long shoreline for a considerable period of time (for a steady state to develop), a dynamic balance is established between the wind stress on the water surface and other hydrodynamic terms in the equation of motion. Figure 1 illustrates schematically the process of coastal upwelling development due to longshore wind. Two simplifications of the hydrodynamic equation can best explain the cross-shore motion and the resultant episodes of upwelling or downwelling. The first is popularly known as the Ekman (Swedish oceanographer Vagn Walfrid Ekman, 1874–1954) transport that results from a dynamic balance between the wind-induced frictional force at the water surface and the Coriolis (French mathematician Gustave de Coriolis, 1792–1843) force. Ekman first introduced this concept (Ekman 1905) to explain quantitatively the ice drift observations of Nansen (Norwegian explorer and scientist Fridtjof Nansen, 1861–1930) in 1893. Due to a balance between these forces, a mass transport of water takes place in the surface layer, which is directed to the right of the blowing wind in the Northern Hemisphere (to the left of the blowing wind in the Southern Hemisphere). Depending on the wind direction relative to the shoreline, water level is lowered or elevated along the shoreline, and mass conservation requirement causes vertical motions of upwelling or downwelling, respectively. For the northwesterlies along California and Oregon coast, this process creates an offshore transport of water mass, and underwater upwells to the surface (McEwen 1912) along the coast. The second simplified explanation of the same motion is based on a geostrophic balance between the pressure gradient force induced by cross-shore Ekman transport and the Coriolis force. The pressure gradient force arises due to the development of a slope in the water surface (Fig. 1).

Coastal Upwelling and Downwelling, Fig. 1

A schematic indication of the coastal upwelling processes initiated by longshore wind.
(a) Plan. (b) Elevation



Further discussion on these processes is given in the section on mathematical models of coastal upwelling and downwelling.

Offshore and Onshore Wind

On a more localized scale, upwelling and downwelling also occur due to offshore and onshore winds. Onshore winds pile up water along the coast, and downwelling occurs to satisfy the mass conservation requirement. The opposite happens, when an offshore wind lowers the water surface at the coast and the underwater upwells to the surface. Thomson (1981) and Henry and Murty (1972) reported occurrence of upwelling due to an offshore wind in the Departure Bay of Strait of Georgia. They found that due to upwelling, surface water temperature dropped by 4 °C within less than 24 h of offshore wind blowing. Field et al. (1981) measured upwelling and downwelling episodes in response to offshore and onshore windstorms, respectively, in a kelp bed in southern Benguela off Cape Peninsula in South Africa.

Underwater Ridges, Shoals, and Large-Scale Bedforms

When interrupted by underwater ridges, shoals and large-scale bedforms, the horizontal flow upwells to the surface in the form of a surface boil. Such local upwelling, especially when associated with large horizontal current, may occur in many places in response to large discontinuity of the seabed. Thomson (1981) reported several upwelling episodes in many passes of the Juan de Fuca Strait and Strait of Georgia along British Columbia coast.

Open Ocean Divergence and Convergence

From June to October, the southeast trade winds cause upwelling in the equatorial Pacific (Thomson 1981). The Coriolis force deflects the trade wind generated South Equatorial Current to northward in the north of the equator and to southward in the south of the equator. A belt of divergence, known as

equatorial divergence, is thus created that prompts upwelling of cold underwater to the surface. A similar divergence is caused at about 10° N latitude in response to Northeast Trade Winds. As a different process, divergence also occurs due to a moving pressure field associated with cyclones. A cyclonic low-pressure system causes surface water divergence, and upwelling occurs around the eye of the cyclone.

Geography

The following areas are some of the notable large-scale upwelling regions in the world (Bowden 1983; Smith 1983). These are the regions associated with anticyclonic high-pressure systems in the atmosphere:

1. In the North Pacific, between 25 and 45°N, along California and Oregon coast in response to equatorward California Current during summer northwesterlies
2. In the South Pacific, between 5 and 30°S, along Peruvian coast in response to equatorward Peru Current
3. In the North Atlantic, between 10 and 40°N, along the northwest coast of Africa in response to equatorward Canary Current
4. In the South Atlantic between 5 and 30°S, along the southwest coast of Africa in response to equatorward Benguela Current
5. Along the East Coast of Africa and south coast of Arabian Peninsula in response to poleward Somali Current during the southwest monsoon

Large-scale sea surface temperature anomalies covering the whole width of the Pacific Ocean – known as the El Niño-Southern Oscillation (ENSO) – may in part be linked to upwelling events.

Consequences

Coastal upwelling is a very well-known phenomenon because of its practical interest. The important consequences of upwelling are abundance of fish resources in the upwelling region, influence on coastal climate, development of sea fog, and intensification of existing coastal current.

Very often, the upwelled water brings upward with it a higher concentration of nutrient than the resident surface water. The reason for lower concentration in resident surface water is that over time the available nutrients are used up by the growth of phytoplankton. The upwelled nutrients are usually nitrate, phosphate, and silicate salts. The enhanced nutrient content of the upwelled surface water helps the growth of phytoplankton that supports a greater zooplankton concentration. This primary productivity maintains a high fish population, and the whole interdependence of lives in the region is known as upwelling ecosystem. Most of the important marine fisheries in the world are found in regions of upwelling. Although coastal upwelling areas comprise only 1% of the total area of the ocean, they are estimated to produce more than 50% of the total marine fish harvest.

In an upwelling region, the cold upwelled water causes a drop in air temperature of the region, and sea fog develops locally in the presence of moist air. Thomson (1981) mentioned such effects along the West Coast at Oregon and British Columbia coast.

Investigations off Oregon and Washington coast showed that the broad and slow southward flowing California Current is intensified into a narrow jet of stream with speeds exceeding 100 cm/s in the upwelling region.

Mathematical Models

In this section some mathematical models are described that deal with large-scale coastal upwelling due to longshore wind. In most models of large vertical motions in the ocean, Coriolis force is balanced against different forces such as pressure gradient force, friction force, etc. In the simplest of this model, known as Ekman-Sverdrup (Norwegian oceanographer Harald Ulrik Sverdrup, 1888–1957) model of upwelling, the Coriolis force is balanced against the steady wind forcing in homogeneous water. Together with the continuity equation, the solution of the system of equations (taking x-axis across and y-axis longshore) gives the Ekman transport (Pond and Pickard 1978; Smith 1983; Bowden 1983; Stewart 2004):

$$E_x = \rho_w w L = \rho_w u D = \frac{\tau_{sy}}{f}, \quad (1) \quad \text{and}$$

where E_x is offshore Ekman mass transport (kg/m/s), ρ_w is density of water (kg/m³), w is vertical velocity (m/s) of upwell-

ing, L is length (m) across the shore of upwelling influence, u is cross-shore horizontal velocity (m/s), D is Ekman layer depth (m), τ_{sy} is wind stress (kg/m/s²) on the water surface in the longshore direction, and f is Coriolis parameter (1/s) and is given by $2\Omega \sin \phi$, where Ω is angular speed of the earth (7.29×10^{-5} radians per second) and ϕ is latitude. The Ekman layer depth is a function of wind speed and latitude. At 30° North or South latitude, a wind speed of 10 m/s would cause $D = 61$ m. But for an Ekman layer to develop, a steady wind is required to blow for the duration of a pendulum day $= 4\pi/f$. For winds at the same latitude at 30°, the required duration is about 48 h. Some of the discussed wind-induced motions were applied by Barua and Kana (1995) in the Ganges-Brahmaputra Delta coast. The wind stress shown in Eq. 1 can be estimated using the quadratic friction law as

$$\tau_{sy} = C_d \rho_a U_{10y}^2, \quad (2)$$

where C_d is air friction drag coefficient (≈ 0.0015), ρ_a is air density (≈ 1.25 kg/m³), and U_{10y} is the wind velocity (m/s) at 10 m above mean sea level in the longshore direction.

The offshore Ekman transport defined in Eq. 1 is expressed in volumetric terms known as upwelling index, UI (Bakun 1973):

$$UI = \frac{E_x}{\rho_w}. \quad (3)$$

The UI is also referred to as Bakun index (m³/s per meter along the coast). The National Marine Fisheries Service of NOAA publishes monthly summaries of upwelling indices for major upwelling regions along the US coast.

Figure 2 shows an indication of UI magnitude responding to the change in longshore wind speed. The UI is estimated using Eqs. 1, 2, and 3.

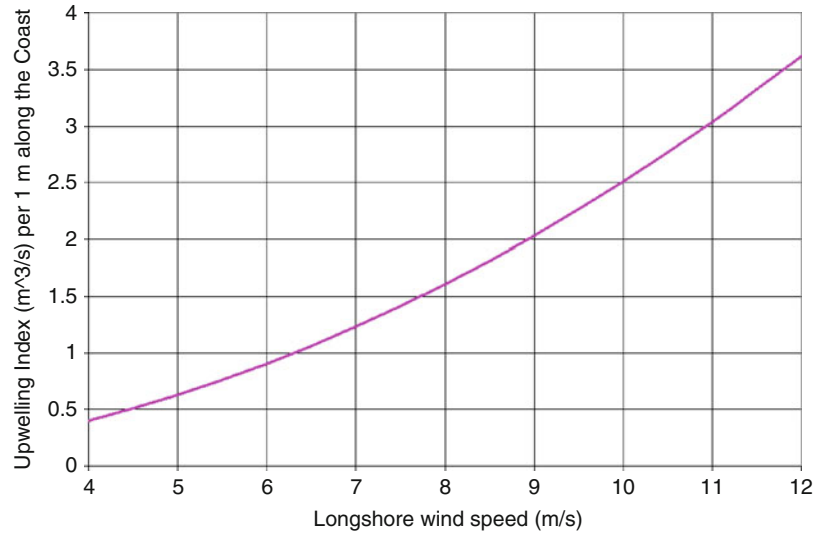
Hidaka (1954) obtained an analytical solution to the system of hydrodynamic equations that considered a balance between the Coriolis forces and the diffusive transports in the vertical and horizontal directions. From his findings, some scales of upwelling motion can be estimated:

$$S_v = \sqrt{\frac{N_v}{f}}, \quad (4)$$

$$S_h = \sqrt{\frac{N_h}{f}}, \quad (5)$$

$$\frac{w}{u} = \frac{S_v}{S_h} = \sqrt{\frac{N_v}{N_h}}. \quad (6)$$

Coastal Upwelling and Downwelling, Fig. 2 Upwelling index (UI) in response to the presence of persistent longshore wind



In Eqs. 4, 5, and 6, S_v is vertical length scale, S_h is horizontal length scale analogous to mixing length, N_v is vertical eddy diffusivity coefficient, and N_h is horizontal eddy diffusivity coefficient. Garvine (1971) made an analysis by including the longshore water slope and showed that beyond the narrow coastal boundary layer, frictional effects could be neglected.

In reality, ocean water is stratified, and a complete solution can only be obtained by taking into account all the forces. Some early contributions in this direction were given by Yoshida (1967), Allen (1973), and Hamilton and Rattray (1978). Their solutions are dependent on defining some characteristic baroclinic terms:

$$E_v = \frac{N_v}{fh^2}, \quad (7)$$

and

$$R_b = \frac{Nh}{f}. \quad (8)$$

where N is Brunt-Väisälä or buoyancy frequency and is given by

$$N = \sqrt{-\frac{g}{\rho_w} \frac{\partial \rho_w}{\partial z}}. \quad (9)$$

In these relations E_v is Ekman number in the vertical direction, R_b is baroclinic Rossby radius of deformation, z is vertical coordinate and is positive upward, h is water depth,

and g is acceleration due to gravity. Note that Ekman number is a ratio between the friction force and the Coriolis force and R_b defines a horizontal offshore scale in which upwelling occurs.

Measurements

Field measurements of surface temperature and conductivity in horizontal and vertical directions and anomalies of these parameters from the expected variation provide a quantitative indication of upwelling and downwelling. Some examples of the stable expected variation in the vertical direction (measured positive upward) are (1) a positive temperature gradient, (2) a negative salinity gradient, and (3) a negative density gradient. Any deviation from such stable gradients is an indication that upwelling or downwelling processes exist or are going to be initiated. Visible and infrared satellite remote sensing of color and temperature are also used to identify and show the dynamics of upwelled water (Chase 1981; Petrie et al. 1987).

Summary

This entry discusses coastal upwelling and downwelling phenomena that occur ubiquitously in many coastal waters. By defining the parameters, this entry describes the physics of the processes together with mathematical models. An estimate is presented on the upwelling index (UI) to indicate its magnitude in response to persistent longshore wind. The

geographical distribution, consequences, and measurement aspects of the phenomena are also highlighted.

Cross-References

- [Coastal Climate](#)
- [Coastal Currents](#)
- [Coastal Modeling](#)
- [Coastal Wind Effects](#)
- [El Niño–Southern Oscillation \(ENSO\)](#)
- [Meteorological Effects on Coasts](#)
- [Shelf Processes](#)
- [Time Series Modeling](#)
- [Vorticity](#)

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Coastal Warfare

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Definition

Coastal warfare involved military operations in the broadly defined littoral zone, including ground operations that involve proximity to water, naval operations that involve proximity to land, and air operations support both ground and naval operations.

Introduction

Most people live near the coast, and that concentration has prevailed far back into prehistory. As a result, military history includes many chapters on coastal warfare. Invading armies attack from the sea, and defenders build fortifications to keep them back. This recorded history starts with Homer's story of the Trojan War, and most recently includes the British recapture of the Falkland Islands, and in a new twist, the US Marines night landing in Somalia under the glare of television cameras.

Coastal warfare involves all the elements of warfare: knowledge and use of the environment (terrain, weather, and tides), fortifications and obstacles, deception and surprise, concentration of force at the decisive point, mobility, logistics, leadership, and technology. Coastal operations include naval battles, land battles in coastal plains, and sieges of coastal fortresses. Table 1 lists some important coastal battles.

Many naval battles have been fought in coastal waters, both because of strategic considerations and the need to support land operations. Two classical battles exemplify these operations. In 480 BC, the Greek fleet defeated the Persians at Salamis. The Greeks used constricted coastal waters to negate the size advantage of the Persian and Phoenician ships, and the naval victory ended the Persian invasion. In 31 BC, Octavius blockaded Marc Anthony's fleet at Actium on Greece's western coast. Anthony's fleet finally sortied with 220 heavy ships fighting 260 lighter ships. The

bloody but indecisive fighting ended when Cleopatra fled with 60 ships. When Anthony followed, his remaining ships despaired and Octavius won the battle that made him sole leader of Rome.

Coastal plains often provide key avenues of approach. In 490 BC, the Spartans and allies defended the pass at Thermopylae to stop the Persian advance along the coast. Betrayed when the Persians learned of a path around their positions, the Spartans conducted a heroic and doomed fight to buy time for the Greeks. As at Marathon earlier that year, coastal topography dictated the battle's location.

Coastal fortresses control land and sea avenues of approach. King Edward I of England created "Edwardian" castles in Wales. Five of these six classic castles had coastal locations allowing resupply and reinforcement by sea. Perhaps the greatest siege in history ended the millennium-long Byzantine empire, when the Ottoman Turks cutoff Constantinople by land and sea. Until the very end, the defenders hoped for relief from the sea. Improvements in technology have decreased the value of fortresses, and such sieges have become much less frequent, but the potential still exists for urban siege warfare in a coastal setting.

Coastal Warfare, Table 1 Selected coastal battles

490 BC	Marathon	Athens defeated landed Persian force on constricted coastal plain
490 BC	Thermopylae	Persians annihilate Spartans defending a choke point along the coastal plain
480 BC	Salamis	Greek naval forces defeat Persians in coastal water where the Persians cannot take advantage of larger ships
31 BC	Actium	March Antony's naval force defeated while trying to break out of blockade by forces of Octavius
1274, 1281	Mongol invasions of Japan	Two invasions defeated by typhoons
1453	Fall of Constantinople	Ottoman siege of Byzantine coastal capital, cutoff both by land and sea
1759	Quebec	British assault force landed at night, scaled Abraham Heights, and captured city
1915	Gallipoli	British land at multiple locations but cannot advance inland and eventually withdraw
1943	Tarawa	US forces capture small and heavily defended Pacific atoll
1944	D-Day invasion of Normandy	Allied forces unleash the largest amphibious assault ever on German defenders
1950	Inchon	UN forces broke out of the Pusan perimeter with amphibious landing far behind enemy line
1982	Falkland Islands	British recapture islands using naval and amphibious forces without any nearby land base

Amphibious Assaults

The amphibious assault in the face of a defended beach marks the culmination of coastal warfare. It requires cooperation among army, navy, and air force units, and reached its highest development during World War II. As late as World War I landings at Gallipoli used standard ships and small boats to land troops. In the early 1930s, the Japanese developed landing craft with a bow ramp. World War II saw the development and fielding of specialized landing craft and amphibious vehicles (Table 2). Four different theaters saw amphibious landings: leapfrogging from New Guinea to the Philippines (Fig. 1), island hopping in the Central Pacific, the Mediterranean theater movements from North African to Sicily and finally Italy, and the largest single amphibious operation in Normandy.

In 1950, UN forces in Korea staged a repeat of the World War II Pacific landings, with old landing ship tanks (LST) and many Marine veterans recalled to active duty. On environmental grounds Inchon presented an unlikely site. A narrow river channel and wide mudflats restricted ship operations. A 10 m tidal range restricted the landing to a few days each month. Surprise and impeccable execution overcame environmental drawbacks to make the Inchon landings a turning point in the Korean War.

Technology improvements have greatly increased the ability of large navies to successfully land forces on a hostile beach. Helicopters, and specially configured aircraft carriers to support them, allow airlifting of troops and bypassing of coastal defenses. Parachute and glider units filled this role in

Coastal Warfare, Table 2 Selected amphibian vehicles of World War II

Acronym	Name	Capabilities	Number produced
DUKW	Amphibious truck "Duck"	2.5 tons, 25 soldiers five knots at sea, or 80 km/h on land	20,000
LVT	Landing vehicle tracked-"amtracs," "amphtracs," or Alligator	8 m long, 6.5–7.5 knots; additional 2,850 armored LVT (A)	15,501
LCVP	Landing craft vehicle, personnel	3.6 tons of cargo or 36 soldiers	23,398
LCM	Landing craft, mechanized	15 m, one medium tank	11,392
LCI	Landing craft, infantry (large)	48 m, 200 soldiers two gangways instead of bow ramp	1,051
LCT	Landing craft, tank	Six medium tanks	1,435
LST	Landing ship, tank	100 m, 20 Sherman tanks; max. speed 11.5 knots; 1,490 tons; 26 lost in action and 13 in accidents	1,051
LSD	Landing ship dock	4,490–4,546 tons	18

Coastal Warfare, Fig. 1 LST (left) and LCVF (right) in the Philippines during World War II (Photo: M. Schwartz)



World War II landings like Normandy, but the helicopters can return to the ship and bring in additional troops and supplies ashore as needed. Squared-off bow ramps have been replaced with extendable ramps that allow a hydrodynamic bow and greatly increase the cruising speed of the amphibious ships while still allowing roll-off of assault vehicles. Large landing ship docks (LSDs, over three times the size of the 18 built during World War II) feature a large docking pool that can be flooded, allowing smaller amphibious vessels to rapidly float away to begin the assault. LCACs – landing craft air cushioned – use jet engines to float on a cushion of air, riding rapidly over the waves and obstacles up to 1.2 m high, and up onto the beach. Only one conflict since 1950 has tested the new technology, the British recapture of the Falkland Islands in 1982. When the Argentines attacked without warning, budgetary cuts left the British without a full complement of new amphibious equipment. Success relied as much on rapidly requisitioned and modified commercial vessels as on specialized naval vessels. In the First Persian Gulf War, the American amphibious forces had a diversionary role without actually performing a landing, and the new equipment and doctrine did not receive a battle test.

Military principles of amphibious assault mirror those of any offensive action. The attacker must mass forces at the critical location, and rapidly exploit the advantage. In many cases, this proves relatively easy to achieve, because the defender must spread forces over a large area and many beaches. The danger of mistaking a diversionary landing for the main attack always dulls defender reactions, allowing the attacker to exploit the initial numerical superiority on the landing beaches. Throughout history, many defenders have decided that beaches cannot be defended, and have kept large forces in reserve to fight a mobile defense once the situation becomes clear. German and Japanese attempts to fortify the beaches never stopped the attacker at the water's edge in World War II.

The key to the offense becomes rapid exploitation of the initial numerical advantage; failure here doomed the Gallipoli

landings in World War I, and led to stalemate after the Anzio landings during World War II. In contrast, the 1759 British capture of Quebec resulted from a nighttime landing, and an immediate scaling of cliffs the French had considered impassible. Rapid exploitation requires good planning, forceful leadership on the beach, and the ability to rapidly reinforce the beach. In almost all opposed landings, units in the first wave become militarily ineffective. Despite planning and rehearsals, boats land units at the wrong place and destroy unit integrity, and casualties decimate units. Junior leaders and individual initiative can carry the assault well inland and secure the beachhead, but the strategic goals of the invasion require that additional units be rapidly landed with unit integrity and supporting armor, artillery, and air defense support. Units in the follow-on waves must seize the initiative and expand the beachhead before the defenders can respond and mass their units.

Concurrently with the follow-on combat units, the attacker must land supplies to keep the units on the beach moving inland. At Gallipoli, the British had trouble keeping units supplied with water, and despite multiple successful landings at different locations and times they never exploited the initial success to expand the beachheads. At Inchon, the large tidal range required that LSTs be beached overnight to unload. Beached LSTs provide a tempting target, a necessary risk because assault troops need heavy vehicles and supplies. Logistics remain important until ports can be captured. Allied artificial harbors in Normandy failed when a major storm damaged the structures, increasing the time supplies had to come over the beach.

Environmental Considerations

Requirements for logistical support and combat reinforcement often dictate the time of a landing. A significant tidal range usually restricts landings to a few days each month, with two tides during daylight hours. The first tide near

sunrise brings in assault waves, and the next tide 12 h 30 min later brings supplies and a second wave. Depending on conditions, planners may prefer low or high tide. At Normandy, the assault came at low tide. This exposed obstacles in the surf zone, but combat engineers had to neutralize obstacles with the tide rising faster than obstacles clearance. At Inchon, planners chose high spring tide so assault vehicles could get to the seawall without getting mired in the mudflats. Neap tides might be selected to avoid the extremes in tidal range, and insure there is enough water to get over obstacles even at low water. Poor knowledge of the tides at Tarawa led to problems because the reef grounded many assault vessels, and there were not enough amphibious tractors to drive over the reef and up out of the water, but a strategic need for speed led planners to accept the risks. At Tarawa, the actual tide encountered did not correspond to either predictions or later hind casts, demonstrating the large effects of meteorological conditions with a tidal range of 1.2–1.9 m. Depending on location, favorable windows can be either two weeks (Normandy) or four weeks apart (Inchon). Tidal currents also influence the assault timetable, since at Inchon the maximum current speed equaled the speed of many amphibious assault vessels.

Geologic conditions for assault beaches generally require low slope and firm sediment. On a reflective–dissipative continuum, this generally places assault beaches in the middle of the spectrum. Assaults cannot target beaches backed by cliffs because there these lack easy egress, although landings at Quebec in 1759 and the American Rangers scaling cliffs at Normandy show that even this rule can be violated to achieve complete surprise or to provide supporting assaults. Moderate slopes offshore allow amphibious vehicles easy egress from the water, and moderate slopes ashore allow rapid expansion of the beachhead. The landing beaches at Gallipoli were small sandy pocket beaches along a coast that generally provided steep terrain right up to the water. Extreme low slopes generally must be avoided, because they come with small sediment sizes that will not allow heavy military vehicles to operate and rapidly exit the beach. The landing at Inchon came at high tide to avoid extensive mudflats; the selected “beaches” allowed the landing craft to get to the base of a seawall and minimized problems on mudflats. Favorable geologic conditions accompany favorable wave conditions, as low slopes correlate with smaller wave heights. Assaults also generally try to avoid offshore reefs. At Tarawa, the landing craft approached from inside the lagoon, but grounded on the reef and dropped Marines into the water where many drowned between the reef and shore. Other abandoned gear to get ashore, and emerged without weapons or supplies.

Weather always affects amphibious operations. Mongol invasions of Japan in 1274 and 1281 ended in defeat when intense cyclones destroyed much of the fleet. The Japanese called these *Kami-kaze*, or “divine winds,” and used the

term during World War II for suicide attacks, many at ships supporting amphibious landings. The D-Day landings had to wait 1 day, and nearly had to wait 2 weeks for favorable tides, because of a severe Atlantic storm. The Inchon fleet suffered minor damage from one typhoon while preparing in Japan, and put to sea a day early to avoid a second typhoon. For both D-Day and Inchon landings, correct favorable weather predictions by the staff meteorologist allowed the invasions to proceed without having to wait weeks for favorable tides.

While defenders have put obstacles in the water and along the beach to slow or stop the attackers, and dug reinforced positions for troops and weapons, they still defeated no major landings in World War II. Conversely, despite massive naval and air bombardments preceding most of the landings in World War II, attackers never dislodged and rarely seriously disrupted the defenders until assault boats hit the beaches. This paradox has led many military thinkers to assert that a mobile defense back from the beaches will work better than building the Maginot line at the shore, while for the attackers, a relatively short bombardment maintains surprise while achieving almost the same effect as a much longer bombardment.

Supporting landing operations can be among the costliest naval actions. During the Gilbert Islands campaign, the Marines lost 1,115 dead and missing on Tarawa. The sinking of the escort carrier *Liscome Bay*, supporting an army landing, led to the loss of 644 men. The largest naval battle ever fought, at Leyte Gulf, occurred when the Japanese attempted to destroy the forces landing in the Philippines. US losses included two escort carriers, including one lost to kamikaze. During the invasion of Okinawa in April 1945, kamikaze attacks sank 36 ships and landing craft. In the Falklands campaign, the British saw 9 of 23 frigates and destroyers either sunk or seriously damaged.

The SEAL (Sea, Air, and Land) Teams date only to 1970, but have become respected both in the military and Hollywood. Their primary missions involve military special operations along coastlines, but they often serve as general purpose special operations troops in non-coastal environments.

Twenty-First Century Developments

Coastal warfare in the twenty-first century has not involved conflict between nation states, but has seen three trends: combatting piracy, construction of artificial islands, and challenges in selecting appropriate technology. Recent piracy has involved three major hotspots: oil production areas near Nigeria, the choke points in sea routes through the Philippine and Indonesian archipelagoes, and the Somali region. Somali piracy, starting in 2005–2008 in coastal waters, peaked

in 2011 and brought significant ransom money to an impoverished country with no effective government. In response to a coordinated multi-national naval effort, pirates initially responded by going up to 2,400 km offshore in small, fast boats, before being brought under control in 2012. Periodic flare-ups, and the lack of improvement for the Somali state, suggest the continued threat of piracy, and a continued naval mission.

The South China Sea sits on one of the world's busiest shipping routes leading to the piracy hotspot in the Philippine and Indonesian archipelagoes, and has seen recent military activity including small scale armed clashes. China has been building a "Great Wall of Sand" of artificial islands, dredged to create land and building port facilities, air fields, and other military facilities in the Spratly Islands to press claims for fishing, possible oil and gas reserves, and as general territorial assertions. Other countries, with longer established territorial claims, have responded in cheaper fashions, such as the Philippines beaching the LST Sierra Madre on Second Thomas Shoal and sending a squad of marines to inhabit the rusting World War II ship that previously served in the US and South Vietnamese navies. This cat and mouse game among some of the most productive coral reefs in the world has environmental, legal, political, and military implications.

Piracy and monitoring territorial claims complicate the job of navies, which must prioritize procurements to deal with competing missions. Recognizing the importance of the coast environment, in the late 1990s the US Navy began developing the Littoral Combat Ship (LCS), which most navies would classify as a corvette. The design featured easy to reconfigure modules for different roles, including anti-submarine warfare, mine countermeasures, anti-surface warfare against small boats, but also had to handle intelligence, surveillance and reconnaissance, homeland defense, maritime intercept against piracy, special operations, and logistics missions. The LCS suffered from mission creep, cost overruns, reliability issues, and questions about survivability in combat, leading to reductions in the number of ships to be acquired. In 2015, the Navy decided to change future procurement to a frigate class as it continues to balance coastal warfare requirements with deep water missions requiring larger ships.

Conclusion

Even as active coastal warfare operations have become rare, military forces must acquire specialized hardware and train to fight in an increasingly complex environment. Population migration to the coastal regions continues, as the risks from rising sea level and global climate change intensify, makes the coastal environment critical to militaries around the world.

Cross-References

- [Coastal Climate](#)
- [Coral Reefs](#)
- [Human Impact on Coasts](#)
- [Pacific Ocean Islands, Coastal Geomorphology](#)
- [Scour and Burial of Objects in Shallow Water](#)
- [Tidal Environments](#)
- [Tides](#)
- [Wave Environments](#)

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Coastal Wells

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And he looked, and saw a well in the field, and behold, three flocks of sheep were lying there beside it, for from the well they watered the flocks. Now the stone on the mouth of the well was large.

Genesis 29, 2

When all the flocks were gathered there, they would then roll the stone from the mouth of the well...

Genesis 29, 3

Historical Background

Water in arid and semiarid land is one of, if not the most important and vital factor, for the existence of people,

livestock, and farming all along the history of mankind. Therefore, in many regions where freshwater was not available, occupation was very difficult and much sporadic if at all. In general, people were dependent on permanent water sources such as springs or rivers that supplied water all year round, and temporary sources such as seasonal rivers, natural water cisterns, etc. Later, when agricultural and farming irrigation systems were developed, an immense need for larger water quantities existed. Throughout history, when there has been a necessity and demand for expansion of settlements, either as a result of political pressure or due to the growth in population, people moved (to their sorrow), from the prolific and rich water regions into much poorer areas.

Coastal areas in general, and more specifically the near-littoral zone, reached an ever-growing importance when the sea became another nutrient source to the communities of the gatherers and hunters. As a result, they looked for new water resources, preferably natural, and if these did not exist, they searched and created artificial sources.

In arid and semiarid regions where natural water sources are very scarce, settling was totally dependent on their talent to either transport water to these regions, or locate and discover other sources, mostly underground. The existence of ground-water at a relatively shallow depth below the surface in the nearshore region permitted water exploitation through well digging at least as early as the Pre-Pottery Neolithic Period (ca. 7900 years BP; Galili and Nir 1993). This period parallels the beginning of the development of canal irrigation systems in China and Anatolia (Hübener 1999).

There is no doubt that during the early periods of humanity, people experienced seasonal living along the coast

(Oren 1992), in regions where mountains reach the coastal zone, in most cases where there had been no water problem. Springs or river water easily covered the daily needs of people and their livestock. While in shallow relief or hilly coastal plains, in those semiarid lands where no springs or rivers exist, the water problem was very crucial to their survival.

Coastal Water Wells

At a certain stage of development, people discovered the fact that fresh or semi-fresh water may be reached very easily by digging shallow holes on a sandy beach quite close to the coastline (perhaps in imitation of animal behavior). It is assumed, although there is very little chance of finding any evidence at all, that this method was practiced for a very long period at much earlier stages before digging of water wells commenced.

The oldest known well in Israel is found at 10–12 m. water depth in an Atlit Pre-Ceramic Neolithic village, located offshore, opposite the southern flank of Mount Carmel (Galili et al. 1988; Galili and Nir 1993; Nir 1997). This well is about 5 m deep, mostly built of stones. It was found in a relatively large prehistoric site with many dwellings and other village installations. The well, dug in a clayey soil reached, at its lower part, a permeable carbonate cemented quartz-sandstone layer, the Pleistocene “kurkar” rock, which is known to contain a very prolific high-quality aquifer.

The majority of the historical wells are dug as a vertical pit, having a cylindrical shape, with inner diameters varying from

Coastal Wells, Fig. 1 A typical Anti I Man coastal well at Tell Ashqelon, located on the southern coast of Israel (Photo: Y. Nir)



80 to 200 cm (Forbes 1955; Nir and Eldar-Nir 1987, 1988). Wells of this type are technically constructed from top to bottom, layer after layer, strengthening its.

walls with local stones, undressed in the old wells, but nicely dressed and fit with the well's round shape in the more "modern" wells from the Hellenistic and Roman periods and on to the present.

The water sources of the coastal wells depend on the ground water table, water flowing from inland higher topography to the drainage basin—the sea. This water table fluctuates with changes in sea level that fully controls its level.

In the coastal plain of Israel, and no doubt in similar geologic terrain elsewhere around the Mediterranean and other coastal regions, ground-water has an average slope of about 1/1000, and its surface is almost linear (Kafri and Arad 1978). Many sub-recent wells were dug along the coastal plain of Israel, most of which are in the nearshore strip. One of the most famous sites, where a high concentration of wells is found, is at Tell Ashqelon, located on the southern coast (Nir 1997). More than 50 water wells of ages from Roman to the twentieth century were mapped there (Fig. 1).

Water wells of the dug-pit method have been in operation up to the present time, mainly in underdeveloped regions. Wells of different shapes, construction materials, depth, etc., exist along most of the world coasts, located in all regions where the ground water table is close to the surface. Their importance in the developed countries, has dropped, while in the developing countries where there are no other natural sources they are still of high importance to the coastal population.

Cross-References

- [Archaeology and Sea-Level Change](#)
- [Archaeology](#)
- [Changing Sea Levels](#)
- [Desert Coasts](#)
- [Hydrology of the Coastal Zone](#)

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Coastal Wind Effects

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Wind can have several effects on the processes influencing coastal geomorphology. These include wind stress on the water surface in major storms, such as hurricanes or typhoons, inducing short-term above normal sea elevations or storm surge, short "choppy" waves in estuaries and fetch-limited harbors, downwelling and upwelling processes in the coastal oceans, and sea breeze effects.

Storm surge is caused by the frictional wind stress on the water surface inducing a current in the surface water. Empirically, the resulting current in the top few centimeters of the sea surface will be about 2% of the wind velocity (Komar 1976). For Class 5 storm (hurricane) winds, a considerable mass transport is advected along the sea surface waters in the wind direction, so that onshore storm winds force the seawater up against the land. Of course this occurs concomitantly with large, wind-forced, storm waves. The total wind effect is therefore a combination of both the direct wind stress effect and the indirect wave run-up effect. Storm wind stress effects are maximized over broad shallow continental shelves and semiencloded seas, such as the Gulf of Mexico, the North Sea, and the Baltic Sea, where extreme surge effects can raise the still water level by ~5 m and flood over the low-lying barriers, estuarine wetlands, and coastal land. The storm surge effect enhances the beach erosion process along open sandy duned coastlines by raising the water table in the beach and lowering intergranular friction between the sand grains, so that the pore water pressure from the raised water table induces the sand from the lower beach face to flow out into the surf zone. But wind-generated surge also piles up

Terry R. Healy: deceased.

tidal waters inside estuaries, tidal inlets, and river mouths, often leading to saline flooding of adjacent low-lying coastal land. As a result of the onshore winds and storm surge, the tidal cycles within the estuaries are altered to a much higher than normal tides and reduced low tide.

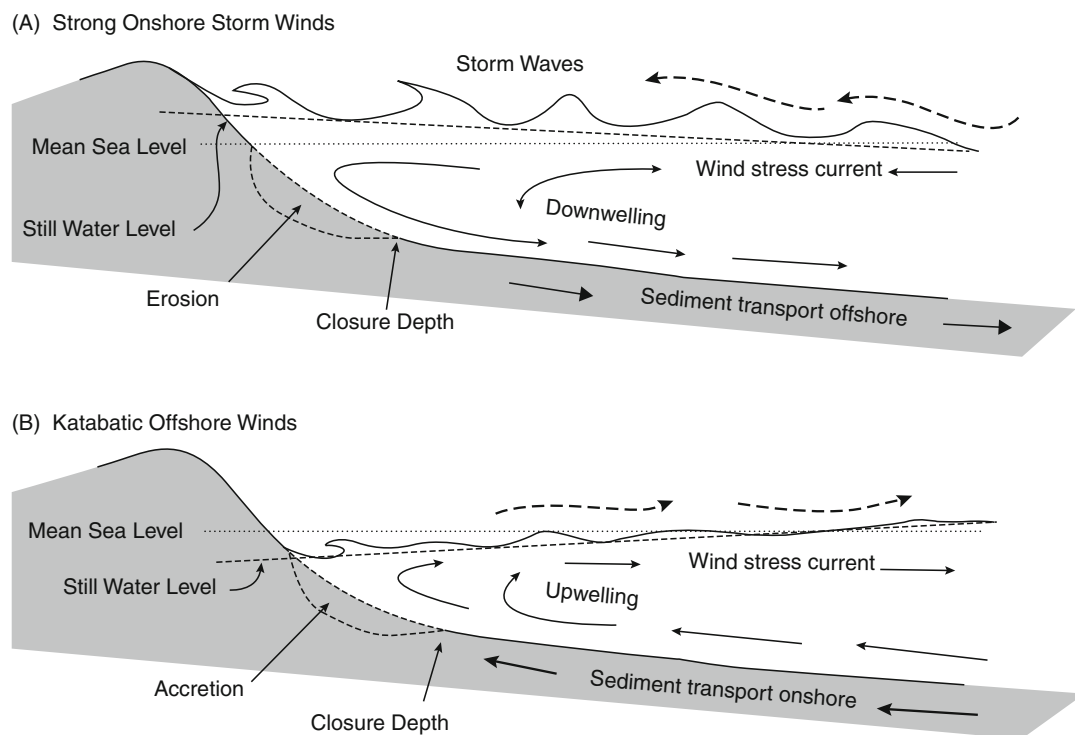
Short steep, “choppy” waves are generated by fetch-limited winds inside estuaries and semi-enclosed seas. These “choppy” waves have a number of effects including stirring of the bottom muddy sediments in the shallow near-shore creating the characteristic high muddy suspended load “turbid fringe” observed around shallow estuaries on windy days. These waves may even generate “fluid” mud in muddy estuarine environments from the constant bottom agitation of the loosely cohesive bottom mud sediments. Wind-forced choppy waves in enclosed bays, on top of a storm surge, result in erosion cutback of low terraces and marshland above the normal high tide levels, and the lagoonal shore of barrier islands and spits—this is the prime mechanism for shore erosion within sheltered harbors and estuarine lagoons.

Downwelling and upwelling in the coastal zone are largely wind-generated processes (Fig. 1). These processes have marked, but often overlooked, influence on the diabathic (cross-shore) sediment transport. For the case of significant onshore winds blowing for some hours, surface waters are forced up against the beach and dunes, raising the still water level at the beach. As a requirement of continuity, the onshore surface current in turn generates a bottom return current (downwelling). Thus sandy sediment grains initially eroded off the

beach in a storm, when uplifted by the bottom orbital motion of the waves, are acted upon by the downwelling current as they fall back to the seafloor, and thus become transported offshore, typically to “closure depth.” Conversely, with consistent significant offshore winds, wind stress on the sea surface forces the surface water out to sea, and lowers the still water level. This generates an onshore-directed bottom return current (upwelling), and as the longer period (non-storm) swell waves uplift the bottom sediment, it is transported onshore to the beach, facilitating beach face accretion, and rapid rebuilding of the beach berm. Of the two processes, the latter likely transports bottom sandy sediment at a greater rate because the longer period waves have greater bottom orbital

velocities which uplift more sediment. It is often observed that beach recovery at the end of a storm (when winds reduce in strength and change to an offshore direction) is more rapid than the loss of sand from the beach during the erosion phase.

Sea breeze effects arise from the thermal heating of the land generating a steady onshore wind. The effect of sea breeze on beach morpho-dynamics has been studied on the Western Australian coast, where sea breezes are some of the strongest in the world. Here the sea breeze cycle is an important influence on beach processes, with onshore winds reaching magnitudes >8 m/s. Masselink and Pattiaratchi (1998) researched the effects of the sea breeze on the near-shore processes, demonstrating how with the onset of a strong onshore sea breeze of 10 m/s, the root mean square (rms) wave height increased from 0.3 to 0.5 m, the wave period



Coastal Wind Effects, Fig. 1 Effect of wind generation inducing coastal upwelling and downwelling and associated diabathic sediment transport

decreased from 8 to 4 s, and mean cross-shore flows reached velocities of 0.2 m/s directed offshore. The short period sea breeze-generated waves became superimposed on the background swell. During the sea breeze event, nearshore sediment resuspension became almost continuous, and cross-shore transport was directed offshore. But in addition, the sea breeze affected the beach longshore transport rates when it approached the coast obliquely, and the longshore-suspended sediment transport rate increased by a factor of 100. These sea breeze-induced hydrodynamic changes were associated with erosion of the beach face, and deposition in the surf zone, with a consequent flattening of the profile. The effect was similar to a storm of medium intensity.

Cross-References

- [Beach Erosion](#)
- [Coastal Upwelling and Downwelling](#)
- [Cross-Shore Sediment Transport](#)
- [Storm Surge](#)
- [Tides](#)

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Coastline Changes

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Definition

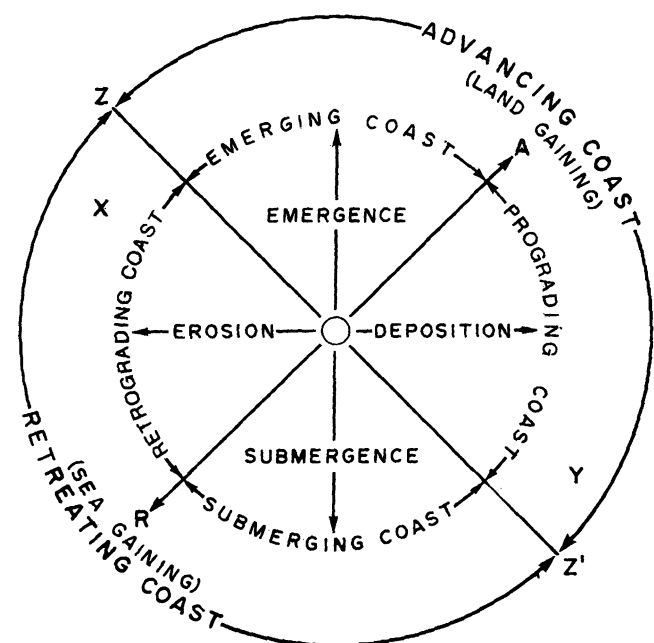
Coastline changes (here defined as changes on the high-tide shoreline) can be mapped and measured on various coastal sectors over time scales ranging from a few hours or days (related to tidal cycles, weather events, earthquakes, or tsunamis) to long-term trends over decades or centuries.

Evidence of Coastline Change

Evidence of coastline changes can be obtained from comparisons of dated historical maps, charts, air photographs, and satellite imagery with the present coastal configuration. Such

evidence is available on many coasts over the past century, but in some countries, such as Britain and Denmark, maps and charts surveyed around the beginning of the nineteenth century permit the determination of changes over longer periods: Houston and Dean (2014) used measurements of shoreline change on the east coast of Florida since the mid-1800s to demonstrate long-term beach accretion. In the Mediterranean region, and locally elsewhere, there is sporadic evidence of coastline changes over periods of at least 2,000 years, using historical descriptions and dated archaeological sites. On some coasts there is geological evidence of changes over the past 6,000 years, since the Late Quaternary marine transgression brought the world's oceans to approximately their present levels: cliffed coasts have receded and depositional coasts have advanced, and some coasts show evidence of alternations of advance and retreat during this period.

Advance or retreat of a coastline may result from combinations of erosion and deposition, emergence and submergence (Fig. 1). Coastline changes are usually expressed in linear terms, as the extent of advance or retreat measured at right angles to the land margin, but they can also be considered in areal terms, as the extent of land gained or lost on a coastal sector, or volumetric terms, as the quantity of material added to, or lost from, the coast. Measurements made over varying periods are commonly averaged as gains or losses per year, but coastline changes may be sudden or intermittent rather than steady and gradual. Attempts have been made to



Coastline Changes, Fig. 1 Analysis of coastline changes, based on Valentin (1952), Z-Z' indicates a stable coastline where erosion has been offset by emergence or deposition by submergence, or where no changes have taken place (O) (Copyright-Geostudies)

document the advance or recession of coastlines in several countries: the Royal Commission on Coast Erosion in Britain (1907–11), the National “Shoreline” Study in the United States (US Army Corps of Engineers, 1971), and the survey of the Netherlands coastline by Edelman (1977); and in 1984 the International Geographical Union’s Commission on the Coastal Environment completed a global review of coastline changes over the preceding century (Bird 1985). More recently there has been monitoring of changes on particular coastlines using remote sensing and photogrammetry, but this has not yet been pursued systematically around the world’s coastline. In the absence of another global review the indications are that changes during the past 30 years have been mainly on the pattern identified previously, apart from large-scale coastal land reclamation projects, particularly in Asia.

Some coastlines have changed very rapidly (advance or retreat >100 m/year), but changes of more than 10 m/year are unusual, and few have changed by more than ± 1 m/year. If, as is predicted, a worldwide sea-level rise develops and accelerates as a consequence of global climatic warming resulting from the enhanced greenhouse effect, submergence and erosion will become widespread, and coastline changes are likely to intensify (Bird 1993). Coastline changes can be considered in the following categories.

Cliffed Coastlines

Cliffs in hard, massive rocks have changed little over the past century, even where they are exposed to stormy seas, but cliffs in soft materials can be cut back very rapidly. Sample rates of cliff retreat were tabulated by Sunamura (1992: Appendix 2). Recession of up to a meter a day has been measured on cliffs cut in unconsolidated volcanic ash, as on the Isla San Benedicto, off the Pacific coast of Mexico, and cliffs cut in

alluvium and peat weakened by seasonal freeze-thaw oscillations have retreated up to 100 m in a few summer months before a winter cover of snow and ice briefly stabilized them. On parts of the Polish coast, cliffs cut in glacial drift deposits have shown an average recession of a meter a year, but have lost up to 5 m in a single storm, and similar rates have been documented on the east coast of England, notably on the Holderness cliffs, also cut in glacial drift. Recession of cliffs cut in soft formations is often accelerated by slumping and erosion by runoff after heavy rain or from melting snow. Storms have modified structures such as stacks and natural arches on cliffed coasts, as on the Port Campbell coast in Victoria, Australia, where several of the stacks known as the Twelve Apostles have disappeared in recent years and natural arches collapsed at London Bridge in 1990 (Figs. 2 and 3) and Island Archway in 2009.

Glaciated Coastlines

Where glaciers and ice sheets reach the sea, as in Icy Bay, Alaska, southernmost Chile, and sectors around the Antarctic continent, the coastline may advance with forward movement of the ice, or retreat as the result of melting. Ice cliff coastlines have receded at rates of up to a kilometer per year in Antarctica as the result of melting and calving induced by recent climatic warming. In the Gulf of Alaska, however, rivers augmented by glacialfluvial outwash from melting ice have advanced parts of the coastline by more than a kilometer in recent decades by building deltas and prograding beach ridge plains. Rapid progradation occurred on the southeast coast of Iceland when melting of ice, hastened by volcanic eruptions, generated torrential outwash, delivering large quantities of sand and gravel to prograde beaches and build barriers and spits.

Coastline Changes, Fig. 2

London Bridge, Port Campbell coast Victoria as it was in 1989 (Copyright-Geostudies)



Coastline Changes, Fig. 3

London Bridge, Port Campbell
coast Victoria, in 1990
(Copyright-Geostudies)

**Emerging and Submerging Coastlines**

Where land uplift has occurred the coastline has advanced. Isostatic uplift of formerly glaciated areas in Scandinavia and northern Canada has led to the emergence of former nearshore sea floors, as in the Gulf of Bothnia, where the coastlines of Sweden and Finland have advanced more than 100 m over the past century. On some sectors progradation has been augmented by shoreward drifting of sea floor sediment, notably sand from submerged eskers, to build up beaches. Sudden emergence during earthquakes has advanced coastlines on Montague Island, Alaska (1964), Talcahuano in Chile (observed by Charles Darwin during the voyage of the *Beagle* in 1835), the Orontes delta in Turkey (1958), the Makran coast in Pakistan (1945), Cheduba Island in Burma (1762), Tokyo Bay in Japan (1923), and Wellington (1855) and Napier (1931) in New Zealand. Emergence occurred during the phase of sea-level lowering between 1929 and 1977 on the coasts of the Caspian Sea, when islands expanded and became attached to the mainland, while shoals emerged as islands.

Submergence results in coastline recession, except on vertical cliffs or where it is compensated by coastal deposition. Land subsidence in the vicinity of the Mississippi has led to the submergence and erosion of former subdelta coastlines, and land subsidence in the Netherlands, augmenting the effects of a rising sea level, would have submerged a large part of the country had the coastline not been maintained by seawalls and reclamation works. Subsidence in the Venice region, due partly to subsurface compaction and the extraction of groundwater, has accelerated beach retreat on the Lido barriers and diminished the extent of salt marshes in the Venice Lagoon. Increasing depth and frequency of storm surge flooding here indicates that submergence is partly due to a rising sea level in the Adriatic. Since 1977 the Caspian Sea has been rising and the coastline receding, much of the

coastal land that emerged previously having been inundated. Sandy barriers have formed and are migrating landward (Ignatov et al. 1993).

Sudden submergence during earthquakes led to coastline recession at Anchorage in Alaska (1964), in southern Colombia (1979), at Concepcion in Chile (1960), and on the Indus delta (1819, 1845). Tsunamis following sea floor mass movements have caused sudden recession of the Valdez delta in Alaska and along the beach-fringed Ivory Coast.

A correlation has been noted between alternations of water level in the Great Lakes and the advance or retreat of bordering coastlines. During phases of falling lake level the coastline advances as the result of emergence and shoreward drifting of sand to accreting beaches, while during phases of rising lake level the coastline retreats as the result of submergence, beach erosion, and increased cliff recession.

Volcanic Coastlines

Volcanic eruptions prograde the coastlines where lava and ash are deposited, as on the southeast coast of Hawaii. They also augment river sediment loads, leading to the advance of deltaic coasts, as on the Rio Samala in Guatemala, and the progradation of beaches, as in the Bay of Plenty, New Zealand, after the Tarawera eruption in 1886 and on the south coast of Java after recurrent eruptions of the Merapi volcano.

Some oceanic islands are entirely of volcanic origin. Surtsey, off the south coast of Iceland, was built by volcanic eruptions in 1966–1967, and has since been reshaped by the rapid recession of cliffs cut in lava and ash on the more exposed southern coastline, generating sand and gravel that have drifted round to accumulate on a triangular northern foreland. The explosive eruption of Krakatau Island,

Indonesia, in 1883 left caldera-side cliffs and a surrounding mantle of ash which initially advanced the coastline, but has since been cut back as cliffs receding at up to 6.6 m/year.

Mass Movements

Coastal landslides and rock falls temporarily prograde the coastline, forming lobes and fans that are subsequently cliffed and cut back by marine erosion. This sequence has been documented at Tillamook Head in Oregon; Pacific Palisades, California; Black Ven and Fairy Dell in Lyme Bay, Dorset; Rosnaes and Helgenaes in Denmark; and on the Bulgarian and Crimean coasts of the Black Sea. On the Dorset coast erosion pin measurements have shown intermittent mudslide movements totaling 8 cm in 30 h, graded movements of about 3.5 m in 17 h and surge movements of up to 3 m in 20 min (Allison and Brunsden 1990). Some landslide lobes have persisted long enough to act as natural breakwaters, intercepting drifting beach sediment and causing local progradation, as at Golden Cap in Dorset.

Deltaic Coastlines

The arrival of large quantities of fluvial sediment at the mouth of a river has led to rapid accretion of sectors of deltaic shoreline, with rates of up to 80 m/year at the south-west Pass of the Mississippi delta and up to 70 m/year on the Irrawaddy delta. On the north coast of Java deltas fed by sediment-laden floodwaters have prograded by up to 180 m/year. Progradation results from deposition of fluvial sediment on river banks to form natural levees, prolonged seaward as sedimentary jetties, and the delivery of sand and gravel to prograde bordering beaches. In addition, there is often a seaward advance of bordering salt marshes, as on the Fraser delta in Canada, or mangrove swamps, as on most tropical deltas, and sedimentation within these eventually builds up new land at high tide level.

Even where deltas have not formed, sediment supplied by rivers can prograde adjacent coastlines. Sand from the Columbia River on the Pacific coast of the United States has been distributed northward and southward to prograde Long Beach and the Clatsop Plains. Fluvially nourished beach progradation has also occurred on the west coast of South Island, New Zealand, and in north-east Queensland.

Where the mouth of a river is diverted during a major flood, the coastline that had formerly prograded begins to recede, while new progradation develops at the diverted outlet. This has happened on the Huang He in China (1852), the Rio Sinu in Colombia (1942), the Ceyhan in Turkey (1935), the Medjerda in Tunisia (1973), and the Cimanuk in Java (1947). Artificial diversion of a river mouth by canal cutting

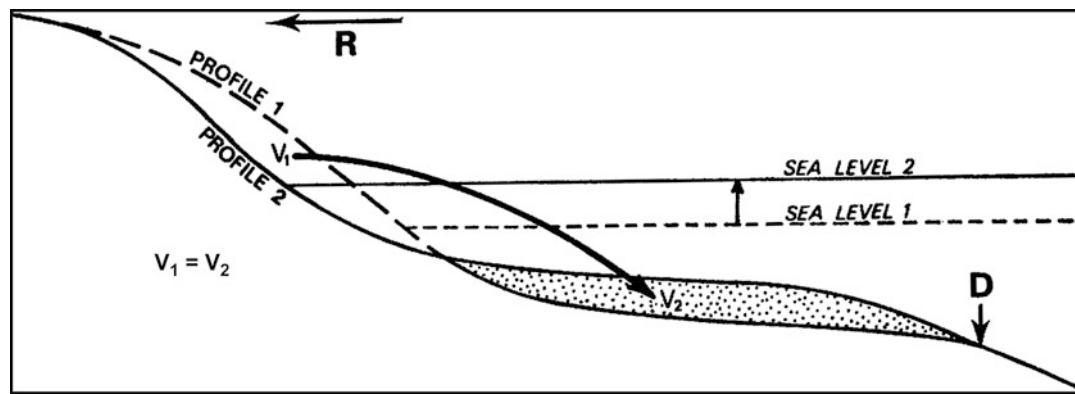
has had similar effects on the Brazos delta in Texas (1938), the Rioni in the eastern Black Sea (1939), the Shinano in Japan (1922), and the Cidurian in Java (1927).

Progradation of deltas and coastal plains has accelerated where river sediment yields have increased as the result of soil erosion in their catchments, due to deforestation, overgrazing, or cultivation of steep hinterlands. In Java this has led to rapid progradation of deltas in Jakarta Bay, and the Segara Anakan bay on the south coast is shrinking rapidly as the result of increased sediment input from the Citanduy river. On some coasts, progradation has followed an increase in fluvial sediment yield due to hinterland mining. Hydraulic sluicing in the Sierra Nevada a century ago led to the growth of the Sacramento delta into San Francisco Bay, and open-cast mining has resulted in beach progradation at Par and Pentewan in Cornwall, Chanaral in Chile, Jaba on Bougainville, and around New Caledonia, especially on the Thio and Népoui deltas. Reduction in fluvial sediment yield to the coast following the building of dams on rivers has led to the onset or acceleration of erosion on many deltas. The best-known example is the Nile delta coastline, which has shown recession of up to 40 m/year since the completion of the Aswan High Dam upstream in 1964. Similar erosion has been documented on the Rhône delta in France. Deltaic coastline erosion has also followed extraction of sand and gravel from river channels or the reduction of runoff by irrigation schemes, soil conservation works, and afforestation in catchments, notably in Italy, Greece, and Turkey.

Beach-Fringed Coastlines

Changes on beach-fringed coastlines result from variations in the rate of supply of sediment from various sources or in losses from the beach system, as shown in Fig. 4. Progradation of beaches and growth of spits supplied with sand and gravel from the erosion of nearby cliffs is well illustrated on coasts dominated by glacial drift deposits, as on Puget Sound on the west coast of North America, New England and the Canadian Maritime Provinces on the east coast of North America, and the coasts of the North Sea and the southern Baltic. Changes in the configuration of spits generally include their prolongation, often accompanied by recession of their outer coastlines with longshore drift of eroded sediment to their growing ends: an example is Cape Henlopen, a northward-growing spit on the southern side of Delaware Bay, on the Atlantic coast of the United States.

Construction of breakwaters to protect ports and harbors has often resulted in changes on adjacent coastlines, notably where longshore drift of beach sediment has been intercepted to cause local progradation. In some cases accretion updrift of such breakwaters has led to erosion downdrift, as at Lagos Harbor in Nigeria, but there are also examples of progradation



Coastline Changes, Fig. 4 The various ways in which sand is gained by, or lost from, a beach. When gains have exceeded losses beach progradation is likely to occur, while losses exceeding gains leads to

the lowering, flattening, and cutting back of the beach profile. Gravel (shingle) beaches show similar patterns of gain and loss, without the wind-blown component (Copyright-Geostudies)

on both sides of a harbor entrance resulting from longshore drift convergence or shoreward sediment supply, as at Lakes Entrance in Victoria, Australia.

Beach erosion can occur when the sediment supply from eroding cliffs diminishes as the result of seawall construction or stabilization, or a reduction in incident wave energy, as at Bournemouth and Brighton in southern England, the Melbourne Bayside coast on Port Phillip Bay, and Byobugaura in Japan.

On many coasts the onset of beach erosion has been due primarily to a rise in relative sea level, but as shown in Table 1 there may be other contributory causes. A recent increase in beach erosion around Port Phillip Bay, Australia, is thought to be the result of deepening the entrance from the sea by dredging to improve navigation, leading to a rise in high tide levels (Bird 2011). Beach erosion has been widespread around the world: examples of coastline advance by continuing beach accretion in recent decades are comparatively rare, many beach resorts having erosion problems, but such an advance has occurred at Seaside in Oregon, Malindi in Kenya, and Riumar on the Ebro delta in Spain. Successful beach renourishment projects have led to the advance of the coastline, as at Praia da Rocha in southern Portugal (Fig. 5).

The spilling of waste from coastal quarries has prograded beaches, as at Hoed in Denmark, Porthoustock in Cornwall, and Rapid Bay in South Australia. At Workington in Cumbria, the dumping of waste from coal mines and steel works has locally augmented beaches. On the other hand, quarrying of sand, gravel, or shelly material from beaches has not only depleted them, but has led to accelerated erosion on nearby cliffs, as at West Bay and Seatown in Dorset, Gunwalloe in Cornwall, Hallsands in Devon, and St. Ouen's Bay in Jersey.

Information supplied to the International Geographical Union's Commission on the Coastal Environment in 1972–1984 indicated that about 20% of the world's coastline consisted of sandy beaches backed by beach ridges, dunes, or other low-lying sandy terrain. Of this, more than 70% had

shown net erosion over the preceding few decades and less than 10% sustained progradation, the remaining 20–30% having been stable, or shown no measurable change. The initiation or acceleration of beach erosion was attributed to several factors, listed in Table 1. The relative importance of these factors differs from one coastal sector to another, but analysis of particular eroding beaches has shown that usually more than one has contributed. A detailed discussion was provided in Bird (1996).

Swampy Coasts

The seaward margins of salt marshes and mangrove swamps (generally close to the mid-tide shoreline) have advanced seaward where the sediment supply has been abundant, particularly on coasts sheltered from strong wave action and on the shores of prograding deltas. Sedimentation within the widening salt marshes and swamps eventually builds them up to high-tide level, so that the coastline advances. Extensive salt marsh progradation has occurred around Chesapeake Bay during the past 350 years, and has been attributed to increased sediment yields from inflowing rivers as the result of soil erosion in their catchments. Mangrove swamps have prograded on some tropical coasts, particularly near growing deltas, as on the shores of southern and eastern Kalimantan in Indonesia. On the west coast of Peninsular Malaysia, mangroves spread seaward by up to 54 m/year between 1914 and 1929 when sediment yields from agriculture and mining were high, but after dams were built on rivers and tin mining activity reduced erosion became prevalent (Kamaludin 1993). The major tsunami in the Indian Ocean in 2004, generated by an earthquake off Sumatra, caused extensive marine flooding and erosion on the coast from Indonesia and Malaysia through Thailand and India around to Somalia: the effects were diminished where a mangrove fringe persisted, but were severe where mangroves had been cleared.

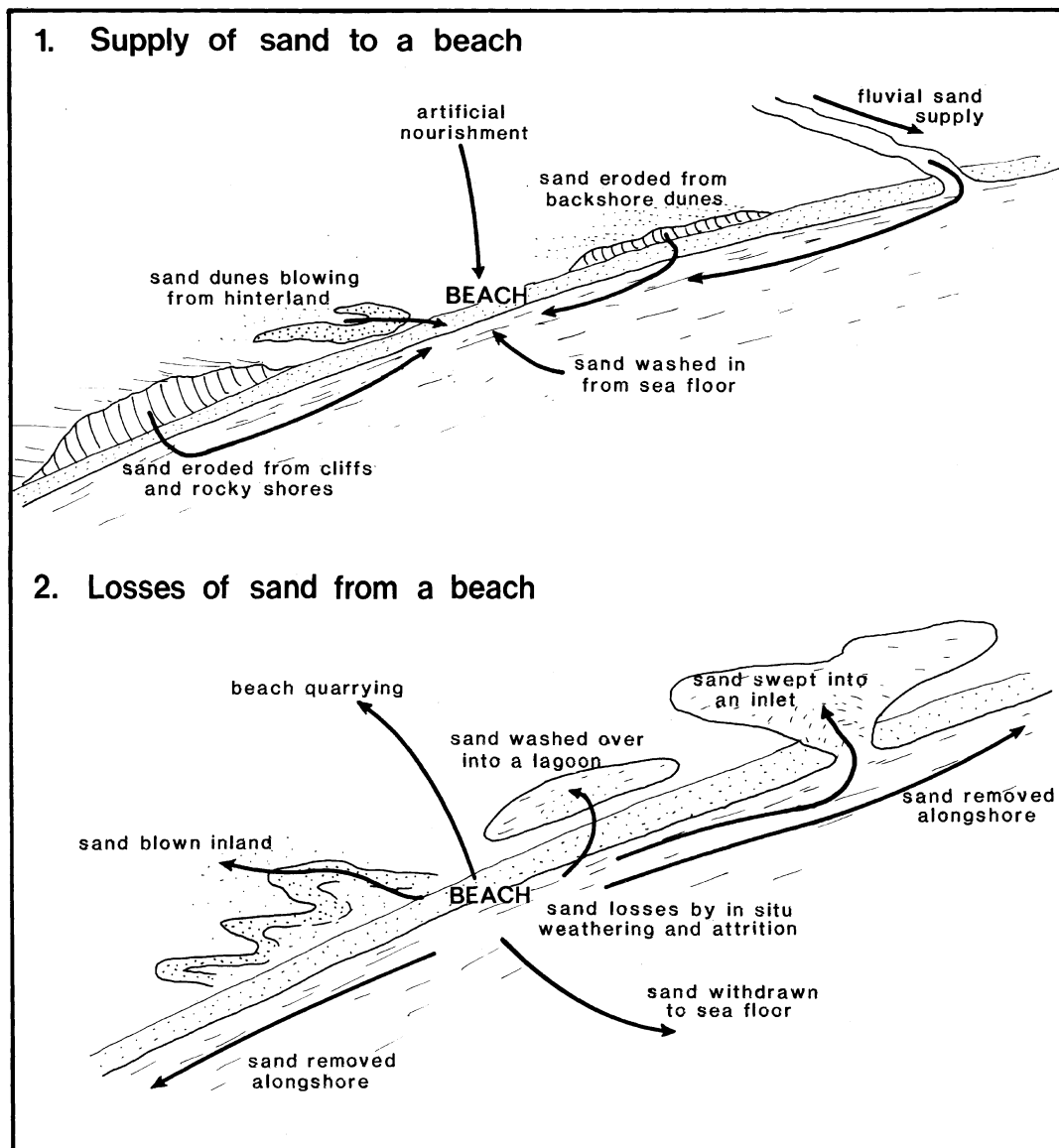
Coastline Changes, Table 1 The causes of beach erosion

1. Submergence and increased wave attack as the result of a rise in sea level or coastal or nearshore land subsidence: the “Bruun Rule” (Fig. 5)
2. Diminution of fluvial sand and shingle supply to the coast as a result of reduced runoff or sediment yield from a river catchment (e.g., because of a lower rainfall, or dam construction leading to sand entrapment in reservoirs, or successful revegetation and soil conservation works), or the natural or artificial diversion of a river outlet, or because of the removal of much or all of the weathered mantle from slopes in the river catchments, exposing bare rock that yields little or no sediment to runoff
3. Reduction in sand and shingle supply eroded from cliffs or shore outcrops (e.g., because of diminished runoff, stabilization of landslides, a decline in the strength and frequency of wave attack, or the building of seawalls to halt cliff recession), or the exposure by cliff recession of formations (such as massive resistant rock, or soft silts and clays) that do not yield beach-forming sediment
4. Reduction of sand supply to the shore where dunes that had been moving from inland are stabilized, either by natural vegetation colonization or by conservation works, or where the sand supply from this source has run out
5. Diminution of sand and shingle supply washed in by waves and currents from the adjacent seafloor, either because the supply has declined (e.g., where ecological changes have reduced the production of shelly or other biogenic material), because the transverse profile has attained a concave or steeply declining form which no longer permits shoreward drifting, or because such drifting has been impeded by increased growth of seagrasses or other marine vegetation
6. Removal of sand and shingle from the beach by quarrying or the extraction of mineral deposits, or losses from intensively used recreational beaches, for example, as the result of beach cleaning operations
7. Increased wave energy reaching the shore because of deepening of nearshore water (e.g., where a bar or shoal has drifted away, where seagrass vegetation has disappeared, where dredging has taken place, or where the seafloor has subsided)
8. Reduction in sand and shingle supply from alongshore sources as the result of interception (e.g., by a constructed breakwater or groins) or interruption by the growth of a fringing coral reef or some other depositional feature
9. Increased losses of sand and shingle alongshore as a result of a change in the angle of incidence of waves (e.g., as the result of the growth or removal of a shoal, reef, island, or foreland, or because of breakwater construction or a change in wind regime)
10. Intensification of obliquely incident wave attack as the result of the lowering of the beach face on an adjacent sector (e.g., as the result of dredging, or scour due to wave reflection induced by seawall or boulder rampart construction or beach mining, or bow waves generated by ships passing through nearshore waters)
11. Increased losses of sand from the beach to the backshore and hinterland areas by onshore wind action, notably where backshore dunes have lost their retaining vegetation cover and drifted inland, lowering the terrain immediately behind the beach and thus reducing the volume of sand to be removed to achieve coastline recession
12. Increased wave attack due to a climatic change that has produced a higher frequency, duration, or severity of storms in coastal waters
13. Diminution in the caliber of beach and nearshore material as the result of attrition by wave agitation, leading to winnowing and losses of increasingly fine sediment from the beach, either landward into backshore dunes or seaward to bars and bottom deposits
14. Diminution of beach volume by weathering, solution, attrition, or impactation (e.g., by heavy vehicle traffic), resulting in lowering of the beach face and coastline recession
15. Increased scour by waves reflected from backshore structures, such as seawalls or boulder ramparts, leading to reduction of the beach that fronted them
16. Migration of beach lobes or forelands as the result of longshore drifting – there is progradation as these features arrive at a point on the beach, followed by erosion as they move away downdrift
17. A rise in the water table within the beach, due to increased rainfall or local drainage modification, rendering the beach sand wetter and thus more readily eroded
18. Increased losses of beach material due to outwashing by streams or from drains carrying an augmented discharge of water, either because of increased rainfall or from melting snow or ice, or because of natural or artificial modifications (such as urbanization) in the catchments that have increased runoff through beaches
19. Intensification of wave action where tide range has diminished, as on the shores of a bay or lagoon that has been partly enclosed by natural or artificial structures, impeding tidal ventilation
20. Erosion where driftwood deposited on the beach is jostled by wave action, leading to erosion of sand and gravel from the beach face
21. On Arctic coast beaches the removal of a protective sea ice fringe by melting, so that waves reach the beach (e.g., for a longer summer period) and subsidence due to the melting of permafrost within a beach

On many coasts the seaward margins of salt marshes and mangrove swamps have retreated in recent decades, so that they terminate in small cliffs, fronted by tidal mudflats or sandflats. In southern England and north-western France, and in New England, seaward cliffing of salt marshes is very extensive, and similar features are seen on mangrove swamps on the west coast of Peninsular Malaysia, in Thailand, and the Philippines, possibly in response to a relative sea-level rise in recent decades.

Artificial Coastlines

Many coastlines have become artificial, especially during the past century, as the result of the construction of seawalls, boulder ramparts, and other structures intended to halt erosion on cliffs, beaches, and deltas. Spits such as Ediz Hook in Washington and Sandy Hook in New Jersey, previously variable in configuration, have been armored with seawalls or boulder ramparts. Breakwaters built to shelter harbors or form



Coastline Changes, Fig. 5 The Bruun Rule states that a phase of sea level rise on a sedimentary coast will be accompanied by erosion (R) of the beach as a volume of sediment ($V_1 = V_2$) is transferred from the

backshore to the nearshore zone in such a way as to restore the transverse profile landward from D, the initial seaward boundary of nearshore sand deposits (Copyright-Geostudies)

marinas are now widespread on the coasts of North America, Europe, and the Mediterranean, and some coastal areas have been embanked and reclaimed for ports, industry, and urban development, notably in the Netherlands, China, Singapore, Hong Kong, and Tokyo Bay. Reclaimed areas are typically bordered seaward by earth banks or concrete walls, which have become extensive on the coasts of Britain, Europe, and Southeast Asia, particularly Japan, where about a quarter of the coastline is armored with seawalls and tetrapods (Walker and Mossa 1986). Artificial islands formed by deposition of sediment off Dubai in the Arabian Gulf have been reduced by erosion except where they are protected by sea walls.

Many coastlines have been modified by land reclamation: an example is Samphire Hoe, an artificial terrace constructed in 1990–1994 in front of chalk cliffs west of Dover in south-east England using rock waste excavated from the Channel Tunnel from Folkestone to Calais (Fig. 6).

The building of artificial structures is still a widely preferred response to erosion and submergence of coastlines, but increasing use is being made of beach nourishment as a means of defending coastlines against erosion, which also provides a recreational resource of scenic value.

Cliffed coastlines have been modified where roads have been built on bridges in front of them, as at Coalcliff in New South Wales (Fig. 7) and on the Muscat coast.

Coastline Changes, Fig. 6

Samphire Hoe, an artificial terrace fronting Chalk cliffs on the Kent coast west of Dover (Copyright-Geostudies)



Coastline Changes, Fig. 7 Sea Cliff Bridge, built in 2004 along the cliffed coast at Coalcliff, south of Sydney, New South Wales, Australia, to provide an alternative to a corniche road that was frequently closed by rock falls (Copyright-Geostudies)

**Summary**

Changes that occur on cliffs, coastal glaciers, volcanoes, deltas, beaches, swamps, and artificial coasts are described and analyzed. Reference is made to mass movements on steep coasts and 21 causes of erosion on beaches are listed.

Cross-References

- ▶ [Changing Sea Levels](#)
- ▶ [Cliffs, Erosion Rates](#)

- ▶ [Coastal Changes, Gradual](#)
- ▶ [Coastal Changes, Rapid](#)
- ▶ [Coastal Subsidence](#)
- ▶ [Deltas](#)
- ▶ [Glaciated Coasts](#)
- ▶ [Shoreline and Coastal Terrain Mapping](#)
- ▶ [Volcanic Coasts](#)

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Coasts, Coastlines, Shores, and Shorelines

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Origins of Terms

The word shore comes from the middle English word *Schöre* that came from the middle low German word *Schöre*. Shore is also an archaic past tense and past participle of shear. In the German language, the verb shear is *scheren* (Betteridge 1965). *Schere* (or shear) means shearing, cutting, or a point of division. Thus, the middle English and low German *schöre* translates as the sheared line (or shoreline) between the ocean and the land. In the German language, *schöre* is a line and not a region. However, the English usage of shore (from *schöre*) became a description of the *area* adjacent to a body of water. Webster's unabridged dictionary defines shore as "land at or near the edge of a body of water" (McKechnie 1979). The term shoreline is used to describe the *line* marking the edge of a body of water.

The German language actually uses a different word to describe the strip of land adjacent to a body of water. The German word *ufer* is used for bank, shore, beach, or edge. Thus, *ufer* is more akin to the English shore, and *schöre* is more akin to the English shoreline.

The word coast comes from the middle English and old French word *coste* meaning rib, hill, or shore. In Webster's

unabridged dictionary, coast, is the edge or margin of land next to the sea. The close similarities between the words shore, shoreline, coast, and coastline have lead to many ambiguous and inaccurate usage that are unacceptable for scientific dialogue. Scientific usage of these terms are very specific and require explanations related to their description and use.

Geometric Attributes

Coasts, coastlines, shores, and shorelines have geometric characteristics that have precise denotations in scientific nomenclature. Casual usage of the words line, surface, and prism is not appropriate because they are too imprecise or ambiguous. Below are formal definitions of these terms as they should be used in scientific dialogue.

Shorelines and coastlines are lines designating the boundaries between fluid and solid media. A *line* is a one-dimensional construct used to signify the sheared edge between two different materials. A line has length but not breadth. The term coastline and shoreline are both boundary lines between water and land. The term coastline is generally used to describe the approximate boundaries at relatively large spatial scales. Shoreline is used to describe the precise location of the boundary between land and water.

Strips of land adjacent to shorelines and coastlines are called shores and coasts, respectively. Shores and coasts are bounded on one side by a body of water and on the other by a change in terrain to an upland landscape. The *surfaces* of these land strips are two-dimensional features having length and breadth but no thickness. While shore surfaces may have relief, the relief does not give the surface a thickness that implies a volume. The relief only describes elevation changes on a two-dimensional surface. The main difference between a shore and a coast is one of scale. Shores are relatively narrow strips of land adjacent to water bodies, whereas, coasts generally depict relatively broad bands of land adjacent to water bodies.

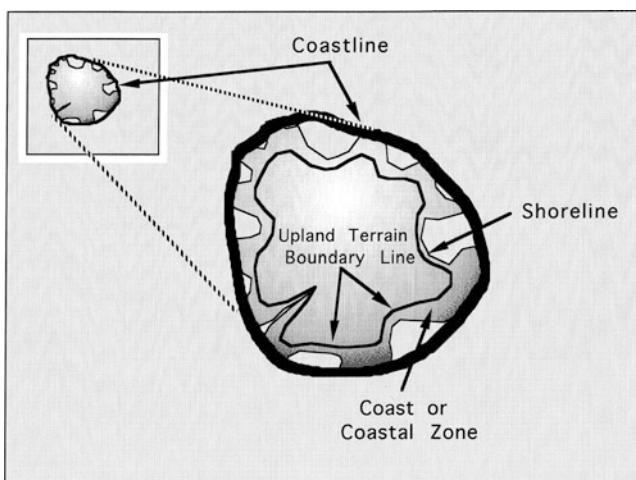
In a geologic context, processes of erosion, transportation, and deposition sculpt the relief on land surfaces. These processes involve the removal, transport, or deposition of a volume or mass of material, and imply a three-dimensional quality of the shore. The shore must be considered a three-dimensional prism when considering the processes that build or destroy it.

Shore surfaces are two-dimensional areas capping the three-dimensional shore prism. Shorelines are one-dimensional boundaries at the edge of the shore. Generally, the terms shore and coast are used to describe the prism of material below the elongate land surfaces at the edge of a body of water.

Coasts and Coastlines

The terms *coast* and *coastline* are used for describing the boundary between the land and the water at broad regional scales. The coast is a strip of land of variable width that extends from a body of water inland to a regional break in terrain features. The coastline is the line that forms the boundary between the coast and a major body of water.

At regional to global scales, the term coastline (rather than shoreline) is used to differentiate the boundaries between the land and the sea. At these scales, changes in margin shapes and boundary orientations are used to describe continents, oceans, and marginal seas. Even at smaller regional scales, coastlines are used to describe the marginal shapes and boundary orientations of estuaries, bays, deltas, barrier islands, and spits. Coastlines may be accurately portrayed on maps having global or regional scales, but the widths of lines constructed at these scales are too large to locate the precise physical position of the sheared boundary between land and water. For example, a volcanic island on a very broad-scale map may look elliptical in shape, but upon closer inspection the border between the land and the sea may illustrate a serrated boundary created by headlands and pocket bays (Fig. 1). The same effect would occur at regions of long straight coasts where the precise water boundaries of small estuaries and bays are scarcely discernible at large scales. The broad-scale maps “smooth-out” small coastal irregularities. Thus, the coastline is generally not as precise as the associated shoreline that traces the land–sea boundary at much finer detail. The coastline is a line connecting the most seaward promontories into the sea. Consequently, a length of shoreline is generally much longer than the associated length of coastline (Fig. 1).



Coasts, Coastlines, Shores, and Shorelines, Fig. 1 Hypothetical planview of a volcanic island showing the apparent shapes of the island from a broad-scale view and a close-scale view. The coastline is oval-shaped, whereas the shoreline illustrates all of the shore irregularities

Since the coast is a strip of land adjacent to the coastline its precise boundaries are indefinite and general. The coast is the strip of land between the coastline and a change in terrain to upland landscapes. The coast includes specific coastal environments or landscapes that are distinct from the upland (e.g., bays, estuaries, coastal dunes, maritime forests, maritime climates, tidal waters, etc.).

At early periods in the Earth's history, there were no images or surveyed maps to precisely locate land–water boundaries. Boundaries could only be deduced from other indirect sources of data. Coastlines from these data sources have imprecise locations with respect to their spatial and temporal setting. For example, at periods of 10^4 – 10^5 years, water volumes in the ocean fluctuate in response to the expansion or contraction of continental glaciers. Glacial waning increases the volume of the sea causing water to transgress over the land surface, while glacial expansion causes water to regress off the land surface. The vertical shift in water level produces horizontal shifts in the position of the boundary between the two media. A paleogeographic depiction of the coastline during these periods may be reconstructed using terrain slope and sea-level projections. These are generalized locations of the boundaries, rather than precise paleo-shoreline locations. A time series of coastline positions on a map may be used to illustrate the regression of water off a land surface, or the transgression of water over a land surface. Thus, temporal scaling also influences our interpretation of coasts and other large-scale areas. Since short-term events generally produce small changes to coastal landscapes, the record of these events cannot be observed at large coastal scales. At broad spatial scales, only large changes produced over long temporal intervals can be observed.

As the sea transgresses over, or regresses off a land surface, the process necessitates a lateral shift in the position of the coastline. At large spatial and temporal scales, the antecedent topography might be considered a passive surface being inundated by rising water. The process is “morphostatic,” meaning that the coastline shift is driven by flooding of a surface rather than erosional sculpting of the surface.

Shore and Shorelines

The terms *shore* and *shoreline* are used for describing the boundary between the land and the water at very fine local scales. The shore is a strip of land of indefinite width that extends inland from a body of water. The terrain of the shore is primarily composed of forms generated by active or recent marine and marine-related processes. The landscape of the shore generally has a unique plant community draped over the terrain. The landward limit of the shore is located at a boundary separating these terrains and plant communities from upland landscapes. The shore is a dynamic area that has many ephemeral forms sculptured by a variety of marine processes. The landward part of the shore that is primarily

molded by aeolian processes is called the backshore (Fig. 2). The surface of the backshore often has a succession of plants draped over a chronosequence of dune ridges between the water's edge and the upland areas. The berm and foredune areas of the backshore are located on shifting sandy surfaces that are sparsely vegetated with specialized herbaceous plants and grasses that tolerate intense light, heat, drought, burial, and salt spray. Dune ridges and swales are located on relatively stable surfaces that have dense concentrations of stress-tolerant grasses and shrubs.

Common grasses that colonize dune surfaces in warm temperate zones include sea oats (*Uniola paniculata*), American beach grass (*Ammophila breviligulata*), and running beach grass (*Panicum amarum*). Common shrubs in the swales between dune ridges include wax myrtle (*Myrica cerifera*), bayberry (*Myrica pensylvanica*), and silverling (*Baccharis halimifolia*).

The seaward part of the shore is called the foreshore, which is defined by tidal range (Fig. 2). The high-water elevation defines the upper limits of the foreshore, and the low-water elevation defines the lower limit of the foreshore (a more comprehensive description of foreshore zones is described below). The foreshore is the portion of the shore where wave-energy is dissipated on the land. The foreshore is sculptured by the dispersion and shifting of granular particles by the energy of waves and littoral currents. The resultant patterns of erosion and accretion make the foreshore one of the most dynamic areas of the shore.

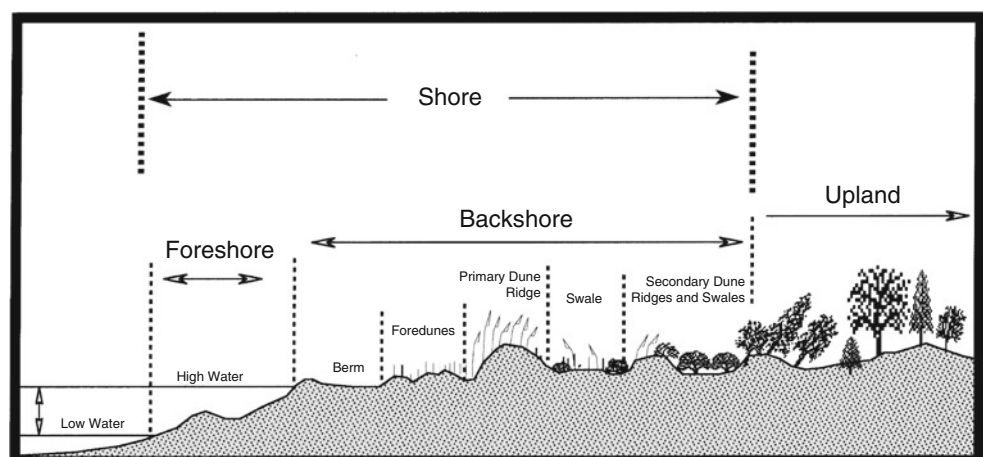
Shorelines

The shoreline is a line that demarks the precise boundary between the shore and the water. Since the position of the water shifts with the elevation of the tide, then the position of the line at high tide is different from its position at low tide. On Fig. 2 it is apparent that there is a high-water shoreline at high tide and a low-water shoreline at low tide. Sea level is generally defined as the elevation halfway between the high-

and low-water elevations, and the generic “shoreline” is meant to describe the line associated with sea level, rather with high or low tide. However, during the rise and fall of the tide, there are an infinite number of shorelines between high tide and low tide.

The elevations of high tide and low tide are not constant. They are dependent on changes in the magnitudes of gravitational and centrifugal forces pulling on the water surface. Changes in the gravitational forces result from the motion and changes in the relative alignments of the Sun, the moon, and the Earth. When the positions of the Earth, moon, and Sun have additive gravitational effects, then the greatest tidal bulges occur (spring tides). A favorable alignment for gravitational addition occurs twice a month producing spring tides. However, the gravitational forces offset each other twice a month producing, small tidal ranges (neap tides). The elevations of each progressive high and low tide are different with the incremental changes in the resultant gravitational forces of the three celestial bodies. In fact, there are over 140 factors involved in the generation and alteration of the tides. The gravitational force of the moon plays a major role in tidal variations. However, the moon's orbit around the Earth is slightly irregular and it takes about 9.3 years for the plane of the moon's orbit to shift above and then below the plane of the solar ecliptic. The cycle is completed in about 18.6 years. These factors can be used to mathematically predict the tidal ranges. An average of all of the high tides and low tides over 18.6 years provides the mean-high and mean-low tide elevations for those predictions. The relationships also can be used to predict the elevations of water level at any position in the 18.6-year cycle, but it assumes no elevation changes are caused by atmospheric conditions. The 18.6-year mean of all high-tide elevations is called mean high water (MHW), the 18.6-year mean of all low-tide elevations is called mean low water (MLW), and the 18.6-year mean of all sea-level elevations is called mean sea level (MSL). There are also means for spring high (MHWS) and spring low tides

Coasts, Coastlines, Shores, and Shorelines, Fig. 2 Hypothetical cross section of a shore showing the classification into backshore and foreshore based on topographic features and water levels



(MLWS), as well as neap high (MHWN) and neap low tides (MLWN). These temporally averaged elevations can also be affixed to associated sheared lines (shorelines) between the ocean and the land. To standardize the use of the term shoreline, it is appropriate to affix it to a specific water level, such as mean high-water shoreline (MHW-shoreline), mean sea-level shoreline (MSL-shoreline), mean low-water shoreline (MLW-shoreline).

Atmospheric conditions also have influences above and beyond the gravitational effects described above. Low-pressure systems can elevate water surfaces, winds can pile water on the shore, and waves may runup on a shore above the still-water elevation of the sea. Daily effects of these atmospheric events on water level result in “actual” water levels that are generally different from predicted values.

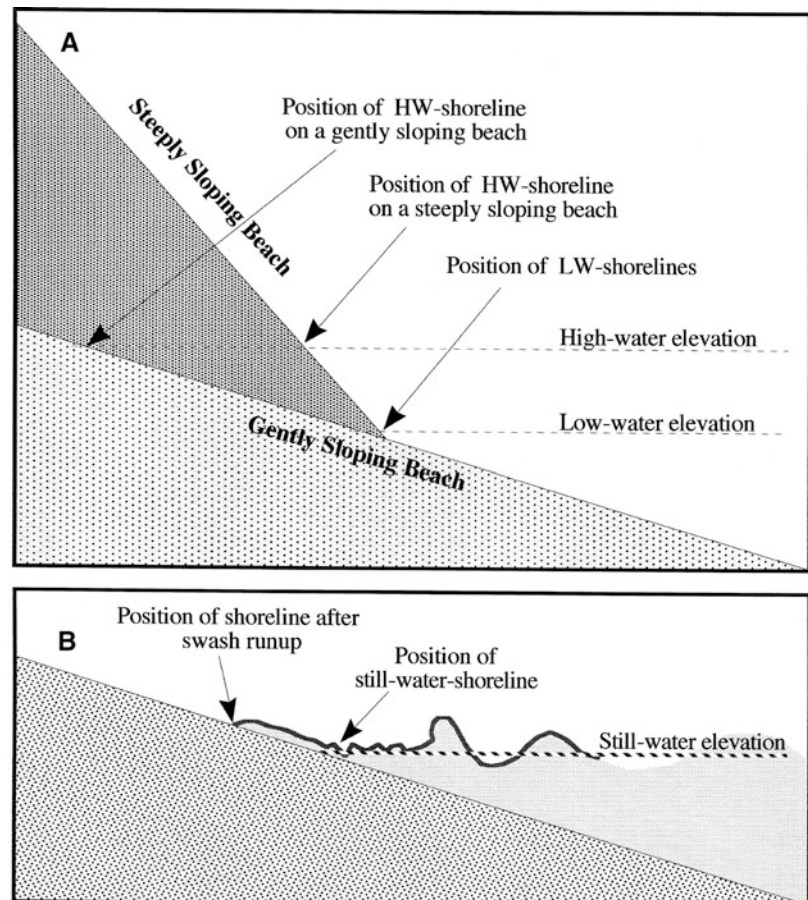
The slope of the shore profile also has a very important influence on the how far water levels may shift up the shore surface (Fig. 3). Typically, beaches have slopes between 1 and 3°. At gentle slopes, a 1 m rise in water level related to atmospheric conditions may cause the position of a shoreline to shift landward 188 m. This type of shift can be accomplished without any addition or removal of sediment from the shore (“morphostatics”).

Breaking waves are other processes that influence the position of the wet and dry boundary at a shore (Fig. 3b). As waves break at the toe of the shore, the wave-bores surge forward and strike the beachface. The momentum of the water continues forward and the water (called swash) rushes up the beach profile. Swash often rises 1.5 times the wave height above the still water level (SWL). Since wave climate is spatially and temporally variable, then the average elevation of the swash limit is seasonally different, and different along dissimilar areas of the coast. If an average wave height can be estimated for a region’s wave climate, then a mean swash elevation (MSE) can be estimated for any reach of shore. Similarly, a standardized mean swash shoreline would represent the wet/dry boundary of a shore influenced by its associated wave climate. Thus, on any given day the predicted water levels from gravitational tides are rarely accurate because of the variability in atmospheric phenomena. Field and remote observations of shoreline locations are temporally very specific, and the sheared line between the water and the land is generally not related to a specific standardized elevation.

Geologists and oceanographers do not agree on the placement of shorelines on maps and charts. Since geologists are

Coasts, Coastlines, Shores, and Shorelines, Fig. 3

(a) Cross-sectional sketch of a steeply and gently sloping shore. Given a constant vertical tidal range, the horizontal spreads of water across a gently sloping surface is considerably greater than across a steeply sloping surface. (b) After a wave breaks, the rush of water up the beachface (swash) shifts the location of an observed shoreline landward of the projected still-water-shoreline



primarily interested in depicting land surfaces, the high-water shoreline is generally used as the datum for these maps. Topographic quadrangles published by the US Geological Survey (USGS) place the shoreline at mean high water or at the North American Geodetic Datum (NAVD 29) which is generally very close to mean high water. On the other hand, the National Oceanographic and Atmospheric Administration (NOAA) publishes charts that are primarily for navigational purposes. Oceanographic cartographers consider any surface that is subaerially exposed as land whether it is always exposed or only exposed periodically by tidal inundation. The datum (and shoreline) for NOAA charts is positioned at MLW or mean lowest low water. It is for this reason that the intertidal zones on maps and charts are often not surveyed, and are simply described as flats and marsh without elevation data.

Shoreline Dynamics

Short- and long-term processes govern the positions of shorelines. A change in shore form created by shifting volumes of sediment is called morphodynamics. The processes of sediment erosion, transportation, and deposition along the shore often result in shifts in shoreline positions. Magnitudes of shoreline shift are expressed as linear units (e.g., meters or feet) where negative values often indicate landward retreat and positive values indicate seaward shifts. During short-term events, shorelines at different elevations on the beachface respond differently. During storms, material is often eroded from the upper part of the shore and deposited at the toe of the shore. As a result, the MHW-shoreline may shift landward, while MLW-shoreline shifts seaward. In this case, the directions of shift for the MHW and MLW-shoreline are opposite, while the MSL-shoreline may have remained stationary. Thus, the use of only one shoreline is insufficient to study the short-term morphodynamic events. However, over long periods, the shift direction of the upper, middle, and lower

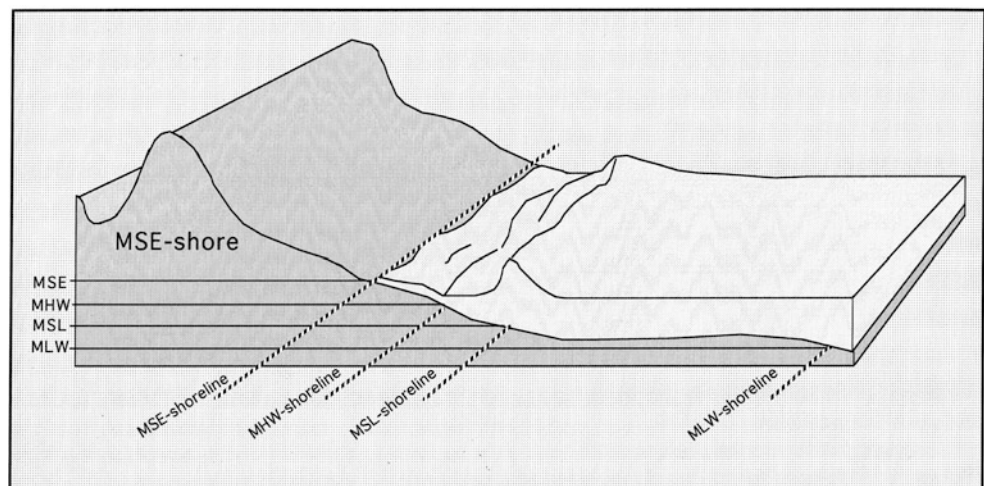
shorelines tend to be the same, and a specific shoreline might be reliable for studying long-term changes.

Rates of shoreline shift may be used to describe how fast a shoreline is shifting. The rates of shoreline shift are expressed in linear units per time (meter/year or feet/year). During short-term events, consistency in the use of a specific shoreline is important since the upper and lower parts of the beachface respond differently to storms. During long-term intervals (annual to decadal), the gross shift rate and the net shift rates may be significantly different since the directions of long-term shift are bipolar. For example, if a shoreline retreated landward 5 m in 5 years and then advanced 5 m seaward for the next 5 years, the gross shift rate would be 1.0 m/year, while the net shift rate would be 0.0 m/year. At very long-term rates (centuries to millennium), shoreline shift rates are impacted by the chronic Holocene rate of sea-level rise.

Shores and Shore Prisms

Since a shore is a strip of land associated with a shoreline, a standardized subdivision of a shore is logically defined by the series of mean shorelines described above. The mean shorelines are associated with specific elevations on the shore surface. These elevations are associated with a horizontal planar surface that extends under the land surface (Fig. 4). The material above each planar surface represents the sand prism of a specific portion of the shore. For example, the mean low-water shore (MLW-shore) is the prism of sand located above the mean low water elevation (and between the MLW-shoreline and the landward limit of the backshore). Each of the standardized shorelines has an associated standardized shore. The mean sea level shore (MSL-shore) is the prism of sediment located above the elevation of mean sea level. The mean high-water shore (MHW-shore) is the prism of sediment located above the elevation of mean high water. The mean swash elevation shore (MSE-shore) is the prism of dry sand

Coasts, Coastlines, Shores, and Shorelines, Fig. 4 A block-diagram sketch illustrating the relationships among four “mean” water levels, four mean shorelines, and four shore prisms located above each horizon.



located above the mean limit of swash runup on the shore (Fig. 3b). These shore prisms are valuable units for describing specific changes to the shore. Since each prism is composed of a specific volume of material, then removal of a portion of that volume is by “geological” definition called erosion. The addition of a volume of material by definition is called accretion. Note that erosion and accretion requires the addition or removal of a mass or volume of material.

Shore Accounting

Beach profiles are often made perpendicular to the shoreline to study beach processes. The area between the profile and MLW elevation represents a two-dimensional section of the shore. When combined with a one-dimensional length of shoreline, the resultant prism is a three-dimensional section of shore. Thus, pairs of profiles can be used to estimate the volume of sand stored in reaches of shore. Standardized elevations can also be used to determine the volume of material stored in each of the standardized sections of a shore prism. Thus, an accounting of material can be made for each shore zone that provides information on the erosional and accretional characteristics of the shore. Using the technique of profile comparisons along a stretch of shoreline, rate scales can be devised to normalize the change rate per unit length of shoreline and/or period of time (Oertel et al. 1989).

During storms, material is often eroded from the upper portions of the shore and deposited on the low portions of the shore. The process may result in no net gain to the total shore volume although there was a shift in the location of a large amount of material. An accounting of material in the total MLW-shore might illustrate a stable volume suggesting a passive condition. An accounting of material in the MSE-shore would illustrate erosion, while the accounting of material in toe of the shore between MLW and MSL would show accretion. The dynamic nature of the shore can only be realized by an accounting of material in vertically stratified zones. The shore zones (related to mean water levels) provide a valuable framework for relating patterns of shore-volume change and shoreline shift.

Common Misusage

The most common misuse of shoreline is the expression “shoreline erosion” and “shoreline erosion rate.” A shoreline is a one-dimensional construct on a planar surface. It has length but no width or depth. In this case, it is a line that denotes the boundary between a solid (land) and a fluid (water). By definition erosion is the removal of a mass or volume of material. Since, the line is one-dimensional, it has no volume or mass. Eroding the line would be equivalent to erasing the boundary between land and water. In strict scientific terminology the expressions “shoreline erosion” or

“shoreline erosion rate” are nonsensical. Shorelines shift landward or seaward across a surface, but do not erode. The landward shift of a shoreline should be described as shoreline retreat, while the seaward shift of a shoreline is called shoreline advance. A rise or fall in water level (that does not require the processes of erosion or accretion) may induce shoreline shifts, or it may be a response to erosion or accretion of material from parts of the shore. The advance or retreat of a shoreline is expressed in linear units (meters or feet). The terms erosion and erosion rates must apply to the change in the size of a volume or mass of material.

Since erosion is the removal of a mass or volume of material, shore erosion is used to describe the removal of material from all or part of a section of shore. It is a three-dimensional process requiring a reduction in the size of the shore sediment prism. An increase in the size of a shore sediment prism is described as shore accretion. A decrease in the size of a shore sediment prism is described as shore erosion. Erosion and accretion are expressed in units of volume (generally cubic meters or cubic feet). Erosion and accretion rates are as expressed in volume units per time (e.g., cubic meters/year of cubic feet/year).

The gross amount of material transported in the littoral system is often used as a surrogate for shore erosion. This is another incorrect use of terminology. Material transported in the littoral zone is described as littoral drift. The rate of littoral drift is generally expressed in units of volumetric units per time (e.g., cubic meters of sand per year). This is a transport rate, and is not a substitute for an erosion rate. The rate is a gross amount of material being transported in the littoral zone. Since, erosion is a measure of the removal of material (not transport) the two may not be equivalent. For example, the gross amount of sediment entrained in a littoral current (in a section of shore) may be constant at about 15,000 m³ in a month. During the transport of this material, the littoral current can remove and add material to the shore. If the amount removed is equal to the amount added then there is no change to the shore sediment prism. However, if the rate of littoral drift is not constant along adjacent sections of shore, then erosion or accretion may result. Gradients in the rates of littoral drift along a section of shoreline may be used to imply erosion or accretion rates. That is, if more material is being transported into a section of shore than is being transported out of a section of shore, then the rate of accretion will be the net difference between the input and output rates.

The use of remotely sensed data is a popular tool for studying rates of shoreline shift. By comparing shoreline locations at two different times, the amount of shift can be estimated over a specific duration. The resultant rates are often used to predict rates of change. Typically, the shoreline is defined by the wet/dry boundary found on an image. Since, each image represents a “snap shot” in time and not one of the standardized elevations (described above),

then it is difficult to discern whether any shift was caused by changes in the volume of sediment at the shore (morphodynamics), or changes in water level (morphostatics). Changes in water levels could be a response to different combinations of tidal stage, barometric pressure changes, waves runup, and wind stresses. On a flat beach, these forces could combine to produce several meters of water-level rise resulting in over 200 m of shoreline shift (without a change to the position of a MSL-shoreline or any change in the size of the shore sand prism). This is the $\pm$ error that should be attached to shoreline shift measurements at low-profile beaches. Therefore, it is very difficult to use remotely sensed data to analyze rate of shoreline change without vertical control.

Cross-References

- [Beach Processes](#)
- [Changing Sea Levels](#)
- [Coastal Changes, Rapid](#)
- [Cross-Shore Sediment Transport](#)
- [Littoral Cells](#)
- [Longshore Sediment Transport](#)
- [Remote Sensing of Coastal Environments](#)
- [Shoreline and Coastal Terrain Mapping](#)
- [Tides](#)

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Cohesive Sediment Transport

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Definition

The transport of fine-grained sediment comprised primarily of clay- and silt-sized particles, mixed with organic matter and, sometimes, quantities of very fine sand.

Introduction

Sediments are a major component of the suspended load and bed load in coastal water bodies (estuaries, lagoons, river mouths, tidal flats, etc.). These sediments can be classified as cohesive and noncohesive. Cohesive sediments are comprised primarily of clay-sized ($<2\ \mu\text{m}$) and silt-sized ($<62.5\ \mu\text{m}$) particles, mixed with organic matter and, sometimes, quantities of very fine sand. Noncohesive sediments are primarily sand and gravel-sized material ($>62.5\ \mu\text{m}$). The underlying property that distinguishes cohesive sediments from noncohesive sediments is that they are cohesive in nature, exhibit poroelastic or viscoelastic rheological characteristics and non-Newtonian behavior when in a fluid-like state (Mehta 2002). Cohesion is the predominance of attractive forces over repulsive forces, such that particles in close proximity can bind together to form flocs (aggregates). This can be attributed to the presence of clay and colloidal particles with surface physico-chemical forces that are significant. A clay fraction greater than about 10% is generally sufficient for the sediment to exhibit cohesive properties (van Rijn 1993).

Cohesive sediments are transported to coastal waters from alluvial and marine sources. Alluvial sediments are derived from various terrestrial processes and human activities and are comprised of both cohesive and noncohesive materials. The fine sediment load is often termed as wash load in river mechanics. The coarser sediments are deposited in the upper reaches of the river/estuary while the fine sediment fractions are carried further downstream into the estuarine/coastal waters where salinity enhances aggregation, forming flocs that possess higher settling velocities. These aggregates then deposit in areas where velocities are sufficiently small. Marine sediments originate from the nearshore bottom. They are first suspended into the water column by the action of waves and currents and are subsequently transported upstream by the net upstream near-bed flow to zones of low velocities where they eventually deposit. Understanding cohesive sediment transport thus requires knowledge of both sediment behavior and water movement.

Because of the affinity of pollutants to cohesive sediments as a result of adsorption, cohesive sediment transport plays an important role in characterizing the transport and fate of pollutants (e.g., heavy metals, radionuclides, pesticides, and nutrients) input to coastal waters. These pollutants originate from various land-use and marine activities (e.g., urbanization, deforestation, industrialization, mining, agricultural practices, hydrosystem construction works, and dredging/disposal activities) posing a potential threat to receiving waters. Ways to mitigate critical water quality problems are receiving the concentrated attention of environmental engineers, biologists, and ecologists. As a precursor to assessing the impact of pollutants input to a water body, a precise understanding of cohesive sediment dynamics is first necessary.

Dynamics

Cohesive sediments in coastal waters come under the influence of various complex processes. In addition to the transport processes of advection and dispersion in the water column, the principal processes influencing cohesive sediment dynamics are aggregation, settling, deposition, consolidation, and resuspension. These mechanisms are not only dependent on the inherent physico-chemical characteristics of the sediment but also on their interaction with the flow field. It is this interdependence that makes the problem of cohesive sediment transport so complex.

Aggregation

Aggregation refers to the formation of flocs comprised of primary sediment particles. The process of aggregation results from destabilization and collision. When the double layer around each sediment particle is compressed by divalent ions, the particles are destabilized and van der Waals attractive forces predominate, creating a condition conducive to aggregation. Collision occurs as a result of three primary mechanisms: Brownian motion, internal shear, and differential settling. Brownian motion occurs when the sediment particles are bombarded by fluid particles moving in random motion resulting from thermal gradients. Internal shear is due to the presence of velocity gradients in the suspending medium. Differential settling is caused by larger aggregates colliding with smaller aggregates. Brownian motion is responsible for aggregation in stationary or quasi-stationary waters forming aggregates that are weakly bonded. Internal shear predominates in dynamic aquatic systems, and the aggregates produced are more dense and durable. Differential settling is important during near-slack periods where concentrations are high. The resulting aggregates are weak and have low density. Aggregation transfers mass through the particle size spectrum toward large-sized aggregates that can be characterized by their higher porosity, increased irregularity and fragility, and higher settling rate (Krone 1963). Aggregation is influenced by sodium adsorption ratio, pH, salinity, sediment size, shape, gradation, density, turbulence, temperature, and the efficiency of collision between particles.

Settling and Deposition

The settling velocity of cohesive sediments is dependent upon the aggregation process, the suspended sediment concentration, and the ionic concentration of the suspending medium. From laboratory experiments of San Francisco Bay sediments, Krone (1962) noted that the settling velocity is independent of the suspended sediment concentrations at low concentrations ($0.1\text{--}0.3\text{ kg m}^{-3}$); with increasing concentration ($1\text{--}10\text{ kg m}^{-3}$), the settling velocity increases with concentration; and at high concentrations ($>10\text{ kg m}^{-3}$), the settling velocity decreases with increasing concentration

and hindered settling occurs. Fluid mud is associated with a lutocline (i.e., strong vertical concentration gradients) when sediment concentrations exceed 10 kg m^{-3} (McAnally et al. 2007). Reported values of settling velocity in estuarine and coastal waters range from 10^{-7} to 10^{-3} m s^{-1} (Mehta et al. 1989).

As the aggregates settle toward the bed, the near-bed turbulence controls whether the particles bond with particles on the bed or are broken up and reentrained into the water column. The stochastic nature of the near-bed turbulence is characterized by a “probability of deposition,” which is defined as the probability that particles reaching the bed actually stick to the bed (Krone 1962). This is conveniently expressed as a function of the bed shear stress and a critical shear stress for deposition. The critical shear stress is a function of the type of sediment in suspension and is generally determined from laboratory flume studies. If the critical shear stress for deposition is equal to or less than the bed shear stress, then deposition will cease provided the sediment has uniform properties (Mehta et al. 1989). The deposition rate is thus a function of the aggregate settling velocity, near-bed concentration, and the probability of deposition.

Consolidation

Once the sediment aggregates reach the bed, the flocculated structure breaks down and the aggregates will move close together. The sediments then consolidate under their own weight when particle-to-particle contact is made. The effective stress (i.e., the difference between the total hydrostatic pressure and the pore water pressure) of the soil increases with the release of pore-water pressure. The void ratio of the bed decreases with depth of sediment (i.e., the density and shear strength both increase with depth). Consideration of the consolidation of cohesive sediment beds is crucial for determining whether the bed is susceptible to erosion and how much material can be eroded. The shear strength of consolidated sediments has been estimated from empirical relationships between shear strength and dry density. Experimental results indicate that the shear strength increases with an increase in the clay content, organic matter, salinity, sodium adsorption ratio, and cation exchange capacity. The shear strength decreases with an increase in temperature, pH-value, and the sand concentration in the sediment bed.

Resuspension

Sediment resuspension depends upon the bed shear stress induced by the flow and the resistance of the bed to erosion. In coastal waters, bed shear stresses are due to the combined effect of waves and currents. These shear stresses can be quite large compared to shear stresses due to currents alone. The resistance to erosion depends upon the sediment type and mineralogy, pore and eroding fluid concentrations, the time history of deposition (i.e., whether the sediments are recently

deposited, partially consolidated, or part of a more dense bed), and chemical and biological processes. To account for the multiple-dependence between the above factors and their spatial variability, investigators have relied on laboratory and field experiments to assimilate and analyze sediment cores from various locations. The results of these studies have indicated the need to obtain system-specific data if accurate predictions of erosion are to be realized. The nature of the deposit (i.e., the bed properties) is generally characterized by a critical shear stress for erosion. Investigations on sediment erosion have indicated that when the bed shear stress at the sediment-water interface is greater than the critical shear stress of the surface in contact, then the material will erode. The amount of material that can erode is described as a function of the stress in excess of the critical shear stress.

There are two primary modes of erosion, namely mass erosion and surface erosion. Mass erosion refers to the resuspension *en masse* of sediments when the bed shear stress exceeds the critical shear strength of the bed at some depth below the sediment-water interface. Clumps of sediment are eroded. This erosion mechanism occurs when the flow is rapidly accelerating as in zones of strong tidal currents and under storm conditions. Surface erosion, on the other hand, occurs at low to moderate values of the excess shear stress where currents are low to moderate (Mehta et al. 1989). Surface bed sediments are eroded particle by particle until the bed shear stress is less than the critical shear stress for resuspension. The amount of material that can erode from the bed is limited to a finite amount. Laboratory experiments (Parchure and Mehta 1985; Tsai and Lick 1987) and field studies (Amos et al. 1992) have revealed that only a finite amount of sediment can be resuspended from a cohesive sediment bed exposed to a constant shear stress as a result of the increase in shear strength with depth of sediment.

Mathematical Modeling

To effectively assess the impact of sediment input to coastal waters, it is necessary to predict their spatial and temporal distributions in the water column and in the sediment substrate. In order to do this, it is imperative to consider the interaction between sediment transport processes and the flow field. The advent of sophisticated mathematical models coupled with inexpensive computing power has spearheaded major advances in developing simulation models. These models now strive to incorporate more complex sediment processes into tractable computational techniques. Since the accuracy of cohesive sediment transport predictions is contingent upon the accuracy of the flow field, an understanding of the hydrodynamic circulation processes is essential. In lieu of this, models for hydrodynamics and waves are generally

coupled to a sediment transport model so that the various models operate within the same computational framework, in conjunction with one another, with output from one serving as input to another.

The simplest model for cohesive sediment transport is based on zero-dimensional modeling (e.g., Krone 1985). These models are based on the sediment conservation equation, which represents the mass balance between sediment inflows, outflows, and storage (i.e., sediment deposition). The spatial variability of the sediment properties is ignored. Other models of cohesive sediment transport range from simple one-dimensional to sophisticated three-dimensional models. Some of the available models are described in Martin and McCutcheon (1998). The governing equation is based on the mass conservation principle in one-, two-, or three-dimensions, with boundary conditions of no-sediment flux at the free surface, and net erosional or depositional flux at the sediment-water interface. Odd and Owen (1972) used a two-layered one-dimensional model to simulate cohesive sediment transport in the Thames Estuary. Other one-dimensional models (e.g., Lin et al. 1983) do not describe the near-bed concentrations, which are required for determination of deposition rates. Two-dimensional models may be depth-averaged or laterally averaged. The assumption is that concentrations are well-mixed over the depth or in the lateral direction. Depth-averaged models (Ariathurai and Krone 1976; Onishi 1981; Cole and Miles 1983; Lick et al. 1994; Shrestha 1996; Shrestha and Orlob 1996) compute the sediment deposition rates from mean sediment concentrations averaged over the vertical. Laterally averaged models (Ariathurai et al. 1977; Onishi and Wise 1982) describe the distribution of sediment concentrations in the longitudinal and vertical directions. Quasi-three-dimensional models account for the vertical distribution of sediments using approximations (e.g., calculates the vertical concentration profile based on the vertical velocity distribution) (Lou et al. 2000). The current state of the art is to utilize three-dimensional finite difference or finite element models for sediment transport (Hayter and Pakala 1989; Sheng 1991; Onishi et al. 1993; Shrestha et al. 2000). These models are suitable for applications to systems where the three-dimensionality of the flow, the salinity and temperature structure, and the associated suspended sediment distribution are important.

Sediment processes embedded in the model generally include aggregation, deposition, consolidation, and resuspension of sediments. In addition to the flow field, model inputs include bed properties (i.e., bed type, grain-size distribution, dry density, and sediment resuspension characteristics), fixed and/or time-varying sediment loading from various sources, and sediment settling parameters. Sediment aggregation is generally embedded in the settling velocity formulation and is described as a function of the sediment concentration. In most models, sediment consolidation is

accounted for by discretizing the bed into a vertically segmented bed model (Shrestha et al. 2000), where the sediment layers depend upon the time after deposition. Each layer is assigned a certain thickness, density, and a critical shear stress for erosion based on laboratory or field estimates of consolidation. The assumption is that a recently deposited bed is more susceptible to erosion than an aged bed. Alternatively, consolidation is directly simulated using a consolidation algorithm in the numerical model. The resuspension from the bed is based not only on the excess shear stress but also on the age of layer. Output from the sediment transport model includes the spatial and temporal distribution of suspended sediment concentrations, the mass of sediment eroded or deposited, and subsequent change in bed elevations. Model performance is generally assessed through model-data comparison of suspended sediment concentrations and sedimentation rates. Sensitivity of the model input parameters is also assessed. Experience has shown that sediment-loading rates need to be estimated correctly if realistic predictions are to be realized.

The hydrodynamic model is essentially the driver for the sediment transport model. It provides the necessary transport information. Martin and McCutcheon (1998) review a variety of hydrodynamic models ranging from one-dimensional to three-dimensional. The dimensionality of the model is selected based on the intended use of the model. The governing equations include the continuity, momentum, and constituent (i.e., temperature and salinity) transport equations, with an equation of state relating density to temperature and salinity. Inputs to the hydrodynamic model include hydrographical (steady-state and/or time-varying flow rates from rivers and nonpoint land-based runoffs), astronomical (tides), meteorological (winds and heat fluxes), and internal density gradients (temperature and salinity distributions). Hydrodynamic models should possess the capability to predict different time scale events, ranging from semidiurnal tide scale, diurnal scale, meteorological scale (few days), spring and neap tidal scale (15 days), in addition to monthly and seasonal time scales. Model output includes water surface elevations, currents, horizontal and vertical diffusivities, and distributions of temperature and salinity. The predictive capabilities of the model are assessed via extensive comparisons with data, and a confidence is established that the model realistically reproduces the predominant flow dynamics. As an example of a particular three-dimensional model application and its comparison with field data, the reader is referred to Blumberg et al. (1999).

Bed shear stresses due to currents and waves are crucial for calculating sediment resuspension and deposition fluxes. Wind waves are high frequency short waves of periods 3–20 s. These waves not only increase mass and momentum fluxes, but also increase bed shear stresses by inducing large velocities near the bottom of the water column. The

development of wind waves in coastal waters is based on a balance between wind energy input, wave energy, and wave energy dissipation. The wave climate is predicted from a wave model such as SWAN (Holthuijsen et al. 1993), HISWA (Booij and Holthuijsen 1995), the Great Lakes Environmental Research Laboratory (GLERL) wave model (Schwab et al. 1984), SMB (USACE 1984), WAVD (Resio and Perrie 1989), or ACES (Leenknecht et al. 1992). Inputs to the wave model are bathymetric depths and a two-dimensional, time-dependent surface wind field. Linear wave theory is used to translate the wave height and period into a near-bed peak orbital velocity and peak orbital amplitude. A wave-current model can be used to calculate the bed shear velocity. The bed shear stress is then calculated as the product of the fluid density and the square of the shear velocity.

One of the critical features in the application of the above models is the type and quality of data available for creating model inputs and for calibration and validation of the model. In this regard, data assimilation and analyses represent a core component of any modeling effort. Because of the widespread use of such modeling tools, sediment transport studies are often supplemented by field sampling and monitoring programs. For many studies involving the fate and transport of sediment-associated pollutants, hydrodynamic, wave, and sediment transport computations are linked to one or more water quality models.

Summary

This entry emphasizes the key mechanisms influencing cohesive sediment transport in coastal waters. These processes are complex because of the interaction between the physico-chemical properties of the sediment and flow parameters. Mathematical modeling provides a tool to assess the impact of the predominant processes on the transport and fate of sediments discharged to coastal water bodies.

Cross-References

- [Beach Sediment Characteristics](#)
- [Cross-Shore Sediment Transport](#)
- [Erosion Processes](#)
- [Estuaries](#)
- [Longshore Sediment Transport](#)
- [Nearshore Sediment Transport Measurement](#)
- [Sediment Budget](#)
- [Tidal Environments](#)
- [Wave Climate](#)
- [Wave–Current Interaction](#)

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Conservation of Coastal Sites

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Conservation of coastal resources has been practiced throughout human history as communities have endeavored to manage the food, water, and other resources of the coast in ways which ensure that each season or year will bring at least comparable bounty from the sea and the land. Protection of coastal sites in the historic past usually meant protection of areas to be used for elite recreation or as locations where coastal resources were to be conserved. Only in recent centuries, however, have communities begun to take specific action to conserve coastal sites for their natural flora and fauna. This has meant that some areas of coastal land and water have been designated as more important for conservation than others. This may have led to distortions in the behavior of coastal ecosystems which are just as important as the distortions which conservation is intended to prevent.

What does “conservation” mean? Usually, it means management in ways which protect the inherent natural and human features and processes of an area. Conservation is often confused with “preservation,” but natural features

change. Change is normal, but it is usually a means by which sites retain their resilience.

There are many different types of coastal sites which are conserved, including archaeological, biological, cultural, geological, geomorphological, and zoological sites. There is also a wide variety of approaches to their identification, designation, and protection (Table 1).

Conservation for Biological Reasons

Ehrenfeld (1976) argued that protection of ecosystems was usually justified by two purposes: scientific and human. Typically, these include research and education, the need to protect genetic information, and as parts of the worldwide and interconnected pattern of ecosystems. Often, however, they

Conservation of Coastal Sites, Table 1 Typology of coastal conservation sites

Level of designation	Features of importance for designation	Exemplar categories	Exemplar sites
Global	Having features of ecological, biological, geological, cultural, archaeological significance at international level	World Heritage Site	Great Barrier Reef (Australia)
		Biosphere Reserve	Danube Delta (Rumania)
		Ramsar Site	Karavasta Lagoon (Albania)
Regional	Habitats of regional significance sites of importance for migrating birds	European SACs	Chesil and the Fleet (UK)
		European SPAs	Wadden See (NL, D, and DK)
National	Key habitats or species	NNRs	Studland (UK)
		Biological Reserve	Isla del Caño (Costa Rica)
		National Wildlife Refuge	Kodiak (USA)
	Protection of fauna and flora with extensive recreational activities	National Parks	Fuji-Hakone-Izu (Japan)
	Protection of fauna and flora with limited recreational activities protection of marine areas	National Monument	Fort Jefferson (USA)
		Marine Parks	Oshimoto (Japan)
		Marine Sanctuary	Farallon Islands (USA)
		MPA	Milieuzone Noordzee (NL)
		Marine Life Conservation District	Waikiki (USA)
	Protection of sensitive estuarine areas	National Estuarine Research Reserve	South Slough (Oregon, USA)
	Protection of site of scientific importance (geological or biological)	SSSI	Dungeness (UK)
		Site	Anse du Gris-Nez (France)
	Recreational site with ecological significance	National Recreation Area	Oregon Dunes (USA)
	Protection of archaeological, anthropological, or historic site	Area of Archaeological Importance (AAI)	Hengistbury Head (UK)
		Anthropological Reserve	Cota Brus (Costa Rica)
		National Historic Site	San Juan Island (USA)
Local	Protection of locally important habitats or species	Local Nature Reserve (LNR)	Lundy Island (UK)
		Ecological Reserve	Tomales Bay (USA)
	Protection of fauna and flora and provision for recreational activities	State Park	Humboldt Lagoons (USA)
	Local protection of marine habitats with extensive recreational activities	Municipal Marine Park	Balicasag Island (Philippines)
		State Seashore	Del Norte (USA)

have been protected because of the perceived importance of particular species. Usually, these are rare and, commonly, higher plants and animals. Conservation was often a response to particular lobby or pressure groups, sometimes even individuals, whose concern about losses of species and habitats led them to establish groups to provide protection to pressure governments into protecting vulnerable species. This was particularly the approach in nineteenth century North America and Europe. Today, the responsibility lies more with government bodies or conservation nongovernmental organizations (NGOs).

Conservation for Geological or Geomorphological Reasons

Geological and Geological and geomorphological sites are usually protected so that stratotypes, fossil assemblages, or structures can be observed and studied or so that specific important features are preserved and protected. The coast is a particularly good location for the observation and investigation of geological features because erosion constantly exposes new strata and because large sections can be observed without excavation. In many countries, the coast provides an accessible location for earth sciences research and education. Sites may also be important because they demonstrate the processes which mold the landscape both now and in the past. Geological sites commonly fall into two groups: “process” sites and sites which display features of geological significance. The British Geological Conservation Review which describes all designated geological and geomorphological sites of national and international importance in Great Britain refers to “process” and “integrity” sites (Table 2). Most fossil locations or stratotypes are integrity sites. At Dungeness, for example, the ridge sequence provides

a time sequence which is an integral part of recent coastline history. If it disappears that part of the history goes with it, hence it is an “integrity” site. In contrast, chalk cliffs of the Seven Sisters exemplify well the processes which occur at the coast and so this is predominantly a “process” site.

Archaeological and Historical Sites

Archaeological and historical sites include major buildings and structures located at the coast as well as sites where artifacts have been identified and excavated. Some sites are now well removed from their original coastal location by the effects of emergence, whereas others have been submerged. In shallow coastal waters, there are often very important submarine archaeological sites associated with wrecks, especially in the more active sea lanes. Some archaeological sites are at risk of disappearance as erosion cuts into them. Unless the site has considerable architectural or historical importance, erosional sites are usually not protected.

Conservation Legislation

Although the purpose of conservation laws is usually the same, that is, the protection of features or processes of high scientific or cultural value, the detailed legislation depends upon the legal framework of each nation or state. France, the United States, the United Kingdom, and Japan each have well-developed systems of conservation legislation which lay down specific procedures for the identification, designation, and management responsibility for scientifically important sites. These include coastal sites. The way in which management takes place also depends upon local tradition

Conservation of Coastal Sites, Table 2 Selected geomorphological coastal conservation sites

	Dungeness	Chesil beach and the fleet	Seven Sisters and Beachy Head	North Norfolk coast
Feature	Cusate shingle foreland	Shingle tombolo and lagoon	Chalk cliffs	Barrier beaches and spits
Key geomorphological importance	Shingle ridges in sequence from c 5,500	Large coastal fringing tombolo	Rapidly retreating	Barrier islands
	Years BP	Sediment grading	Chalk cliffs	Complex spits and
Archaeological importance	Reclamation from		Intertidal platforms	Reclaves
	Roman period to present		Cliff-top Iron Age settlements	Reclamation from mediaeval times
Biological importance	Holly (<i>Ilex aquifolium</i>) wood invertebrates	Little terns (<i>Sterna albifrons</i>) intertidal and subtidal organisms	Rock-boring organisms intertidal flora and fauna	Salt marshes
Conservation status	SSSI	SSSI	SSSI	SSSI
	Geological	GCR	GCR	GCR
	Conservation review	SAC	SAC	SAC
	Site (GCR) SPA	SPA		SPA

and legislation. Thus in Britain, coastal conservation is carried out both by statutory bodies such as English Nature (a government body) and by NGOs such as the National Trust and the Royal Society for the Protection of Birds (RSPB). In addition, smaller local bodies such as Wildlife Trusts also own or lease land so that they can conserve and manage areas in which rare or endangered habitats or species occur. However, the statutory bodies often are advisory organizations rather than owners of the sites. Site management depends upon the landowner or occupier of the site. This means that there may be many different organizations involved on conservation of coastal sites within short distances of each other. Their conservation policies and practices may conflict.

In many countries, there is little or no tradition of nongovernmental conservation bodies practicing coastal conservation and so responsibilities lie with government. Thus in France, the government sponsored body, the Conservatoire de l'Espace Littoral et des Rivages Lacustres (CELRL) was established to make coastal lands more accessible at the same time as protecting landscapes and properties of national importance. Modeled in part on the UK's National Trust, it is a governmental body rather than a nongovernmental membership body such as the National Trust. In some locations, First Peoples have traditionally regarded coastal lands as sacrosanct, but have no tradition of legislation to protect these lands. As a result, there are conflicts in conservation practice between traditional views of the value of coastal sites and their legal protection. Examples include the conflicts between European-based legislation and traditional approaches of the Maori in New Zealand. Legislation derived from British law conflicts with traditional law.

There may also be differences between local, national, or international legislation. Thus in the United States, both federal and state legislatures can define sites of coastal conservation and in Europe, the system of Directives which are mandatory on all member states may supplement or even override national conservation practices. The European Union's Birds (European Commission 1979) and Habitats (European Commission 1997) Directives have established Europe-wide networks of sites including many at the coast. These are known as Special Protection Areas (SPAs) and Special Areas of Conservation (SACs). In Britain, the seaward boundary of coastal sites such as National Nature Reserves (NNRs) and Sites of Special Scientific Interest (SSSIs) established under the national legislation lies at Low Water Mark, but the boundaries of SACs and SPAs can extend to the territorial limit. There is a very wide range of types and levels of conservation designations worldwide which includes World Heritage Sites, Biosphere Reserves, and Ramsar Sites designated at international level, National Parks, and Marine Sanctuaries at national level and local nature reserves (Table 1).

Conservation Measures in Different Countries

In Europe, there are considerable differences in the history of coastal conservation, usually as a result of the relative importance of government and nongovernmental involvement and action at the coast. Furthermore, there has been considerable conservation action in the past decade as a result of initiatives by the European Community.

France has the longest established nature conservation legislation in Europe. Protection of sites is the responsibility of the Ministère des Affaires Culturelles, with advice provided by the Commission Supérieure des Sites, Perspectives et Paysages and the Commissions Départementales des Monuments Naturels et des Sites. Decisions at local level must be approved by the national Commission Supérieure. The designation of sites followed the Laws of May 2, 1930 and July 10, 1976, which identified the following criteria for classification of sites as nature reserves:

1. The preservation of animal and plant species and habitats which are either in danger of extinction in all or part of the national territory or have outstanding characteristics;
2. The restoration of animal or plant populations or habitats;
3. The conservation of botanical gardens and arboreta forming reserves for plant species which are rare, unusual, or nearing extinction;
4. Preservation of biotopes and outstanding geological, geomorphological, or speleological features;
5. Preservation or establishment of staging points on the great migration routes;
6. Scientific and technical studies necessary for the development of human knowledge;
7. Preservation of sites of particular interest for studies of the evolution of life and the activities of mankind.

This approach is reflected within the European Union Directives on Birds and Habitats both of which include important coastal areas.

The largest of the French coastal nature reserves is the Camargue (13,171 ha) on the Rhone delta. In contrast, when it was designated as a Site in 1930, Port Cros, a small island off the coast of Provence, had an area of only 140 ha. The initial designation was based on landscape protection, as more was known about the newly protected site, especially its flora and fauna, the significance of the surrounding seas was acknowledged. The site was extended in 1963 to include the surrounding sea, and became the first Marine Protected Area in the Mediterranean. The inclusion of the surrounding sea recognized the importance of such species as Neptune Grass (*Posidonia oceanica*) and Grouper (*Epinephelus marginatus*). This example typifies the development of coastal conservation as sites are first identified to protect a particular species or habitat and then expanded to account for

improved understanding of the site and its surroundings. However, conservation can be very slow in practice. For example, at Port Cros the conservation objectives have been restricted by the impact of over half a million visitors annually and local resistance to conservation measures, such as a ban on angling from the shore which restricts traditional coastline activities.

The 62 coastal Départements may designate for protection the habitats of protected species that are threatened and acquire land financed through the Taxe Départementale des Espaces Naturels Sensibles (TDENS: Departmental Tax for Sensitive Natural Areas), often in association with the CELRL (Meur-Férec 1997). The Law of May 2, 1930 allows for designation of either Site Classé (classified site) or Site Inscrit (listed site) for sites of special natural and built importance and has been used extensively in the protection of coastal sites (Meur-Férec 1997). The 1986 Loi Littoral (Coastal Law) provided a statutory framework which recognized the economic, social, and heritage importance of the coast and the unique nature of the coastal zone (Cicin-Sain and Knecht 1998). This provided a context for individual sites, including, for example, a setback zone of 100 m wide for the whole coast, in which development is prohibited. Marine areas can be declared protected under the 1968 Law of Maritime Hunting, essentially to protect wildfowl! In many countries with long-established traditions of hunting, conservation of these species has often meant controlled management of habitats. Although there are some strongly held objections to game conservation, it is important to emphasize that without these conservation measures many sites which are valued today for their biodiversity would not exist. The 1960 Loi Relative à la Création de Parcs Nationaux (Law for the Creation of National Parks) includes Marine Protected Areas (MPAs) by allowing the terrestrial parks to extend into the maritime region (IUCN 1994). European Union (EU) regulations and directives further supplement the national policies. Under the Birds (1979) and Habitats (1992) Directives, 133 sites have been proposed for inclusion within the European Community-wide Natura 2000 network of sites.

Two government bodies also have a role in French coastal conservation. The most important is the CELRL established in 1975. It is a public body charged with acquiring coastal and lakeside land, maintaining its ecological character and allowing public access to its sites (Cicin-Sain and Knecht 1998; European Union for Coastal Conservation 1995). Once purchased, the land is managed by the CELRL on behalf of the state in perpetuity. However, any acquisition has to take account of the area development plans and prior consultation with the Départements and Communes. This is because the day-to-day management and costs of an acquired site are handed back to the commune(s) or relevant landowner after purchase (EUCC 1995). It currently owns over

49,000 ha along 650 km of coastline (Meur-Férec 1997), including the D-Day beaches in Normandy and the Domaine de Certes at Arcachon. The Office National des Forêts (ONF, the National Forestry Office) manages 100,000 ha along 500 km of coastline, mainly forested dunes, such as the Landes. It works with the CELRL and other related administrative bodies to produce management plans for the forests. In the coastal region, ONF's work is more social and ecological than its predominantly economic role on the inland forests (Favennec 2000). In particular, it is responsible for dune stabilization work, where it often has to balance the needs of conservation with tourism and public access. Thus at Pointe d'Arçay it has implemented a management plan devised with public participation and at Cap-Ferrat has successfully reintroduced marram grass (*Ammophila arenaria*) to the dunes.

Coastal conservation in Britain started, as so often there, with voluntary bodies such as the National Trust and the Royal Society for the Protection of Birds. Previously, conservation measures were largely in the hands of individual landowners or the Crown and were mainly concerned with management of game or food resources. In 1949, the National Parks and Access to the Countryside Act established the Nature Conservancy as well as the national parks. The Nature Conservancy and its successors (three country agencies: English Nature, the Countryside Council for Wales (CCW), and Scottish Natural Heritage) were concerned with the geological and biological importance of sites for research and education. National Nature Reserves were established together with areas designated as SSSIs. The Royal Commission on Historical Monuments describes and studies sites of historical and archaeological importance, including many coastal sites. The Ancient Monuments and Archaeological Areas Act 1979 provides for the protection of archaeological sites and monuments and does appear to extend into territorial waters, although there is no record of designations at sea. These Acts and bodies are generally empowered only to deal with sites above Low Water Mark. However, the European Directive on Habitats specifies Marine SACs and so specific attention now has to be given to subtidal and submarine sites around the UK coastline. The Directive has force for the coastal waters of all European member states, although the detailed legislation is made at national level.

Submarine and marine sites are dealt with under different legislation. The Merchant Shipping Act 1995 deals with wrecks and salvage, requiring reporting of finds and the Protection of Wrecks Act 1973 provides some protection for important or dangerous wrecks or their sites. War graves may be designated under the Protection of Military Remains Act 1986: diving on and interference with a designated war grave is an offence. International agreement to protect graves in civilian shipwrecks in international waters was given effect in the Merchant Shipping and Maritime Security Act 1997.

The biological sites were reviewed in detail by Ratcliffe (1977) and coastal geomorphological and geological sites have been the subject of a major national review (the Geological Conservation Review [GCR]) which has identified almost 100 coastal geomorphological sites (May and Hansom 2003), plus a number of coastal landslide sites and several hundred geological sites. The Nature Conservation Review (Ratcliffe 1977) described three stages in site classification:

1. Recording the features of each site to describe the range of ecosystem variation in terms of environmental and biological characteristics;
2. Comparing the quality of these sites; and
3. Choosing a series of nationally important sites.

The criteria used in both assessment and selection of key sites were

1. Size of the site;
2. Diversity;
3. Rarity of the species;
4. Naturalness, that is, areas which have been modified least by humans in so far as they are identifiable;
5. Fragility: habitats which are highly fragmented, rapidly decreasing in number and/or area, difficult to recreate or threatened with extinction;
6. Typicalness;
7. The recorded history of the site;
8. The position within an ecological or geographical unit;
9. The potential natural value of a site;
10. The intrinsic appeal of a site.

The Nature Conservation Review identified 140 coastal SSSIs with a mean area of 1,957 ha; 8 sites exceeded 10,000 ha in area. At the same time, France had 131 sites of average size 456 ha. 78% French sites were less than 100 ha in area, whereas 42% British sites exceeded 1,000 ha. For coastal geomorphological sites, criteria have included not only the specific importance of the site, but also the processes affecting particular rock-types or structures. As a result, sites with comparable geological conditions have been selected with micro- and meso-tidal ranges and different wave climates. There are 100 geomorphological sites which include large shingle structures such as Dungeness, major landslides such as the west Dorset coast, major dunes such as Morrich More on the east coast of Scotland, and world-renowned features such as Lulworth Cove and the stack of the Old Man of Hoy.

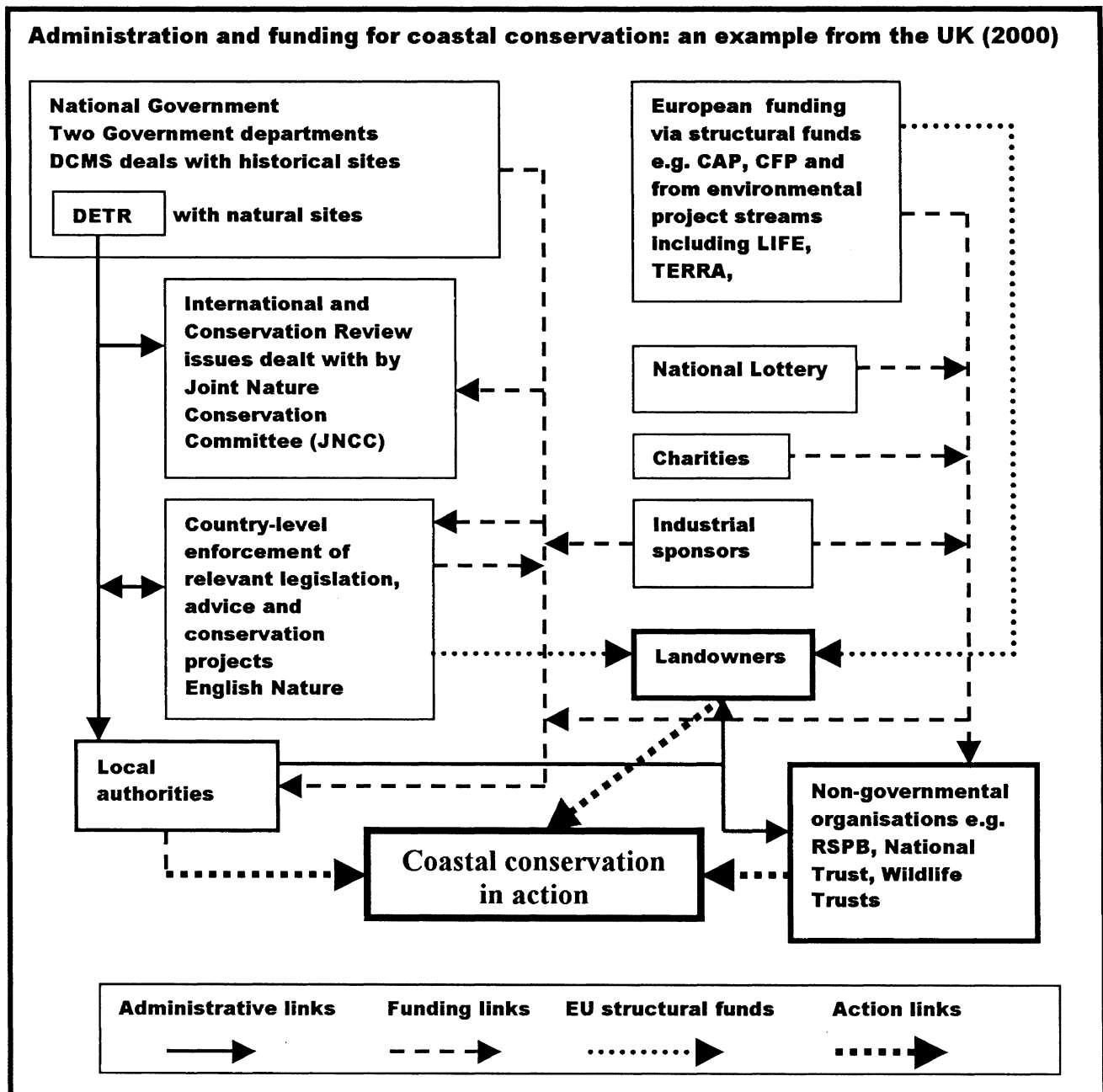
Although their coastal conservation is based upon different traditions both in coastal use and in their laws, France and Great Britain show very well how complex the legislation and administration of coastal conservation has become. Countries with shorter experience have been able to adopt different, and

sometimes more efficient, approaches. Furthermore, the European Community has adopted its own framework for conservation in order to establish a community-wide network of related sites.

Funding of coastal conservation is similarly complex and depends upon receiving levels of funding which allow management to be carried out. The British model relies heavily upon voluntary support, membership, industrial support, bequests, and fundraising (Fig. 1).

Typically, the largest voluntary bodies have strong marketing and sales departments. Funding from national and local government is limited in amount and has to be competed for. Such funding is often restricted in time. In contrast, in France or the United States, for example, conservation is predominantly funded by official sources, with the national and local governments largely responsible for both funding and operation of the conservation sites. The manner and volume of funding may change with a change of government. Thus, although the arrangements shown in Fig. 1 for government funding applied in 2000, a reorganization of government departments following a General Election in June 2001 altered some of the responsibilities. In Britain, local authorities have only limited inputs into conservation funding, whereas the French Départements have considerable fiscal and legal tools at their disposal. They may, for example, acquire land financed through the TDENS (Departmental Tax on Sensitive Sites). Often these acquisitions are made in conjunction with the CELRL, and all the 62 coastal Départements have used these powers to purchase land (Meur-Férec 1997). The day-to-day management and costs of an acquired site are handed back to the commune(s) or relevant landowner after purchase (EUCC 1995). The majority of funding comes from the state but legacies and public/corporate donations also help with the acquisition and management of sites. Apart from the State, the CELRL is France's second largest landowner but interestingly, its budget is one-fifth of the National Trust's, a British NGO that operates a similar protection by acquisition policy (Meur-Férec 1997).

At the time of the French Law of 1976, many European countries had hardly established relevant legislation. Greece, for example, had very good legislation concerning its archaeological sites, but was only just beginning to identify natural sites (May and Schwartz 1981). Similarly, Norway's Nature Conservation Act of 1970 initiated action to preserve and protect plants, animals, and birds and areas of importance to nature conservation which, within a decade, established a network of 13 national parks, almost 200 nature reserves and over 200 locations which were of natural scientific importance. Many are at the coast. In contrast, Poland created its first National Park inland at Bialowieza in 1932 and there are now eight protected coastal areas including two National Parks, the Slowinski (founded 1967: present area 18,619 ha) to the west of Gdansk and the Wolinski Park near the German



Conservation of Coastal Sites, Fig. 1 An example of funding for conservation of coastal sites. Key: *DCMS* Department of Culture, Media and Sport, *DETR* Department of the Environment, Transport and the

Regions, *CAP* Common Agricultural Policy, *CFP* Common Fisheries Policy, *RSPB* Royal Society for the Protection of Birds

border (founded 1960: present area 10,937 ha). The Polish Environmental Protection Bill 1992 provides for three “levels” of designation: Strict (usually no admission to the public: 22% of sites), Partial (where management controls are flexible: 59%), and Landscape Protection (where so-called “soft laws” apply: 19%). This recognizes that although some areas must have strict controls over access in order to protect sensitive areas and species other areas are more resilient and more appropriate for human activities such as

recreation. National ecological policy requires that 30% of Polish territory (including coastline) should eventually be designated as “protected,” 22% of the area had been designated by 1997. Coastal conservation in Poland is thus a national responsibility rather than a regional or local one. This is a typical pattern of many countries which have relatively recently begun to conserve coastal sites. The strength of national legislation often, but not always, ensures that designation and appropriate funding occur.

Spain combines national and local approaches to conservation. It has a long history of environmental legislation but the first specific conservation (in its modern sense) legislation was the 1975 Law on *Espacios Naturales Protegidos* (ENPs: Protected Natural Areas). This proposed four categories of protected area: national park, natural park, natural area of national importance or interest, and reserve of scientific interest. Other later laws have established over 25 different categories (WCMC 1992). The national body responsible for administration of protected areas is the Instituto Nacional para la Conservación de la Naturaleza (ICONA: National Institute for the Conservation of Nature). Set up in 1971, its functions include the encouragement of renewable resource use and maintenance of ecological balance; the creation and administration of national parks and sites of national interest; and the development and exploitation of inland fishing and hunting assets. With decentralization of the government in 1978, responsibility for many conservation issues passed to Spain's 17 autonomous communities (*Comunidades Autónomas*) which are responsible inside territorial waters for the protection of marine and coastal areas from human activities. The 1989 Law on the Conservation of Natural Areas, Flora, and Wildlife strengthened this responsibility further. The protection of coastal areas is covered by the *Ley de Costas* (the Shore Act: July 28, 1988) based on the French model (Barragan 1997). The Shore Act requires protection of the sea up to 100 m offshore and restricts the number of road and pedestrian routes to the coast. It also defines a zone extending 500 m inland which affects land and urban planning as well as prohibiting the disposal of waste in this area (WCMC 1992). As in France, this provides a national framework within which conservation of coastal sites may take place. Laws concerning sustainable use and conservation of living marine resources have been adapted from the Directives of the European Community since these require member states to bring them into force.

The Waddensee: An Example of International Cooperation in Coastal Conservation

International cooperation is sometimes essential if the integrity of coastal ecosystems is to be protected, one of the best examples being the Waddensee in the southern North Sea. The Wadden Sea is the largest unbroken stretch of intertidal mudflats in the world and is extremely productive. With a biomass considered equivalent to rainforests, it is recognized as one of the most important wetland areas in the world. Up to 30 km wide, the Waddensee extends from the northeast Netherlands through Germany to the southwest coast of Denmark and includes low barrier islands, tidal flats, salt marshes, sand dunes, woodland and grassland, and low barrier. It is a very important breeding and overwintering area for a high proportion of the coastal birds of the North Sea. It is also, however, an area of reclamation, oil and gas

extraction, commercial fisheries, and resort development. The three countries have cooperated in its conservation since 1978 using a joint memorandum which provides the basis for planning, conservation, and management of the area. In 1997, they adopted the Waddensee Plan, which was developed by involving all stakeholders and has as its focus the conservation, restoration, and development of all the region's ecological features. The interests of the local population, traditional uses of the area, and habitat conservation must be balanced with the other needs of sustainable management and preservation. Seven principles have been used to bring about regulation of activities that threaten its environment and to provide for active intervention as necessary, namely (de Jong et al. 1999):

- Careful decision making – based on using the best available information
- Avoidance – avoiding potentially damaging activities
- Precautionary principle – avoiding damaging activities, even if there is insufficient scientific evidence to link them to specific impacts
- Translocation – moving damaging activities to areas where they will cause a reduced impact
- Compensation – using compensatory measures to offset unavoidably damaging activities
- Restoration – restoring areas which have been damaged
- Best available techniques and best environmental practice should be adopted.

The most important lessons are that international cooperation is essential for the conservation of some vital sites, that there must be agreed targets and procedures, and that local people must be involved especially when their traditional activities form part of the ecosystem as a whole.

In contrast, although some of the less-developed countries of eastern Europe have taken few steps to conserve coastal sites, the role of international conventions such as the Ramsar Convention on Wetlands of International Importance, 1971 and international cooperation in species protection, especially for marine turtles, have had an impact. In some respects, the significance of tourism as an economic driver for conservation has been overlooked although tourism provides a very strong incentive for conservation and sustainable activities. In Albania, for example, the isolation and centralization of the economy and society over the last 50 years affected community behavior as local communities were excluded from the planning process and decision making about natural resources upon which they depended. Not surprisingly, they consider that conservation and habitat protection have ignored their needs and imposed restrictions without providing benefits.

Albania designated its first Ramsar site in 1994 (RAMSAR 1996, 2001), comprising the Karavasta Lagoon and adjacent Divjaka National Park. In the 1960s, much of the

wetland was drained for agriculture, affecting the biodiversity of the lagoon, as well as the surrounding catchment, the Divjaka forest and hills, and its human communities. The total area of the site is 20,000 ha, made up of native forest of *Pinus pinea*, a species unique to the Albanian coast, and the lagoon itself (5,800 ha) which is a brackish water system connected to the sea by artificially maintained channels. The freshwater input to the lagoon comes from runoff from the surrounding agricultural land and Divjaka hills. Karavasta Lagoon has the only coastal breeding population of the Dalmatian pelican (*Pelecanus crispus*) along the Adriatic and Ionian coasts, supporting up to 5% (700–1,000 individuals) of the total world population. A significant proportion of the European breeding populations of Little Tern (*Sterna albifrons*) and Collared Pratincole (*Glareola pratincola*) use the Karavasta wetlands during migration, as well as large numbers of waterfowl, gulls, and terns, including the globally threatened pygmy cormorant, and white-headed duck.

Despite its designation as a Ramsar site, the lagoon has been seriously threatened by inappropriate tourism development which has often damaged fragile coastal habitats upon which its wildlife depend. Pollution and overfishing have added to the stresses on the lagoonal ecosystem. Some of the reclaimed wetlands are affected by high levels of salinity and so are of limited agricultural value. The area is state-owned and managed jointly by the General Directorates of Forestry and Fisheries (both part of the Ministry of Food and Agriculture). In practice, management depends upon a small number of rangers with limited experience of specialized wetland management, training, or equipment. Future strategies include improved planning and management of the lagoon to both protect the ecosystem and its biodiversity (including the pelicans) and derive economic benefits for local communities. A management plan for the site was prepared with international, technical, and financial cooperation through the European Union's PHARE program (PHARE 1996). The Karavasta Lagoon is an example of where there is a need to involve local communities in biodiversity conservation activities and demonstrate their benefits. Karavasta exemplifies the problem that conservation of coastal sites depends upon management of the human systems which decide how the area will be sustained.

The United States of America

A very wide range of different designations are used in the United States, many within the framework of the National Parks legislation. They range from areas such as National Parks where they exist for both preservation of natural and cultural features and for recreation to specialized research areas, such as the National Estuarine Research Areas that are devoted to research (Table 1). There is a wide network of sites at state and federal level, but the initial establishment

of important sites depended upon very strong individual lobbying of the federal government and the role of individuals such as John Muir in getting presidential support. Coastal sites were generally established later than inland sites.

The origins and variety of the coastal sites is well illustrated by the coastline north of San Francisco (Table 3). From very humble beginnings, some sites such as the Farallones Islands have grown to major national coastal and marine sites. Designations overlap so that within the Point Reyes National Seashore, for example, there are not only marine and terrestrial reserves, but also historic ranches within the Philip Burton Wilderness Area (California Coastal Commission 1987). State, county, and private parks and reserves also occur. However, there are few sites owned or managed by private organizations. Estuaries, such as San Francisco Bay, have been extensively reclaimed with the accompanying loss of wetland habitat. This is a worldwide feature of coasts and serious concerns about the loss of biodiversity and productivity have been expressed within the United States for many decades. Since 1850, reclamation has cut off over half of the bay's tidal marshes and open water from tidal action and 95% of the original wetland habitat has been lost (California Coastal Commission 1987). Nevertheless, San Francisco Bay contains 90% of the coastal wetlands of California and remains a major part of the Pacific coast bird migration route. Responsibility for the wetlands is divided between the Bay Conservation and Development Commission (tidal wetlands) and the US Corps of Engineers (nontidal wetlands), a division which does not recognize the natural hydrological links between them.

In many less-developed countries, funding for conservation is largely provided through tourism receipts, or by international aid programs and funding from international bodies such as the World Wide Fund (WWF) for Nature. Pressure for conservation action and designation typically comes from pressure groups. The implications of conservation action and site or species protection are not always well considered. For example, a ban on turtle fishing along the Pacific coast of Mexico had immediate impacts upon the local economy and knock-on effects to other parts of the coastal ecosystem. The contrast between the approach in many less-developed, but biologically rich, regions is well demonstrated by two examples, Costa Rica and the Mexican state of Oaxaca.

Costa Rica

The Central American country of Costa Rica has a well-established system of designation and protection of its ecosystems and biodiversity, including coastal sites. As the examples in Table 4 show, these sites include very large areas of land and sea which protect different habitats and some small sites which are most

Conservation of Coastal Sites, Table 3 Coastal conservation in the United States – California examples

Site	Designation	Responsible authority	Key features	Area (ha)
Southeast Farallon Island	Bird Reservation (1909)	Presidential order following pressure initiated by Director of California Academy of Sciences	Major seabird nesting site protection to prevent large-scale damage to birdlife by commercial egg companies	26
Southeast Farallon Island	Farallon Reserve (1911)	Re-designation. First statutory protection	Public concern over wildlife endemic Farallon weed major seabird nesting site	26
Farallon Islands	National Wildlife Refuge (1972)	US Fish and Wildlife Service and Point Reyes Bird Observatory	300,000 birds in 12 species nest annually. Over 350 species of birds migrate via the islands World's largest nesting populations of ashy storm petrels, Brandt's cormorants and western gulls northernmost breeding site for northern elephant seals	40
Farallon Islands and surrounding waters (up to 6 nautical miles offshore between bodega head and rocky point and waters within 12 nm of the islands)	Gulf of the Farallones National Marine Sanctuary (1982)	Gulf of the Farallones National Marine Sanctuary (Federal Agency)	Diverse productive marine ecosystem based on upwelling	
Point Reyes National Seashore	National Seashore	National Parks Service	San Andreas fault very diverse ecosystems >360 bird species recorded	Almost 29,000
Limantour Estero	Reserve (within National Seashore)	National Parks Service	Estuarine habitats: removal of all life forms prohibited without permit	202
Bodega dunes	Sonoma State Beach	State Department of Parks and Recreation	Dunes up to 50 m high	364
Bodega	Marine Reserve	University of California statewide Natural Reserve	Undisturbed habitat for research	132
	Marine Life Refuge	System State Department of Fish and Game	Marine area up to 330 m offshore	
Duxbury reef	Marine Reserve	State Department of Fish and Game	Largest exposed shale reef in state Endemic acorn worm and one of only two sites for burrowing anenome (<i>Halcampta crypta</i>)	ca. 16
Tomales Bay	Ecological Reserve	State Wildlife Conservation Board	Wetlands and salt marsh	>200
Tomales Bay	State Park	State Department of Parks and Recreation	Sand beaches	
Bolinas lagoon Olema	Nature preserve	Marin County Audubon	Marsh habitats and species privately owned research and wetland restoration project	485
Marsh	Not designated	Canyon Ranch	Freshwater marsh	16

important for particular species. These sites are all nationally designated and form part of a network of sites designed to represent and preserve Costa Rica's wide biodiversity.

Mazunte: An Example of Conservation in Action

Mazunte on the Oaxacan coast of Mexico is, unlike the Waddensee, in one of the poorest areas of the world. It shows how changes in conservation policy can be accompanied by local community involvement in coastal conservation especially when this is focused upon sustainable

development. However, it also demonstrates the tenuous nature of much coastal conservation in less-developed areas.

For more than 30 years, Mazunte (the name comes from a Nahuatl word meaning "please lay eggs") was dependent upon turtle catching and canning. Most people in the village were either fishermen or worked in the cannery. A Presidential decree in 1990 banned all turtle fishing following significant lobbying by WWF. The immediate effect was to remove the community's main source of income, as the local community express it, throwing "our

Conservation of Coastal Sites, Table 4 Examples of the variety of coastal conservation sites in Costa Rica

Location	Cabo Blanco	Isla del Caño	Manuel Antonio	Ostional
Status	Strict nature reserve	Biological reserve	National Park	National Wildlife Refuge
Area	1,172 ha	300 ha land 5,800 ha marine	682 ha	162 ha land 587 ha marine
Key faunal features	Largest Brown booby (<i>Sula leucogaster</i>) colony in Costa Rica	Coral reefs	Major nesting site for Brown booby	One of the world's most important nesting sites for
			109 species of mammal, 184 bird species, 10 species of sponge, 19 coral, 24 crustacean, 17 algae, 78 fish	Olive Ridley turtle (<i>Lepidochelys olivacea</i>) on beach only 900 m in length
				Also important for Pacific green (<i>Chelonia mydas</i>) and leatherback turtles (<i>Dermochelys coriacea</i>)
				102 bird species
Key plant features	Mainly evergreen forest with especially spiny cedar (<i>Bombacopsis quinatum</i>), chicle tree (<i>Manilkara chicle</i>), and Espave (<i>Anacardium excelsum</i>)	Tall evergreen forest dominated by cow tree (<i>Brosimum utile</i>) coral reefs	Range of habitats: primary and secondary forest, mangrove swamp, marsh and littoral woodland and lagoons major nesting site for Brown booby	Forest includes deciduous trees (especially Frangipani: <i>Plumeria rubra</i>), mangrove, cacti and succulents. <i>Acanthocercus pentagonus</i> cactus grows on beaches and provides shelter for ctenosaurs (<i>Ctenosaura similes</i>)
Archaeological importance		Pre-Columbian cemetery	Pre-Columbian turtle trap	
Geological and geomorphological importance		Plate tectonics: subduction of Cocos plate beneath Caribbean plate along middle American trench	Sand tombolo	

people into misery!" Protection of one species (turtles) was enacted at the cost of another. Some local people abandoned the area and moved inland to occupy land within the coastal dry forest, the associated clearance having potentially damaging effects on coastal lagoons as sedimentation increased. The turtle populations were recognized as being under stress from poaching, predators, and disease. The Mexican Center for Marine Turtle Research (CMT) was established in late 1991, using the buildings of the former Society of Industrial Fisheries of Oaxaca, and now has a unique role as the only center which studies seven out of the eight species of marine turtle. In addition, a community-based project known as Ecosolar and a Reserva Ecologica Campesina were established. Within Mazunte itself, a small ecotourism project was based upon sustainable principles and designed to provide accommodation for mainly foreign visitors to the area. Further to the south, the large-scale resort development

of Huatulco had a major impact on the unusual dry coastal forest and the coastal beaches and lagoons.

The attempts to conserve not only the turtles and their nesting beaches but also to build a sustainable community have not proved easy, for both human and natural intervention have conspired to threaten the coastal ecosystem. In the aftermath of an uprising in 1994 in Chiapas, there was considerable local unrest which affected Huatulco in August 1996. This had a serious impact on the turtle ecology because, when police were sent to the resort, up to 200 poachers went to local beaches where Olive Ridley turtles were arriving to lay eggs. It is estimated that up to one million eggs were lost and large numbers of turtles killed. In October 1997, the area was struck by Hurricane Pauline which caused serious damage to both the CMT and other buildings but also devastated the nesting beaches. 806,000 nests were affected, with half all the nests on Escobilla beach, four million eggs and up to ten million

hatchlings lost. Nevertheless, the turtles continue to nest on these beaches. The CMT carries out detailed studies of turtle behavior and ecology both in the Center and on the beaches and the growth of the resort of Huatulco has brought increasing numbers of visitors to Mazunte. The concept of the Reserva Ecologica Campesina is unusual in that it links deliberately the human community with the coastal ecosystem. Knowledge which once served the commercial interest of turtle canning has been harnessed to develop an approach to sustainable development which has shown itself to be very resilient to significant human and natural disruption. The financial support for the CMT is very small, less than US\$70,000 annually, and receipts from the ecotourism project are also small. Conservation here is not about conservation of a site, so much as conservation of a whole ecosystem (including its human occupants). For many communities, this will be the future for coastal conservation for it is firmly based in local knowledge, the needs of the local community, and their own initiatives.

Conclusion

The conservation of coastal sites takes many different approaches, often with detailed legal and administrative frameworks for designation and protection, but also in relatively informal ways with the involvement of NGOs and private ownership. In some countries, it has only been through the effective active involvement of individuals and NGOs that conservation has been put into practice. The example of Mazunte is a modern, but more community-based, expression of people's wish to protect and conserve features of their environment which they value which was expressed in different words and actions by the founders of the National Trust or John Muir and the Sierra Club. Once established, such initial routes to coastal conservation need the support of the national and international communities to make sure that they are truly sustainable and not threatened by external political, economic, and social pressures. Although coastal conservation is well established in practice, there are nevertheless enormous threats to many sites from urbanization, resource extraction, tourism, pollution, large-scale coastal projects such as barrages, and reclamation and neglect. Sea-level change and climate change also threaten many sites, although their history shows that most sites are resilient to change provided that human activities allow natural resilience.

Note

The International Union for the Conservation of Nature (IUCN) World Conservation Monitoring Centre (WCMC) in

Cambridge, England, is the most important source of information regarding protected sites.

Cross-References

- [Archaeology](#)
- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Marine Parks](#)
- [Rating Beaches](#)
- [Tourism and Coastal Development](#)
- [Tourism, Criteria for Coastal Sites](#)

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Continental Shelves

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Introduction

Continental shelves are the submarine platforms that surround continents. They are found in the open ocean and in many of the coastal seas. Similar features surround subcontinents, such as Greenland, and many oceanic islands. These features exhibit a wide variety of physical dimensions and characteristic relief. Most commonly, they are relatively flat with slopes varying between $1:10^2$ and $1:10^4$. Depths at the seaward edges of continental shelves are generally on the order of 80–100 m, although there are places where depths are less than 50 m or greater than 200 m. Beyond the shelf edge is the continental slope, which is typically inclined on the order of $1:10^2$ to $1:10^3$ and frequently has much more local relief than the continental shelf.

General Morphology

In most places the continental shelf has distinct subzones. Although the nomenclature is not standardized, it is customary to divide the zone between the surf zone and the shelf edge into four major reaches. These are: (1) the *shoreface*, (2) the *inner shelf*, (3) the *mid-shelf*, and (4) the *outer shelf*. On most open-ocean shelves the shoreface, with a slope on the order of $1:10^2$, occurs between depths of 5–25 m. It has a convex upward slope which is steeper at its top. This zone is also occasionally called the shore-rise and is considered by some to be part of the inner continental shelf. The inner continental shelf is usually taken to extend seaward to the point that the coastline no longer significantly influences wind-driven circulation. This distance is strongly affected by the strength and duration of the wind at any given time. It is usually on the order of 5–15 km wide.

The outer-shelf and mid-shelf environments are also defined by their dynamic processes. The outer shelf is where there is significant interaction between deep ocean and shelf circulation patterns. On the mid-shelf, circulation responds to the restricted depths but is not dominated by interaction with ocean currents or near-shore circulation. Because continental shelves have widths in the range of 10^0 – 10^2 km, the characteristics of these subzones are quite variable.

Large-Scale Characteristics

Continental shelves can also be divided according to their modes of origin, dominant processes, and geographic

locations. In most places around the world where the coastline has been tectonically stable the shelves are broad and of low relief. Often these are simply submerged continuations of adjacent coastal plains, which themselves are extensive. Conversely, where continental margins are overriding adjacent seafloor as a result of continental drift, the shelves tend to be narrow and steep. These are often adjacent to coastal mountain ranges. The inner part of continental shelves at low latitudes is a common site for coral reefs and other forms of carbonate platforms. Shelves around the Antarctic continent, the Arctic Ocean, and certain other high-latitude places are strongly influenced by processes related to sea ice. On some high-latitude shelves, permafrost can be found a few tens of meters below the sea bottom. This attests to the combined effects of severe cold and the relatively recent rise of sea level.

The sedimentary environments and deposits on continental shelves are quite varied. In some places rock platforms extend beneath the shelves and sediments are distributed as thin sheets or in patches. These conditions are encountered where the underlying rock is relatively young (such as in volcanic areas), where the supplying of sediments is limited (e.g., west Florida Shelf), or where recently drowned shelves had been scoured by glaciers. Deposits greater than 10^3 m occur where prolific sources of sediments exist (Mississippi Delta foreshore). In some places, the predictable patterns of sandy nearshore sediments grading to silt and mud near the shelf edge are found. However, there are few general rules. In some locations muddy sediments occur all the way from the shore to the shelf edge. Elsewhere, the entire shelf is covered with sandy sediments, possibly with local patches of silt and mud.

Many broad continental shelves, especially those in stable tectonic areas, are sediment-starved today. In these areas the coastline is characterized by broad estuaries. Sediments, transported in river bedload accumulate at the estuary head. The sandy adjacent continental shelves are the result of reworking of previous deposits. In many cases, the previous deposits represent ancient shore processes from times of lower sea level. At the opposite extreme, there are cases where river deltas extend across the modern shelves so that the seaward face blends into the continental slope. The delta of the Mississippi River is an especially good example.

Shelf Features

Although continental shelves tend to have broad areas of low relief, there are many morphological features that characterize this environment. These can be conveniently divided between erosional features, depositional features, and modified relic features. The most common erosive features originated during periods of low sea level when rivers extended entirely across to the shelf edge. Because there were several

major sea-level excursions during the Pleistocene, there have been several extended periods where river and river canyon development have been promoted. In some cases, there are no remnants of these ancient river courses evident in the bathymetric shape of the bottom. However, subbottom profiles show channels now filled with sand. In other cases, quite obvious river valleys extend across the continental shelves to submarine canyons that extend down the continental slope. In some places drowned river terraces and bluffs can be found.

Depositional features dynamically coupled to present sedimentary processes and features representing the reworking of previous deposits are both found. The dynamic sedimentary features include sand ridges, sand waves, and a host of lesser bedforms. *Sand ridges* are long features with characteristic heights of 1–10 m, spacings (i.e., wavelengths) of 10^3 m and lengths up to several kilometers. These may be either shoreface-connected or freestanding sets on the lower shoreface and mid-shelf regions. The *shoreface-connected sand ridges* tend to be oriented obliquely to the shore with the acute angle pointing in the direction of dominant longshore sediment transport (Duane et al. 1972; Swift and Field 1981; Caballeria et al. 2001; Calvete et al. 2001). These features are generally thought to migrate slowly down the coast. The freestanding sand ridges occur in sets in a wide range of shelf environments (Nemeth et al. 2001). They may occur on the tops of shallow sand banks close to the shore. They are also common features in mid-shelf environments. These too are prone to slow migration, although shelf sand ridges in some locations occupy remarkably stable locations.

Sand waves are geometrically similar features that are smaller than sand ridges. They have heights on the order of a few meters, are spaced at distances (i.e., wavelengths) on the order of 10^2 m, and extend 10^2 – 10^3 m in length (Nemeth et al. 2001). These may appear in extensive fields over broad sandy areas or as secondary features on the crests of sand waves and sand banks. Recent research has shown that sand waves are dynamically stable bottom forms that are brought about by shelf currents through a complex feedback mechanism that includes systematic changes in the structure of turbulence. These large bed forms are stable as either fixed or migrating features.

Shoreface-connected and open-shelf sand ridges as well as sand waves tend to occur in parallel sets with quasi-regular spacing. Many other forms of bottom relief are less regular in shape and spacing. These include sand ribbons and irregular bars. The reasons for these features are generally poorly understood, although many represent the results of reworking of sand deposited during an epoch of lower sea level.

In order to explain many of the morphological features that have been developed on continental shelves through the reworking of previous features, it is necessary to briefly

consider sea-level history. Many of the shelf features developed as a result of the migration of the shore across the shelf since the last glacial maximum approximately 18,000 years ago. Worldwide sea level rose approximately 110 m, but the rate was not uniform over this period. The rate of sea-level rise was greatest between 12,000 and 7000 yr. BP. However, throughout the entire recent period the rate of rise has varied within shorter intervals (1500–4000 yr.). More details are provided in Donoghue et al. (2003). Variations in the rate of sea-level rise have caused the shore migration rate to change considerably. At times of slow rise, or rise-reversal, the shore position remains nearly constant. Shore features become well-established and considerable deposits of sand develop. Subsequent episodes of rapid sea-level rise can cause overstepping of these features. They are reworked, but not entirely obliterated. As the availability of dense bathymetric data sets has increased it has become easier to detect and explain these features.

The concept that barrier islands retreat landward in the face of rising sea level is well established. This is brought about by processes that carry sand over and behind the islands while the seaward face erodes. Features formed off the seaward side of these islands can be preserved if their initial volumes are large. In several locations along the northern Gulf of Mexico coast, the huge volumes of sand in ebb tide delta are too great to be disbursed as they are drowned by the ongoing rise of sea level. This results in leaving elongated sand bodies stretching many kilometers onto the shelf. In most cases these features correlate with a position of present estuaries, although they may not be aligned with present tidal inlets.

Because the rate of sea-level rise since the last glacial maximum has been unsteady, with millennia-scale fluctuations, the migration of shore features has also been episodic. In areas where there are plentiful supplies of sand, barrier island complexes extending from the shoreface, across the island, and including the washover deposits in the lagoon, have tended to form during periods when the rate of sea-level change is low. Again, these huge sand deposits resist obliteration when they are drowned during episodes of rapidly rising sea level. Much of the upper portion of these *shore-complexes* is reworked and lost. However, the general morphology can be distinguished where there are sufficiently detailed bathymetric data. In these places subbottom profiles reveal dipping beds representing lower shoreface and back bay deposits (Donoghue et al. 2003).

Many ancient coastal river deltas have also been large enough to avoid being obliterated during periods of rapid sea-level rise. These are now located on the continental shelf. The shelf edge is where many of these are found because the last sea low-stand was below this elevation. Additionally, some major sand shoals on the continental shelf originated as deltas during the transgressive phase of the past sea-level cycle.

Subaqueous deltas are also found on continental shelves. These are sediment clinoforms with nearly flat tops and seaward-dipping foreset beds. However, unlike coastline deltas, these features formed with their tops on the order of 10 m below sea level. Examples of these can be found off the mouth of the Amazon River (Nittrouer and DeMaster 1996), and in the Italian Adriatic (Trincardi et al. 1996). Unlike the drowned deltas described above, these subaqueous deltas are actively forming at the present time.

There are many extensive morphological features, consisting largely of sand, which occur as isolated *banks* or *shoal complexes*. Although many of these have complex histories and origins, there are others that are members of feature classes with common explanations. Swift (1973) has introduced the term *massif* in conjunction with these features. Generally, a *massif* is a very large volume of sand that has originated near the coastline and has subsequently been drowned and reworked due to sea-level rise. *Shoal retreat massifs* originated from the complex of sand shoals that form across the mouth of many broad estuaries. The ancestral location of these estuaries may be far out on the present shelf. As sea level rose, both the estuary mouth and the related shoals migrated landward leaving huge volumes of sand behind. Although these are now being reworked by the action of waves and currents, the very size of these deposits protect them against obliteration. However, there must be many cases where smaller features have not been so persistent.

Cape shoal retreat massifs have also been described (Swift et al. 1972, 1973). These are thought to originate at the apex of coastline capes in barrier island systems that existed far out on the shelf. The system was subsequently drowned by sea-level rise and traces of the barrier islands obliterated. However, the large collection of sand at the point where the littoral transport systems converged produced a linear shoal mass that extends far onto the present shelf. These are often still connected to capes in barrier island systems.

Shoal retreat and cape retreat massifs, along with many other sand banks that originated from the reworking of previous massive sand deposits, often provided the sites for extensive sand ridge and sand wave fields. This is an excellent example of how a portion of many shelf features represents the result of ongoing processes while other aspects of the same features represent more ancient causes.

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Cross-References

- [Barrier Island Landforms](#)
- [Coral Reefs](#)
- [Deltas](#)

- [Offshore Sand Banks and Linear Sand Ridges](#)
- [Offshore Sand Sheets](#)
- [Reefs, Non-Coral](#)
- [Shelf Processes](#)

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Coral Reef Coasts

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Coral reef coasts occur almost entirely between the Tropics of Cancer and Capricorn, though functional exceptions involving fossil reefs may extend beyond these limits. The reef component of a coral reef coast is constrained by the

physiological requirements of the corals themselves, which include clear, warm water, neither too fresh from river outflows (>30 ppt) nor too saline from high evaporation (<40 ppt), with a temperature range of between about $15\text{--}30^\circ\text{C}$, and not heavily sedimented. Exposure levels are rarely limiting; corals thrive from very sheltered to extremely exposed conditions.

Because reefs grow vertically to the low tide level, they provide very effective shelter to the coast, which in turn greatly affects the habitats which lie in their lee. The more closely a reef is located to the coastline, the greater the shelter will be. Extremely sheltered coasts lie behind fringing reefs with broad reef flat components, over which water always remains $<1\text{--}2$ m deep. Reef development may take place some distance offshore, in which case greater fetch and depth between reef and shore leads to higher energies impinging on the coast. In numerous cases, the series of coastline reefs may lie far offshore (east Africa, Great Barrier Reef), possibly even on a line along the edge of the continental shelf, so that coastlines are little protected, added to which the presence of the reef itself may funnel longshore water movement which may add to net coastline erosion and dynamics. In numerous locations, fringing reefs and more distant barriers of reefs, or series of patch reefs, combine to provide complex patterns.

Reefs are biogenic limestone structures, sometimes called bioherms. Thus, over time, reef development shapes the coastline itself, and the shape and morphology of the system decreasingly depends on, or even reflects, the inherited landscape form. While it is usually the case that the limestone-depositing organisms usually can only settle on preexisting hard substrates, they also may develop on unconsolidated substrate where the latter exists in sufficiently sheltered conditions. Good examples of this are found in the Red Sea, a region famous for both its general aridity and excellent fringing reefs lying close to and contiguous with the coast. Here, cores have shown layers of solid reef alternating with layers of unconsolidated sand and rubble of mainly terrestrial origin. The formation of these series comes from alternating phases of reef growth, followed by reef obliteration when heavy rains and sediment deposits carried by flash floods deposit terrestrial materials in thick blankets over the reefs. Alluvial fans develop, prograding to seaward, which then support vast areas of productive shallow water between their leading edge and the shore. Sheltered conditions then allow a new cycle of reef growth to develop on the fans. Depending on current patterns (and repeat episodes of heavy rain) these fans may then infill, extending the land to seaward until greater depth and steep slopes at the seaward edge precludes further progradation.

Uplift of reefs also affects coastline morphology. Clear examples again occur in the Red Sea (Fig. 1) where series of tectonically uplifted reefs extend inland on the coastal

plain. The Red Sea is an example of a spreading plate with current seismic uplift but, equally, substantial examples of uplifted old coral reef coastlines occur on most continents, notably east Africa and Australia.

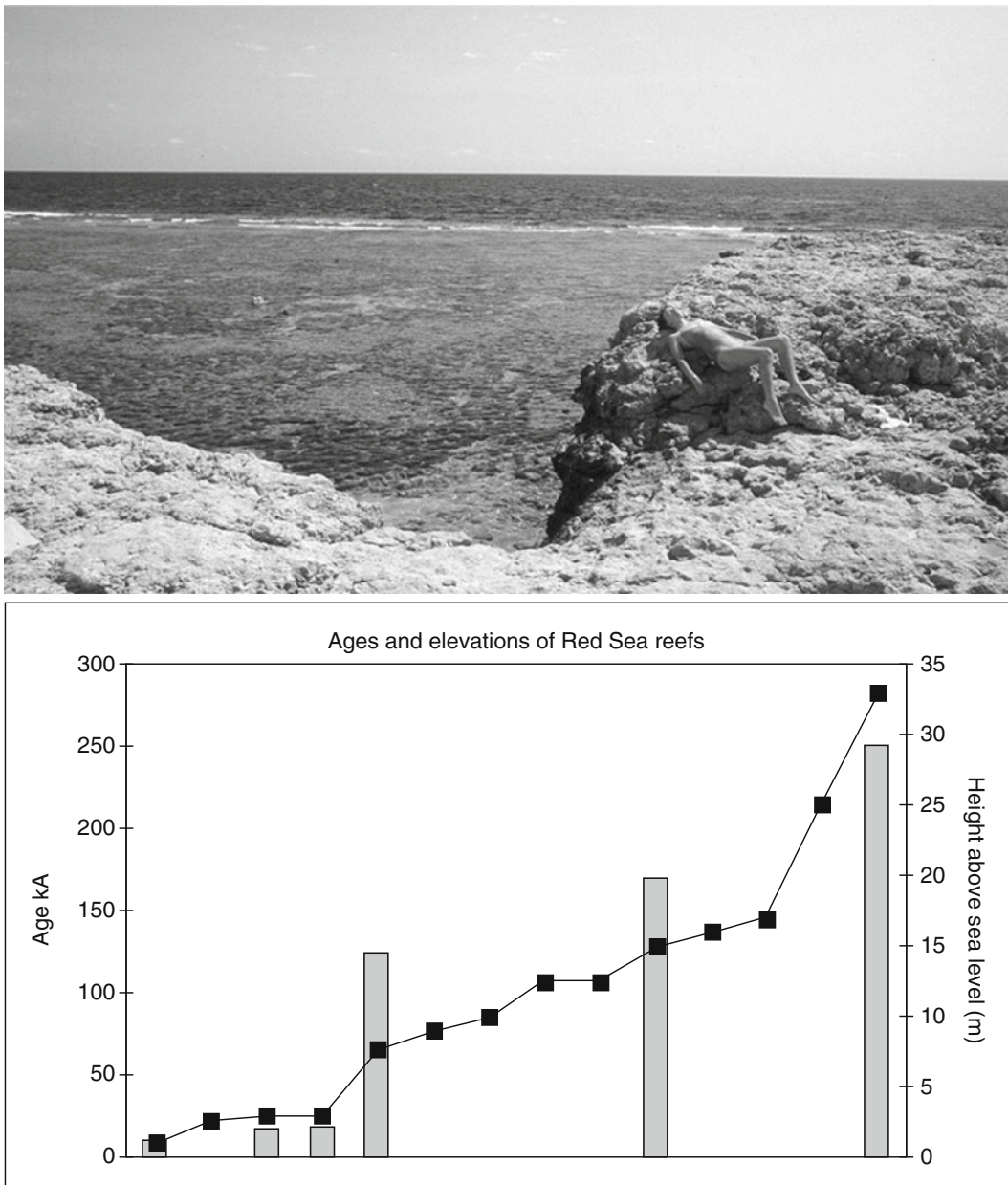
Biological Systems

Biological systems supported along reef shores are usually rich and diverse. Apart from the productivity and biodiversity of the reefs themselves, any areas of hard substrate between the reef and shore are likely to be colonized extensively by corals. Sand accumulates along these shores from wave pumping over the reefs – mostly the sand being limestone, derived from eroding reef material, and fragments of coral and coralline algae. Soft substrates commonly dominate the shoreward coastal marine environment. If stable enough, seagrasses may thrive, themselves being a rich habitat supporting high productivity and a considerably greater biodiversity than adjacent, uncolonized sand. Once these angiosperms take root, they stabilize the habitat further.

Along the coastline itself in such areas, mangroves may dominate the intertidal habitat, particularly if estuarine or at least slightly lowered salinity conditions prevail. These forests may be extensive and grade inland into true terrestrial forests, or may be very shallow-rooted in very limited sediment (Fig. 2) and are known as “hard substrate mangroves.” Most mangrove trees have roots and shoots which trap and stabilize sediments. This can also lead to progradation of the coastline to seaward as well as laterally alongshore. The existence of the mangroves, as with seagrasses, commonly is the direct result of the shelter provided by the reef lying to seaward. Conversely, the existence of the large, shore-colonizing stands of rooted plants, traps and consolidates substantial amounts of sediments that otherwise would reach and adversely affect the reefs, which are extremely intolerant of sedimentation. Thus, complex feedback processes determine both the physical and biological nature of the coral reef coastlines, as well as their physical extent. Seaward progradation may attain a balance, depending on the extent of shallow substrate on which the reef exists and over which it may extend, but rates of progradation are fairly slow; considerably slower than may be the case, for example, where mangroves extend the land seaward near heavily sedimented estuaries in the absence of reefs.

Coral island Shores

These are a special case in which reefs themselves are the only component of the coastline. The modern component may be only a veneer on pre-Holocene reefs, or may themselves



Coral Reef Coasts, Fig. 1 (Top) A central Red Sea fringing reef, whose reef flat is at low tide level. The edge of the reef flat is seen <100 m to seaward, following which the slope dips steeply to hundreds of meters. The person is lying on a tectonically raised fossil reef whose

elevation is about 4 m. (Below) Plots of ages of reefs and their heights above sea level. (Data taken from review material in Sheppard et al. (1992), Table 1.1)

develop into thick deposits. Their shallow existence traps sediments, then limited vegetation, birds and further vegetation, in well-documented patterns of biological succession. Known as cays, keys, or islets of atolls, globally their total coastline is considerable, especially in deep oceanic regions distant from continental coasts. In these, all hard substrate is limestone, and all soft substrate is derived limestone sand. Most of the latter will undoubtedly be modern, though where hard substrate exists, outcrops of pre-Holocene material may be expected.

Typically, these coastlines have classical profiles of a low island, followed by a reef flat at low tide level, and a reef slope which dips to seaward, smoothly or in cascades. These reef slopes may be very narrow, less than 20 m wide, this distance in extreme cases being all that separates an island shore from water which rapidly attains depths of several hundred meters. More commonly, reef flats extend for upwards of 100 m, and may be dry or become extremely shallow for up to about a quarter of the tidal cycle. In exposed areas these may be bordered by an algal ridge; a construction of calcareous red

Coral Reef Coasts, Fig. 2

Fringing reef backed by “hard substrate” mangroves and, on the shallow reef flat, dark patches denote areas of rich seagrasses. The three habitats of reef, mangrove, and seagrasses are closely linked. (Location is in the Sinai peninsula and these mangroves are amongst the most northerly in the world)



Coral Reef Coasts, Fig. 3 The spur and groove structure along the seaward edge of an oceanic atoll. Formed from growth of calcareous algae (*Porolithon* spp.) it is the primary defence against erosion for the reef and the shore behind (Chagos atolls, Indian Ocean)



algae which faces the oncoming waves (Fig. 3). Usually algal spurs extend more or less regularly to seaward from the ridges for a further 100 m, dipping to 5–10 m depth. These structures have a tremendous wave-baffling role, in that their length and spacing appears to be “tuned” in such a way that incoming waves meet the outgoing swash from the preceding wave, their energies partly cancelling each other out. It appears that the very existence of many coral reef shores depends to some degree on this algal, biogenic coastline structure.

Human Habitation on Coral Reef Coastlines

Coral reef coastlines have always been populated, and because reefs and the other rich habitats along these sheltered shores are rich in food sources, notably fish protein, limitations on human numbers have commonly been due more to the availability of freshwater rather than to the supply of food (Sheppard 2000). In rainy tropical areas, human populations have been considerable for long periods of time, while in arid

locations, such as the Middle East, there usually have been far fewer – and more nomadic – people despite the richness of marine foods.

As well as the limiting effects of little or highly seasonal rainfall, the permeable nature of these fossil coral and modern coral island coastlines have long affected the amount of water which may be stored and extracted, and this also has greatly affected the ability of people to live on them. Freshwater lenses on modern coral islands are small and clearly change rapidly in size; they track the quantity of rainfall and they are clearly tidally linked, rising and falling diurnally, often with phase lags of only a very few minutes. They are thus very sensitive to over-abstraction. The available freshwater is also sensitive to increased runoff to sea which takes place following development of human habitation; here significant proportions of the land become covered with built-up areas which are impermeable, causing water to runoff rather than enter the water table. Both the latter and over-abstraction readily cause a collapse of the water table – one marked example is seen in the current sulfurous and saline nature of the water table on Malé, the very built-up and now artificially shaped capital town (and island) of the Republic of the Maldives (Fig. 4). Interestingly, it has also been suggested that where the water table seeps laterally to sea in the region of the intertidal zone in carbonate shores, chemical changes, coupled with microbiological growth, causes solidification of “beachrock” a durable form of limestone which also contributes to coast line stability (Fig. 5).

Where freshwater and biological systems are not over-stretched, human populations may continue to live. The productivity of all three major habitats (reefs, seagrasses, and mangroves) are amongst the most productive in the world (Sheppard 1995) and, in addition, offshore production in

these areas may also yield substantial additional fish such as tuna.

However, numerous factors currently combine to create unsustainable conditions which have led to numerous system collapses and abandonment of coastal villages throughout tropical Asia, Africa, and Central and South America. Pollution from sewage and industries, unsustainable methods of fishing (e.g., dynamite and DDT “bombs”), unsustainable amounts of previously sustainable types of fishing, extraction of coral reef material for building, and infilling of the shallows between reefs and shore to create more land, have all caused ecosystem collapses in coral reef coastlines and have spelled disaster for the local inhabitants. Currently, nearly two-thirds of reefs have been damaged (Bryant et al. 1998) though this figure is nearer 80% in areas such as Southeast Asia. Similarly, removal of the mangroves for timber or to clear space for unsustainable shrimp farming, and whose existence protected the seaward reefs by trapping terrestrial sediments, permits previously stabilized sediments to blanket and kill the reefs leading, eventually, to coastline erosion. Even poor agricultural practices far inland have resulted in massive movement of topsoil into these coasts via river flow, again causing mass mortality of habitats and starvation of people (Sheppard 2000). Compounding the damage is the fact that such reef-sheltered coastlines tend to be disproportionately rich in their provision of breeding, spawning, and nursery areas for marine species.

Effects of Climate Change and Sea-Level Rise

Although reefs grow vertically to the low tide level, and did so during the rapid and substantial Holocene transgression (the

Coral Reef Coasts, Fig. 4 The capital city and island of Male, Republic of Maldives; a coral reef island city. The reef flat is almost entirely infilled. A breakwater (on left) was built by Japanese aid following devastating surges and loss of land. Water abstraction has left the water table damaged, now with saline and sulfurous water



Coral Reef Coasts, Fig. 5 Beach rock formation. Where freshwater seeps into the intertidal, microbial growth leads to solidification of the fine sands. These appear to be extremely important in the creation of dipped limestone slabs in some areas (Chagos atolls, Indian Ocean)



sea-level rise of about 150 m in about 8,000 year), their ability to continue to do so under present conditions of rising sea level and temperature is not at all certain. First, sea surface temperature increases and El Niño events in recent decades have caused mass bleaching and mortality of corals. In 1998 in the Indian Ocean, for example, coral mortality in shallow reef areas exceeded 80% in many and probably most atolls, and in many patch reefs and fringing coral reefs in the Arabian Gulf, southern Red Sea and many continental shores. As of mid-2000, recovery appears poor. It is known that coral erosion continues, and since growth and erosion in a healthy reef are closely balanced (in favor of growth) net erosion may now be predicted with concomitant reduction of breakwater effect. The importance of the breakwater effect on reefs is marked, and can be seen when reefs have been removed previously by other reasons: for example, by coral “mining” on reefs around islands which have lowered the surface of the reef by 0.5 m, or from mortality of corals through disease which resulted in similar loss; in such cases the loss of the reef has led to severe erosion of the coastline, sometimes exceeding 1 m per year until a new stable point is reached. Given the attractiveness of development on “reef front” locations, infrastructure damage has often been substantial. Both of these forms of reef loss are likely to mimic the loss following recent episodes of temperature-induced mortality.

It is likewise known that coral growth and reef growth become “decoupled” as environmental conditions become increasingly severe (Sheppard et al. 1992), and that even continued coral growth does not necessarily result in continued growth of the essential hard, durable reef. Further, while it is generally understood that reef growth results in deposition

of vast quantities of carbonate, a point important in the consideration of possible carbon sinks to combat rising global CO₂ levels, in fact a rising atmospheric CO₂ level also alters water chemistry dynamics in the direction of *reducing* deposition of carbonate; estimates of a 21% reduction of carbonate deposition by 2065 have been postulated given Intergovernmental Panel on Climate Change estimates of CO₂ rise (Leclercq et al. 2000). While the latter work has primarily been directed at the question of the global carbon sink, clearly reduced carbonate deposition on reefs which protect coastlines will become increasingly important too, as the fine balance between reef accretion and erosion becomes disturbed in favor of erosion.

Thus, while healthy reefs may continue to keep pace with a rise in sea level, as has been the case before, the increasingly stressed and damaged reefs appear to be decreasingly capable of fixing CO₂, have a decreasing ability to maintain rich coral- and reef-forming (limestone-depositing) communities, and possibly even a decreased ability to maintain food production for human populations. Evidence is accumulating to show that pressures on reefs, including over-fishing, lead to changes to their ecological stable states which cause them to be dominated by different species groups, even if or when the pressure initially causing the change is removed. The replacement biotic groups may be dominated by non-calcareous species, and the trend appears to be one-way, like a ratchet rather than a wheel. If confirmed further, substantial changes to coastlines currently protected by reefs are likely to occur, the direction of the changes being characterized by increased erosion, reduced productivity, and reduced ability of these coasts to support the human populations currently living there and dependent upon them.

Cross-References

- [Algal Rims](#)
- [Beachrock](#)
- [Bioherms and Biostromes](#)
- [Coral Reef Islands](#)
- [Coral Reefs](#)
- [Coral Reefs, Emerged](#)
- [Demography of Coastal Populations](#)
- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Mangroves, Ecology](#)
- [Water Quality](#)

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Coral Reef Islands

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A coral reef island is composed of rocks from coral skeletons, that is, biologically formed calcium carbonate materials derived from the adjacent coral reef and raised above sea level. Coral reef island sizes range from a few square meters to many square kilometers, and they come in all shapes and proportions. Their soils consist of coral fragments, calcareous algae and other limestone detritus, varied amounts of humus, guano from sea birds, volcanic ash, and drifted pumice (Fosberg 1976).

Coral reef islands can develop only where suitable conditions sustain coral growth over time. These conditions include favorable physical factors (high temperature, high salinity, good light penetration, and low nutrients) and biological

factors (especially a supply of coral larvae) in tropical regions that provide a firm substrate in the shallow photic zone (Birkeland 1997). Coral islands are created in three ways, which are not mutually exclusive: (1) accumulation of dead coral reef rubble and sediments on top of a reef flat through wave and storm action, resulting in a low island (several meters in height) that may be ephemeral or stable; (2) emergence of a coral reef due to a drop in relative sea level; and (3) uplift of the ocean floor beneath the reef. Coral reef islands resulting from tectonic uplift of coral substrate may range from several to hundreds of meters in height.

The life span of coral reef islands is determined by the balance of erosional and depositional processes, and by sea-level rise. Coral islands drown when their corals are submerged below a critical depth for coral growth (Grigg and Epp 1989). During their existence, these nonvolcanic islands may become vegetated and support a terrestrial fauna and flora. Coral reef islands are fundamentally different from so-called “coral islands” that consist of fringing coral reefs growing on a foundation provided by an island of volcanic or continental origin. Coral reef islands may be associated with volcanic or continental land masses if they form as the result of depositional processes arising from fringing or barrier reefs. These islands may lie on barrier reefs separated by a lagoon of varying widths from high land, and their proximity to this land may result in some terrigenous sediment input to coral reef islands.

Darwin's theory of reef formation asserts that the three main reef classes, fringing, barrier, and atoll, are each different stages of development of a coral reef (Darwin 1842). Darwin did not define a coral reef island, simply referring to the islands on atolls as islets of various sizes and shapes. Dana (1872) defined “coral islands” as reefs that stand isolated in the ocean (= atolls), away from the shores of high islands and continents which harbor “coral reefs.” His definition is clearly too restrictive, although he clarified it by stating that a coral island has a lagoon in the middle instead of a mountainous island and consists of a very low ratio of habitable land to area. Other terms associated with coral islands also have different local meanings. Extremely small coral islands on atolls are called islets or cays, while others refer to small coastal islands as cays (“keys” in Florida), some of which are not derived from coral reefs. “Coral cay” defines a small low coral reef island. “High” and “low” islands are terms often applied to coral reef islands. High islands are generally not coral reef islands since most consist of volcanic or continental materials with surrounding fringing or barrier reefs. However, uplifted coral reefs may attain a height similar to some high islands. Low islands are true coral reef islands if their sedimentary origin is coral reef material.

Classification of Coral Reef Islands

Stoddart and Steers (1977) placed coral islands into seven different classes based on morphology and sediment characteristics, with vegetation used as a secondary diagnostic. Vegetation patterns depend on elevation, surface features, area, distance from sea, soil development, and human activities (Fosberg 1976). Stoddart and Steers (1977) stress that these islands do not necessarily evolve from one class to another; any change results from external influences, primarily storms. The first class includes unvegetated and vegetated sand cays, which are low-lying accumulations of coral rubble and sand deposited by wave refraction on the reef. Sand cays are generally associated with barrier reefs and patch reefs in protected areas of reefs, for example, in lagoons behind barrier reefs. They occur on reef patches in shallow, wide lagoons and assume a variety of shapes depending on sea level, wave energy, and tidal ranges (Hopley 1982). The windward margins of the cays tend to be composed of loose coarse rubble and the leeward sides accumulate sandy sediments. These terrestrial piles of coral material may eventually develop vegetation, mostly of plants from seeds deposited by birds or by ocean currents. A lagoon may form in between, with mangroves helping to stabilize the sediments (Guilcher 1988). These islands range in maximum dimension from several meters to greater than 1000 m, with a height of several meters. Unvegetated islands are lower in height and ephemeral in nature. The second class of coral island is the *motu*, which is the most common island in the Indo-Pacific. Motus have a seaward (windward) shingle ridge 3–5 m high and a leeward sandy shore and are generally vegetated. These islands occur most frequently on the exposed windward sides of atolls on the inner reef flat on Pacific atolls where sediments and rubble are deposited by strong unidirectional waves. Motus vary in number, size, and shape on a given reef or atoll. Islands in the Tuamotu atolls range from 250 to 300 per atoll, covering 30–35% of the atoll rim. Most atolls in the Marshalls, Gilberts, and Ellice groups contain more than 20 islands, while the Carolines support fewer than 10 islands per atoll (Stoddart and Steers 1977). The number of islands on an atoll depends on the number of erosional channels (“*hoa*”) cut through formerly continuous atoll rims. The vegetation on motus is strongly influenced by elevation above sea level and island size. A freshwater lens can develop on larger islands which will support a more diverse flora and fauna. The third and fourth classes of coral islands are the shingle cay and mangrove cay. The shingle cay is a storm-related event not common on coral reefs, while mangrove cays are formed by colonization of shoal areas by mangroves. Mangrove cays with a windward sand ridge form the fifth class, and low-wooded islands are the sixth class. The seventh class consists of emerged coral reefs (limestone islands). Examples of emerged coral islands include the northern

Florida Keys, the Bahamas, and Bermuda in the Atlantic ocean and Aldabra atoll in the Pacific ocean. These are often dissected into karst topography with steep and strongly undercut shores, sharp peaks, and sinkholes.

Distribution of Coral Reef Islands

Most coral reef islands occur in the Indo-Pacific region. There are over 300 atolls and extensive barrier reefs in the Pacific ocean and only 10 atolls and 2 barrier reefs in the Caribbean region (Milliman 1973). The total number of coral reef islands is unknown, and varies according to change in sea level and storm activity.

Summary

Coral reef islands occur throughout the world at tropical latitudes. They are best viewed as products of the reef and as an extension of the reef ecosystem (Heatwole 1976). Ecological studies are needed to determine the physical and biological linkages between the coral reef and its islands. Although most coral islands occur in isolated areas, human influences on the vegetation and fauna are highly evident and easily quantified. These land forms are excellent geological and ecological systems for studying interactions between marine and terrestrial ecosystems.

Cross-References

- Atolls
- Cays
- Coral Reefs
- Coral Reefs, Emerged
- Small Islands

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Coral Reefs

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Coral reefs are the largest structures built solely by plants and animals and many are clearly visible from space. Gross distribution is limited by a requirement for warm water with an average winter maximum temperature of 18 °C generally regarded as providing the poleward margin for coral reefs which are thus found only in a few areas outside the tropics. High temperatures above 30 °C may also be lethal to corals. Other requirements include good water circulation and a generally low nutrient status for ambient waters. In the open ocean where nutrient availability is very low, coral reefs exhibit a very efficient nutrient recycling, the marine equivalent of tropical rainforests. Inshore reefs can tolerate higher nutrient levels and may even respond with more rapid growth rates, though this may be countered by the skeleton laid down being more fragile. Turbid waters have a detrimental effect, limiting the penetration of light in the water column and compressing the vertical zonation of the coral reef. In clear open waters corals may grow to depths exceeding 110 m. In turbid water this may be limited to 10 m or less.

The Importance of Plants on the Reef

Many of the restrictions on coral growth are imposed not by the corals themselves but by plants. Within the endodermal tissue of the coral polyps are unicellular algae, zooxanthellae, which have a close symbiotic relationship with the coral host. The zooxanthellae not only provide some of the nutrition for the coral but also aid as important primary producers of oxygen and carbon during daylight hours through the process of photosynthesis. They are also nutrient conservators, satisfying their own requirements from the host metabolic waste which they thus assist in removing. Most importantly, they are direct promoters of inorganic growth of coral skeletons. In daylight, the algae produce more oxygen than can be used by

polyp respiration and some of the carbon dioxide produced by the respiratory process is refixed by the algae into new organic matter. As photosynthesis takes up C-12 slightly faster than C-13 the organic matter synthesized by the zooxanthellae has a relative preponderance of C-12 and a pool of carbon compounds enriched in C-13 is left behind. It is from these that the calcium carbonate skeleton is built, aided by pH variations associated with the diurnal photosynthetic process. Corals grow 2–14 times more rapidly in light as opposed to darkness and twice as fast on a sunny day as opposed to a cloudy day.

Other algae also participate in the growth and maintenance of the coral reef. Crustose coralline algae infill and cement the coral structure especially on the high energy margins of the reef where they build distinctive pavements and ridges. *Halimeda* producing delicate plates of calcium carbonate can be extremely prolific in producing sediment and, given appropriate nutrient pulsing, can build bioherms as significant as the coral reefs (as for example in the northern Great Barrier Reef, and the Makassar Straits of Indonesia). At their poleward limits competition from macro algae which can over-shade the corals, is a factor in coral reef distribution.

Coral Reef Diversity and Ecology

Coral reefs are amongst the most complex and diverse ecosystems on earth (Dubinsky 1990). A variety of marine invertebrates and plants contribute to the accumulation of the limestone which are the most massive bioconstructional features in the geological record. Symbiosis with zooxanthellae occurs not only in the corals but in other major calcifying organisms including giant clams and foraminifera which are a major producer of reef sediment. Other important calcifiers include bryozoans, sclerosponges, echinoderms, and a wide variety of molluscs. They contribute to both the reef framework and the infilling sediments. At the scale of the total reef ecosystem gross primary organic production is very high but conversely, because of the amount of recycling which occurs, net productivity may be close to zero. In contrast, calcification is the most active of any ecosystem on earth. With an estimated total area of 617,000 km² the estimated total annual production is estimated at 929 million tonnes CaCO₃.

The stony scleractinian corals are the most important of the reef biota and their diversity typifies that of many other reef-dwelling species including coral reef fish. It is currently estimated that there are about 220 coral genera and at least 1,300 species, though many are rare and/or azooxanthellate. However, there are two distinct centers of diversity. The most diverse is the Indo-Pacific centered on the south-east Asian archipelago, most significantly the Molucca and Banda Seas of eastern Indonesia. Here there are about 450 species belonging to 80 genera. Individual sites may have as many as

140 species. The other center is in the western Atlantic and is much less rich with only 20 genera and 50 species.

The reasons for this lie in the evolutionary history of the scleractinia which evolved from the earlier rugose corals of the Paleozoic (Veron 1995). The first scleractinians appeared in the middle Triassic about 250 million year ago in the western Tethys Sea approximating to southern Europe and the Mediterranean. Corals occupied a circumglobal seaway which included the Tethys Sea and the gap between the two American continents. During the Mesozoic and Tertiary, as the African, Indian, and Eurasian plates collided the Tethys Sea was closed from the west, its diverse coral fauna pushed eastwards into what is now the final relict consisting of the islands and peninsulas of south-east Asia. The Atlantic fauna was isolated first from the east, then in the Pliocene from the west as the Isthmus of Panama isolated the Atlantic from the Pacific.

The gross pattern of diversity is thus a result of geological history. However, local diversity is also dependent on the range of ecological niches available (particularly wide-ranging in south-east Asia) and the frequency of disturbances such as hurricanes, earthquakes, volcanic eruptions, or crown of thorns starfish outbreaks (Connell 1997). As coral reproduce mainly through a single annual spawning event, a medium frequency of disturbance can allow planula larval recruitment from a wide range of ecological niches because of the longevity of the larval phase which can be several weeks, allowing dispersal over hundreds of kilometers.

Rates of Growth and Accretion and the Importance of Bioerosion

The majority of corals are colonial organisms and annual extension rates for single colonies may be more than 10 cm/year for branching species and 1 cm/year for massive corals. Each year's increment consists of a wider high- and narrower low-density band which can be clearly identified by x-radiography. The high-density band is normally laid down during a period of stress imposed by either ambient environmental conditions or reproduction.

However, as the corals and other organisms are laying down calcium carbonate, many others are bioeroding the reef substrate. Some of the sediment produced may help to infill the reef framework but much is lost from the immediate vicinity of the reef. Bioeroding organisms include rock grazers in search of epilithic and endolithic blue-green algae on the reef surface, rock browsers which consume reef fabric and sediments such as parrot fish and holothurians, and a wide range of burrowers and borers. Rate of breakdown can match or exceed the rate of calcium carbonate production. For example, individual chitons can consume up to 13 cm³/year, grazing fish in Barbados reefs produce 1–2 tonnes/ha/year of

sediment, clionid boring sponges 5–10 tonnes/ha/year, and grazing echinoids up to 97 tonnes ha/year, 65% of which is lost from the reef. Boring worms such as Sipunculids and Polychaetes produce up to 20% of the porosity in the reef framework. On Rongelap atoll in the Pacific it is estimated that up to 100,000 tonnes of lagoonal sediment passes through the acid gut of grazing holothurians each year.

Coral reefs which are under stress may have bioerosion rates which exceed the current rate of limestone deposition. However, most healthy reefs are actively accreting. Rates of reef growth are commonly calculated in two different but complementary ways. Biogeochemical measurements of changes to water chemistry over a coral reef can give instantaneous calcification rates related to total reef metabolism (Kinsey 1985). Such measurements have been made worldwide and when converted to annual rates give remarkably uniform results for the different reef environments. All reefs accrete through the mix of three absolute modes of calcification:

100%	Coral cover	10 kg	CaCO ₃ ² /year
100%	Algal pavement	4 kg	CaCO ₃ ² /year
100%	Sand and rubble	1 kg	CaCO ₃ ² /year

Different zones of the reef are thus accreting at different rates. Sediment is produced from the highly active zones such as the reef front, to the less-active parts of the reef such as lagoons. Taking into consideration the density of the calcium carbonate produced (2.9 for aragonite) and the porosity of coral and reef frameworks (about 50%) these results give an upper limit of about 7–9 mm/year for reef accretion.

Using the geological record through drilling and radiocarbon dating of reef cores provides an alternative long-term method for calculating coral reef accretion rates, which are nonetheless very compatible with those from reef metabolism. The synthesized results from the Great Barrier Reef (Davies and Hopley 1983) have been reproduced from many other reef-drilling projects. Framework accretion rates ranged from 1 to 16 mm/year according to the density of the fabric (lowest for algal pavement, highest for open branching coral frameworks) though the modal rate of 7–8 mm/year is identical to that from water chemistry. Drilling has also produced rates for detrital sedimentation. In lagoons and sand aprons it is 1–4 mm/year for sand flats, 7–9 mm/year for steady trade wind accumulation and 13–18 mm/year for high-energy low-frequency events.

Coral Reefs and Sea-Level Change

Coral reef growth is highly constrained by sea level. Uppermost growth may exceed the mean low water mark in moated pools in reef flats and will die off with any fall in relative sea

level. Single tectonic events causing uplift of the land can cause mortality across an entire reef flat and in areas of constant uplift in the tropics, such as at tectonic plate margins, suites of raised reef terraces may occur to several hundreds of meters of elevation. Radiometric dating of such terraces in locations such as the Huon Peninsula in Papua New Guinea and Barbados has determined not only the rates of uplift locally but also global eustatic sea-level records.

As coral reefs have existed for such a long period, global changes in sea level have had a major impact. Global cooling during the late Tertiary and Quaternary, resulting in waxing and waning of ice sheets at the poles and in the northern hemisphere have resulted in falls in sea level of up to 150 m, exposing all extant coral reefs and removing active growth to deeper banks and slopes perhaps severely restricting the area of coral growth at the low sea level times. Fringing reefs were exposed as limestone terraces, shelf barrier reefs as chains of limestone hills, and atolls and other oceanic reefs as limestone islands. The records of sea-level change suggests that over the last million years coral reefs have been exposed for far greater a period than they have been inundated. Reef accretion has been limited to the relatively short high sea-level phases of interglacials.

During periods of exposure, which may have been as much as 100,000 year at a time, the reefs were subjected to karstic erosion processes. Indeed a series of models for the development of coral reef morphology at both the gross and detailed scale, have been hypothesized based on the karst land forms which were etched at low sea-level phases (Purdy 1974; Hopley 1982). This antecedent karst hypothesis has been used to explain the annular shape of atolls and the rim structures of barrier reefs as well as the labyrinthine morphology of many reef surfaces (Fig. 1) which have been likened to the complex ridges and enclosed depressions of tropical karst regions. Some features such as caves with speleothems, “blue holes” up to 500 m wide and 100 m deep are clearly derived from preexisting karst land forms. Blue holes on the Great Barrier Reef for example (Fig. 2) are almost certainly collapse dolines over which a veneer of 10–20 m of Holocene reef growth has occurred (Hopley 1982).

A problem with the antecedent karst hypothesis is the time available during low stands of sea level for major karst landforms to develop (see Hopley 1982, chapter 7 for discussion). Recent studies have shown that some morphology previously attributed to karst interaction has in fact been derived from reef growth in the Holocene. However, it is also possible that the growth which occurs during an interglacial high sea level may enhance that produced by erosion in the previous period of exposure.

This is then further enhanced by the patterns of erosion during the next low sea-level stand. In all probability both growth and erosion produce the complex relief of coral reef systems.



Coral Reefs, Fig. 1 Complex lagoonal relief in the large reefs of the Pompey Complex, Great Barrier Reef Australia

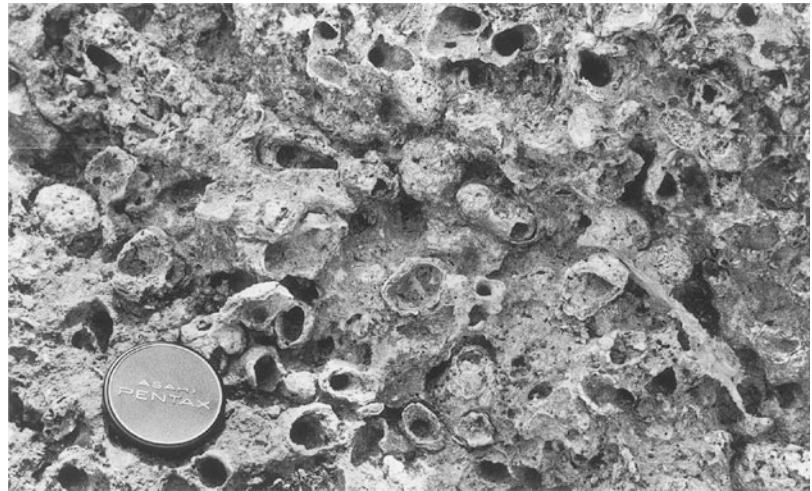


Coral Reefs, Fig. 2 A blue hole approximately 300 m wide and 33 m deep in Molar Reef, Great Barrier Reef, thought to have originated from a collapsed doline

In addition to the formation of karst landforms during low sea-level exposure, coral reefs also experience diagenetic changes. The original skeletal limestone is largely aragonite (corals, molluscs, *Halimeda* sclerosponges) or high Mg-calcite (crustose coralline algae) which in subaerial and vadose environments is unstable and may be replaced by a coarser crystalline low Mg-calcite (Fig. 3). Caliche crusts (Fig. 4) may form on the surface and soils can be retained over parts of the surface even after reinundation. During the subsequent high sea-level phase the older reef is recolonized and new framework deposited over the older surface. However, it remains as a clearly identifiable discontinuity termed a “solution unconformity.” Beneath the Holocene reef framework the solution unconformity formed during the last glaciation is typically 10–20 m beneath the surface. Where coral reefs have been drilled to greater depth 10 or more solution unconformities have been identified especially in atolls and shelf edge barrier reefs which suggests continuous subsidence at these sites during the period of reef accretion.

Coral Reefs, Fig. 3

Diagenetically altered *Acropora cervicornis* in the Pleistocene reef tract, Barbados. The aragonite of the original corals has largely disappeared or altered into calcite. Low Mg-calcite molds result from the more rapid replacement of encrusting coralline algae



Coral Reefs, Fig. 4 A brecciated caliche profile with numerous brown micritic stringers developed on Pleistocene reef limestones ca. 84,000 year BP on Barbados

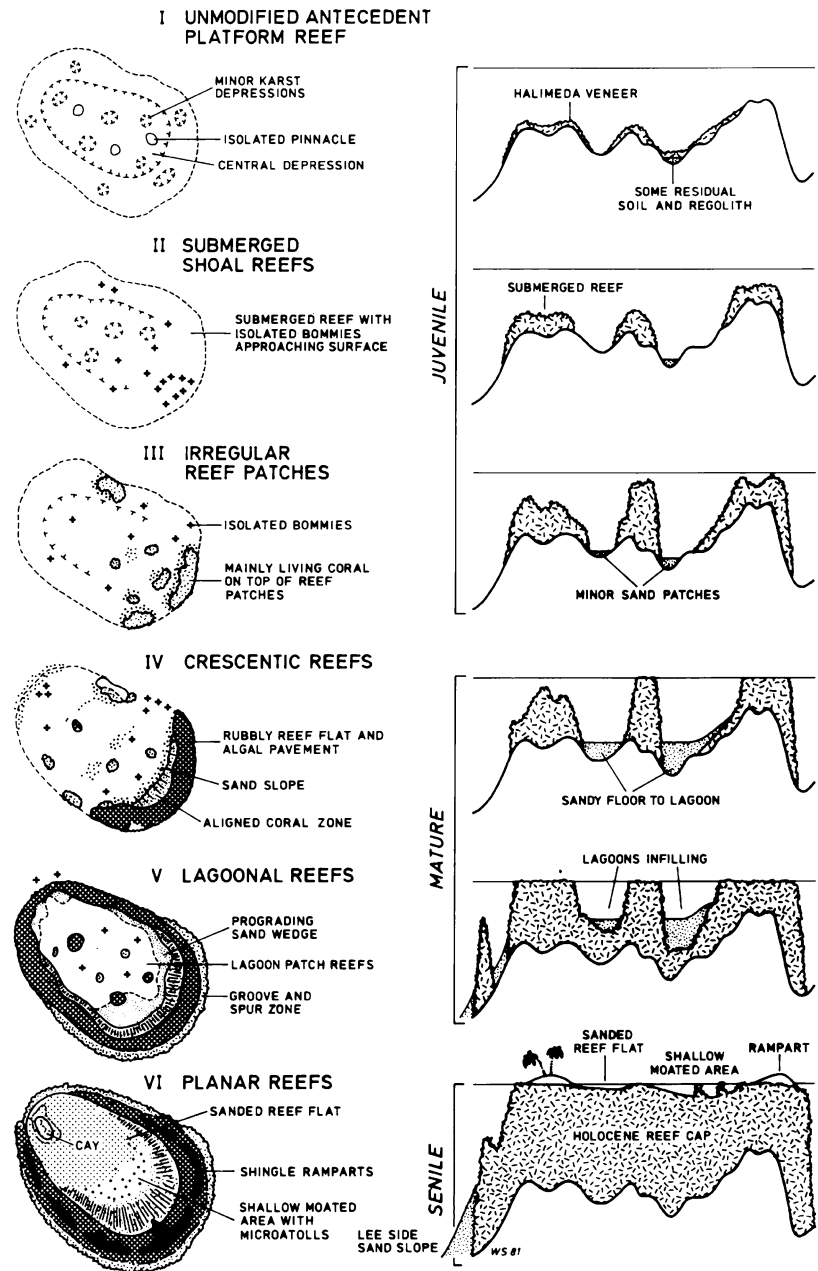
**Reef Growth in the Holocene**

Modern coral reefs have developed entirely during the Holocene period, the majority forming a relatively thin veneer over older Pleistocene foundations which are no more than 10–20 m below modern sea level (Fig. 5). Very few radiocarbon ages greater than 9,000 year have been obtained for this Holocene veneer. Only where reefs were in deep water, such as open ocean atolls, were corals able to merely migrate to the flanks of the reef during the last glacial and thus recolonization was simple and immediate. This is also the situation for some of the midoceanic volcanic islands, which explains why the deepest and longest postglacial record comes from Tahiti. Here the postglacial reef is more than 90 m thick and basal dates older than 13,500 year have been obtained. Australia's Great Barrier Reef in contrast rises from a relatively shallow continental shelf generally less than 50 m deep. Even the lowest parts of the older Pleistocene reefs were inundated less than 11,000 year ago (Hopley 1995). Delays in

re-colonization occurred due to the need for larval recruitment from more distant sources (probably Coral Sea Plateau reefs), and as the result of initially poor water quality as regolith was removed from the Pleistocene foundations and as a result of very close proximity of the mainland of the time to the main reef tract. The major part of the Holocene growth took place only after 8,500 year. BP by which time the mainland coastline was close to its present position and well removed from the outer reef.

On the Great Barrier Reef, and generally many other coral reefs, the major period of Holocene accretion took place between about 8,500 and 6,000 years ago, when average vertical accretion rates of ca. 5 m per thousand years were achieved. Largely for hydroisostatic reasons (see Hopley 1982, chapter 13 for discussion) the pattern of Holocene sea-level rise over this period varied in different parts of the world. Over much of the Indo-West Pacific modern sea-level had been achieved by as early as 6,500 year. BP while in the Atlantic relative sea level may have continued to rise right up

Coral Reefs, Fig. 5 An evolutionary classification of Holocene coral reefs



to the present time, though at a considerably slower rate after 6,000 year. BP. Reefs have responded in different ways with three models recognized (Neumann and Macintyre 1985; Guilcher 1988):

- Keep-up, in which the reef was able to accrete vertically at about the same rate as the rise in sea level and was never more than a few meters below sea level.
- Catch-up, in which the reef lagged behind the sea-level rise due to delays in recolonization and/or slow growth. At times these reefs may have been more than 10 m below sea level.
- Give-up, reefs which after becoming established in the early Holocene died off and remain only as drowned

carbonate banks. Various reasons have been put forward. That they became too deep because of the rapid rise in sea level is unlikely in the majority of examples as most would have remained within the photic zone of active reef growth, unless there were also an increase in turbidity (which would have limited the depth of reef growth). Many “give-up” reefs are on shelf edges and sediment shed from upslope reefs may have been one cause of their demise. In the Caribbean, in particular, it has been hypothesized that as sea level first flushed the shelf, a new wave of sedimentation and eutrophication was the cause of shelf edge reef decline (Macintyre 1988). In addition, at high latitudes

shallow shelf waters may also have killed off reef growth at the shelf edge.

Once coral reefs caught up with sea level then their upward growth phase was terminated and carbonate productivity has been directed into lateral rather than vertical growth.

An Evolutionary Classification of Holocene Reefs

It is possible to examine the growth of modern reefs by way of an evolutionary classification (Hopley 1982), based on the presumption that the majority of reefs are located over pre-Holocene reefal foundations at no great depth and with an irregular morphology resulting from both the original growth form of the last period of reef growth and the erosion which has taken place during subsequent exposure. This presumes transgression by the sea over the foundations and a sea-level rise which most reefs or most parts of reefs cannot match by upward growth (Fig. 5).

Three phases of Holocene reef development are recognized in the classification. In the juvenile phases, initial colonization takes place on the antecedent foundation and upward growth of the reef lags behind sea level rise, enhancing the original relief of the foundation. In the mature phases, reefs reach modern sea level and develop reef flats over the highs in the antecedent surface and especially around the windward reef margins where reef growth appears to be most rapid. One or more lagoons typify this phase but in the later stages of maturity, lateral transport of sediment from the productive margins leads to initial infilling of lagoons, masking of the inherited relief forms, and widening of reef flats. The senile phase is reached when lagoons are infilled completely, and lateral movement of sediment from the windward margins leads to progradation to leeward.

The basic classification is applicable to medium-size reefs growing from antecedent platforms with single depressions. Smaller reefs, where antecedent surfaces may lack a proto-lagoonal relief may develop along similar lines but because the reef flat initially developed around the reefal margins may occupy such a high proportion of the total reef area, any lagoon which forms will be shallow and quickly filled and the mature phases will be either bypassed or be present for only a very short period. In contrast, the largest reefs will be at the mature stages for the longest period because of the low ratio of productive margin to lagoon area. These reefs may also differ from the basic classification during the mature phase by having more than one lagoon.

A major subgroup of reefs are ribbon reefs which form distinctive barriers. They grow from linear foundations which are either discrete antecedent platforms or higher rims on much larger and usually deeper platforms. The basic feature

is depth of water behind the narrow reef flat which develops from the foundations. Although sediment moves to leeward, accretion is very slow and widening of the reef flat is retarded. However, the stages in the development of a ribbon reef can be equated with those in the basic classification.

Development of Coral Reef Zonation

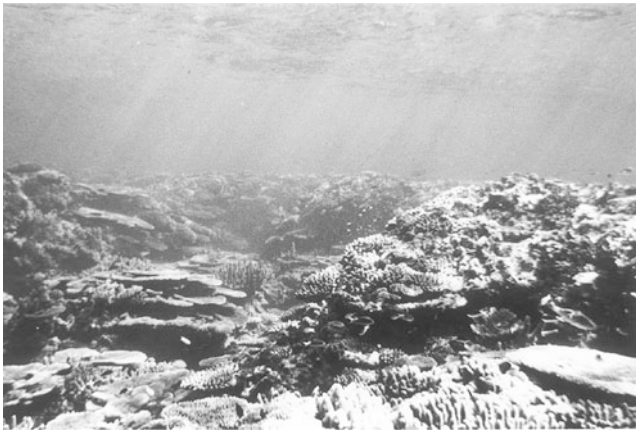
Distinctive morphological and ecological zonation of coral reefs is a characteristic which develops only after the reef has reached sea level and commenced horizontal growth of a reef flat (Hopley 1982, chapter 10). During “catch-up” or even “keep-up” phases, the reef may be too far beneath effective wave base to allow the significant energy gradient from windward to leeward to exist. The reef may be a more homogenous environment of continuous coral cover with the only variable being the attenuation of light on the flanks and the development of more foliaceous light capturing colonial forms. As the reef reaches sea level, however, the contrast between windward and leeward sides is enhanced by sediment movement, accumulation of detrital features, and the colonization of the various ecological niches by distinctive life forms. Zones develop parallel to the reef front, their width and distinctiveness determined by the length of time the reef has been at sea level and the strength of the energy gradient (Fig. 6).

While the species which inhabit the various zones may differ geographically, the morphology of coral reef zones is similar worldwide and typically may consist of:

Fore-reef slope: Increasing light and wave energy are the controlling parameters on the deeper fore reef. At greater depth are flat, platey, and encrusting light capturing corals. Higher up the slope where light is not limiting branching forms may dominate. Moving into the higher energy wave



Coral Reefs, Fig. 6 The windward margin of a mid-shelf reef, Great Barrier Reef showing outer living coral zone with buttresses, algal pavement with isolated reef blocks, aligned and nonaligned coral zones



Coral Reefs, Fig. 7 Reef front dominated by *Acropora* sp. close to wave base. Mainly corymbose are found on the higher crests of the spurs with arborescent and plate-like forms within the deeper channel

zone at about 10 or 15 m, more compact or corymbose forms become dominant and coral cover is dense (Fig. 7).

Reef crest: Particularly on high-energy reefs the crest develops massive spur and groove morphology. Originally thought to be inherited from karst rillen type features developed at low sea levels, most drilling investigations have indicated a growth origin for spur and grooves. They have developed as a baffle against the wave energy and are self-perpetuating as the grooves act as surge channels along which sediment is swept. **Outer reef flat:** The coral covered tops of the spurs are continued onto the outermost reef flat as a living coral zone of encrusting and resistant growth forms up to about mean low water mark.

Algal crest: Particularly on very high energy mid-oceanic atoll reefs a rim built by crustose coralline algae may rise to almost mean sea level. On lower energy reefs the feature may be far less prominent consisting of a rubble pavement cemented by the coralline algae. Under even lower energy conditions turf-like algae may dominate in this zone.

Detrital zone: Beyond the algal pavement, and often deposited over it may be a morphologically diverse zone characterized by deposition of coarse material derived from the reef front. This zone can consist of imbricated clasts of branching corals forming low irregular banks parallel to the reef front. Where tropical storms are relatively frequent a more prominent ridge or rampart may develop and, if stable for sufficient time, its lower sections will become cemented. Also in this zone, and on the reef flat to seawards may be much larger boulders and reef blocks the largest of which may be more than 10 m in diameter. These are sections of the reef front broken off in major storms or by tsunami waves.

The inner reef flat: Behind the energy-absorbing outer zones, the inner reef flat may be another zone of dense coral growth. Where wave set up provides a steady flow of water over the reef crest this zone may display an alignment of

corals normal to the reef front. Elsewhere, at low tide pools may be moated by the rubble ridges or the algal crest and massive corals will be in the form of micro atolls.

Lagoons and back reef zones: In the deeper water of lagoons or back-reef, finer sand size sediment, which is swept across the reef flat, forms prograding sand sheets which may slowly bury lagoonal or leeward side patch reefs. This sand sheet may be up to 500 m wide and 10 m in thickness and is where much of the carbonate productivity of the outer reef is finally deposited. Beyond this patch reefs of various forms may rise from the sandy lagoonal or back reef floor. Each patch reef may have its own "reef flat" surrounded by an algal rim. In shallower lagoons reticulate reefs forming a network of anastomosing ridges may form an intricate pattern of reef growth.

Leeward reef slope: The high energy wave zone is lacking from the leeward side of the reef and coral growth even near the surface may be in the form of branching thickets. If the leeward slope descends into deep enough water, at greater depth more light seeking platey forms may again dominate.

Fringing Reefs

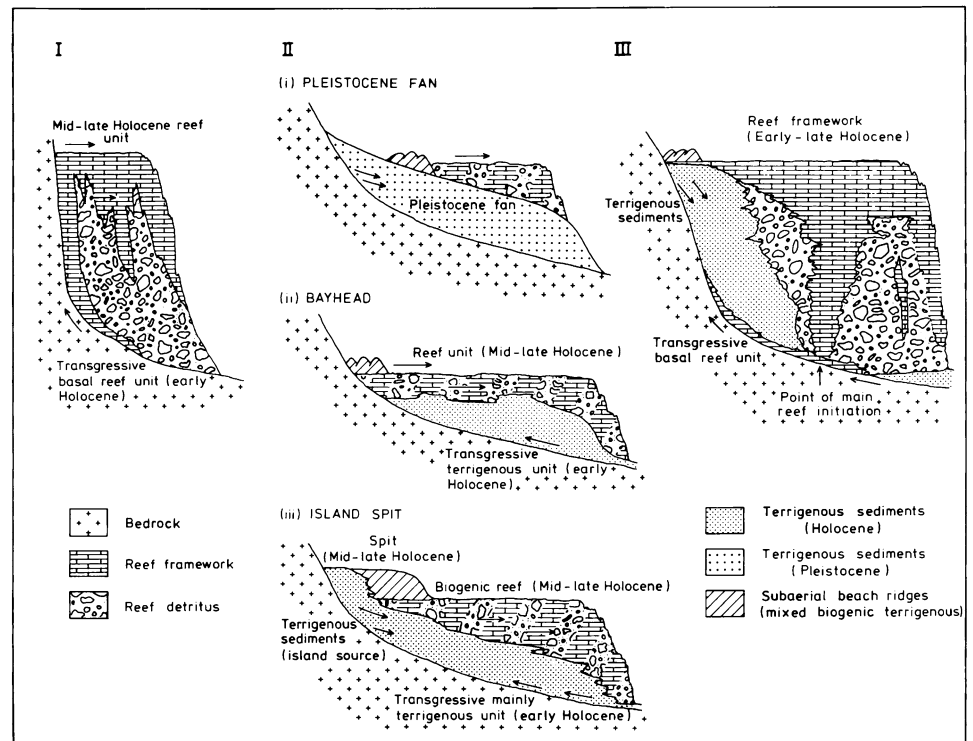
Fringing reefs, attached to the mainland or to a high island, may have the weakest zonal development, especially where sheltered by offshore reefs or other islands. They are the commonest reef type with an area exceeding 320,000 km² or more than 50% of total reef area. Many grow in turbid waters but can still flourish as long as sediment does not settle and smother the corals. However, because of the greater attenuation of light the vertical zonation of the reef front may be compressed into less than 10 m water depth.

Fringing reefs were formerly considered to be the simplest of reef forms, growing outwards from the mainland or island shoreline as originally advocated by Charles Darwin over 150 year ago. However, over the last 20 year a considerable amount of drilling and dating of fringing reefs has been undertaken and they have been shown to have a complexity which matches that of other reef types.

Because they have a foundation on which to grow immediately sea level stabilizes (the island or mainland shore) many fringing reefs have reef flats older than those of nearby offshore reefs. On the Great Barrier Reef micro atolls older than 6,000 year BP are found on many fringing reef flats. However, reef establishment, growth and internal structure can show considerable variation. The structural classification (Fig. 8) is based on the results from Australian fringing reefs, but records from fringing reefs elsewhere in the world suggests that it has a wider application. Three basic types are recognized:

1. Simple reefs formed from the foundation on the lowest portion of the rocky foreshore during the transgression. These reefs were developed while the sea level was still

Coral Reefs, Fig. 8 A structural classification of fringing reefs



rising. Most of their development has gone into upward growth over the rocky slopes of the island, following the transgression. After the still-stand period and when the reef had reached sea level a small amount of outward growth may have been possible over the reef's own fore-reef talus slope. On the whole, such reefs are growing from relatively deep clear water and reef flat development is therefore limited due to the great vertical extent of these reefs. The structure is thus one of a basal framework unit immediately over rock and then a small biogenic detrital frontal unit with thin reef flat veneer.

2. Reefs developed over preexisting positive sedimentary structures. Reefs have long been regarded as requiring hard substrate for initiation. However, there is increasing evidence that the presence of even a muddy sedimentary structure with positive relief may greatly enhance or speed-up reef flat development. Such sedimentary structures may be in the form of terrigenous mud-sand banks or barriers, lee side sand spits attached to islands, boulder beaches, deltaic bar gravels, and low angle Pleistocene alluvial fans. During the transgressive phase, even though a bank may have existed previously, no reef development is possible because of the inhospitable nature of the substrate. However, once the rocky shores of the adjacent island or mainland are inundated reef colonization takes place rapidly on these shores at shallow depth. Progradation of the reef is then rapid over the preexisting structure with hard substrate now being provided over

the sedimentary base by the fore-reef talus from the prograding reef front. The structure of such reefs is thus one of a basal terrigenous sedimentary unit, an inner framework with a prograding carbonate detrital unit extending over the terrigenous base, and an upper, thin, generally less than 4 m, reef flat framework veneer.

3. Reefs developed over more gently sloping substrate, particularly where older foundations of Pleistocene reefs may be present. In these instances, the reef foundation is initiated offshore from the present coastline, although would probably have started as a fringing reef as sea level would have been lower at this time. The rising sea level however, isolated this initial reef which continued to grow upwards during the transgressive phase as an offshore barrier. Possibly, because of poor circulation and terrigenous input, growth behind this barrier was very slow. After the still-stand and after the reef reached sea level, the outer reef became attached to the island by lagoonal infilling, this infill coming from both the land as a terrigenous unit and from the outer reef as a biogenic carbonate unit. Following the still-stand, some small outward growth may also have taken place. The structure is therefore one of a framework unit offshore from the present shore and an interdigitated terrigenous and biogenic detrital fill behind the framework. A thin, reef flat framework veneer may be present over the entire reef and a small reef front biogenic talus may also be present.
4. Davies and Hopley (1983) indicated that fringing reefs grew at rates comparable to middle and outer shelf reefs,

though usually at the lower end of the growth scale, that is, for framework construction in the range of 1–4 mm/year. This may be in part a reflection of the greater proportion of massive as opposed to branching framework. As with outer reefs, the fringing reefs showed a bimodal detrital accretion rate, a lower range of 1–5 mm/year representing accumulation under normal weather conditions, but with higher rates up to 15 mm/year indicative of infrequent high energy cyclonic events. Fringing reefs generally show very low growth rates in shallow water, particularly when compared to the mid-shelf reefs which have both a protected situation and clear water. The lower rate for fringing reefs is probably a reflection of the turbid water conditions and periodic decline in salinities. However, at depths between 4 and 7 m, fringing reef growth seems to be at least equal to the rate of growth of both midand outer-shelf reefs. Below this depth, there appears to be a rapid decline in accretion rate. This is interpreted as the result of the high turbidity of inshore waters and rapid decline in light levels at these depths. Equivalent decline in growth rates for outer reefs takes place below approximately 15 m.

The Future of the World's Coral Reefs

While the store of knowledge about the functioning and evolution of coral reefs has expanded dramatically in the last 50 year, great concerns are currently being expressed about the health and status of the worlds reefs. It is estimated that 10% of coral reefs of the world have already been degraded beyond recognition; 30% are in a critical state such that they will be lost in the next 10–20 year; a further 30% are so threatened that they may disappear as living ecosystems in 20–40 year; and only 30% are healthy and stable (Wilkinson 1993).

The major reasons for this decline are from the anthropogenic stresses of organic and inorganic pollution, sedimentation, and overexploitation all of which are increasing with the exponential growth of human populations. Unfortunately, the majority of the most rapidly growing populations are in the tropics adjacent to coral reefs with subsistence economies at least partly dependent upon the reefs. However, superimposed upon these direct human pressures are the effects of global climate change with coral reefs being regarded as one of the most sensitive ecosystems to the predicted changes (Buddemeier 1993).

Many believe that coral reefs are already showing detrimental effects of climate change with a significant increase in coral bleaching as a response to unusually high temperatures detected globally over the last 25 year (Brown 1997; Hoegh-Guldberg 1999). Bleaching involves the expulsion of the symbiotic zooxanthellae (which gives the coral its color) from the coral tissue. Bleaching may be temporary and may

be an adaptation mechanism that allows more resistant strains of zooxanthellae to recolonize affected hosts. However, as the zooxanthellae are so important to the metabolic functioning of the coral, severe bleaching leads to mortality of whole colonies. Bleaching events appear to coincide with El Niño episodes with significant occurrences in 1979–80, 1982–3, 1987–8, 1991–5, and 1998. As El Niño events are predicted to become more regular and severe in a “Greenhouse” world, the associated short-term higher temperatures will be superimposed on a more general global warming, leading to increased stresses for coral reefs.

There are other harmful effects, including increased sedimentation from more runoff as tropical areas receive more rainfall, possible greater damage from increasing frequency and severity of tropical storms, and inhibition of calcification with CO₂ enrichment of ocean water. Planktonic and larval stages of many reef organisms, including corals, may be particularly susceptible to increasing ultra violet B (UVB) and shorter ultra violet A (UVA) radiation associated with decreasing levels of ozone in the upper atmosphere.

Rising sea levels can produce some beneficial effects. Inundation of reef flats which have been at sea level for 5,000 year or more may regenerate coral growth with reef flats contributing more significantly to carbonate production. Wave efficiency across reef flats may increase and sediments stored there may add to reef island construction. Some reef islands, contrary to popular opinion, can actually grow rather than erode as the result of rising sea level, but with reorientation of many islands also probable, the amount of inhabitable area with older cultivatable soils may decrease. Coral reef islands may not disappear, but their attractiveness for human habitation may decrease. Similarly, the structures which are coral reefs will not disappear as the result of anthropogenic stresses, but their dominance by animals (corals) may be overtaken by algae and other plants for which the Greenhouse scenario is more beneficial.

Cross-References

- Atolls
- Cays
- Coral Reef Coasts
- Coral Reef Islands
- Coral Reefs, Emerged
- Karst Coasts

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Coral Reefs, Emerged

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Corals and other reef-building organisms such as algae live within the photic zone, rarely deeper than 100 m and are limited upwards by intertidal exposure. However, well-preserved reefs emerged up to hundreds of meters above modern sea level are found in many parts of the world. All types of reefs are found emerged although the most common are fringing reefs occurring as terraces built over volcanic or other non-carbonate foundations (Fig. 1). Emerged barrier reefs appear to be rare, even in Indonesia which has both extensive, modern barrier reefs and numerous emerged fringing reefs (Tomascik et al. 1997). However, a section of the very long (2,000 km) Papua New Guinea barrier reef is emerged in Sabari Island (Fig. 2) to the north of the Louisiade Islands, and in the Trobriand and Lusancay Islands to the

north. Many emerged atolls occur in the Pacific (e.g., Henderson, Makatea, Mataiva, Nauru, Walpole) and in Indonesia (Maratua, Kakaban). Detailed descriptions of the geology and hydrogeology of many emerged reef examples can be found in Vacher and Quinn (1997).

Causes of Emergence

It is generally accepted that world sea levels were last ubiquitously above present during the last interglacial about 125,000 years ago and reefs of this age up to 10 m above present sea level have been reported from many quasi stable areas of the tropical world (e.g., Western Australia, parts of Tonga, Aldabra, Curacao). However, most of the world's reefs were exposed during the periods of glacially lowered sea levels (up to 150 m below present). Landforms produced during exposure have been a major influence on modern reefs (see “► Coral Reefs” entry). Emergence during the Holocene is more limited and for isostatic reasons is largely restricted to the Southern Hemisphere (Hopley 1982, 1987). For example, in the Great Barrier Reef region of north-east Australia emerged reefs of about 1 m and with an age of about 5,000 years are found only on the inner shelf islands and mainland and consist of coalescing micro atolls rather than fully developed fringing reefs as the period at the high level was short and followed a rapid post-glacial rise (Chappell et al. 1983).

The most spectacular suites of emerged reefs occur in areas of strong uplift at the margins of the world's tectonic plates (e.g., Ryukyu Islands, Barbados (Fig. 3), Papua New Guinea (Fig. 1), Indonesia). In mid-ocean areas uplift may also occur around hot spot swells or over asthenospheric bumps and is especially responsible for emerged atolls such as Ocean, Nauru, and Marcus Islands (Scott and Rotondo 1983). Long-term average rates of uplift may be as high as 5 m/ka with the highest reefs being of Tertiary age. For example, on Gunung Dirun Timur, Indonesia, terraces at 1,293 m are of Pliocene age (Tomascik et al. 1997). The majority of the best preserved reefs, however, are Pleistocene and because of the rate of uplift represent not only interglacial high stands but also reef-building periods at stable sea levels which were well below present. Emergence in these unstable areas continues with individual earthquakes having the potential to uplift living reef flats >2 m above sea level (e.g., Cortes 1993).

Modifications After Emergence

After emergence, coral reefs undergo changes to both the geochemistry of the reef fabric and to the morphology of the surface features. The original limestone deposited beneath the sea by coral and other plants and animals is largely

Coral Reefs, Emerged,

Fig. 1 Tectonically uplifted emerged reef terraces rising to 270 m on Misima Island, Louisiade Archipelago, Papua New Guinea



aragonite or high magnesium calcite. Once uplifted, however, the limestone is unstable in the freshwater vadose and phreatic environments and diagenesis through solution and reprecipitation as stable phase low magnesium calcite takes place (Matthews 1974). The open framework of the original reef becomes denser, infilled, and with greater cementation (Fig. C78, ► “Coral Reefs” entry). A coarse calcite crystalline structure is produced and in the uppermost layers a caliche horizon may be formed consisting of a distinctive brown crust with numerous micritic stringers (e.g., Hopley 1982), (Fig. C79, ► “Coral Reefs” entry). These processes also operated on reefs exposed during major sea level falls of the Pleistocene. Subsequent recolonization during post-glacial transgressions results in distinctive unconformities (see entry ► “Coral Reefs”), several of which may be found beneath modern reefs.

Erosion of emerged reefs may be relatively slow as the permeable limestone does not support surface drainage. Karst processes dominate but whether or not they have been rapid enough to produce large-scale relief features during the limited exposure of reefs emerged only by sea level fall as suggested by Purdy (1974) is debatable (Hopley 1982, Chap. 7). Bloom (1974) has suggested a facies control over rates of solution which would emphasize the constructional relief differences between reef rim and lagoon floor. Even in ancient reefs, structure may remain evident through facies guidance of solution. For example, the Devonian reefs of Western Australia were buried beneath Permian sandstones then truncated by a Mesozoic-Tertiary planation surface, yet uplift and rejuvenation has resulted in emerged reefs (the Limestone Ranges of the Canning Basin) which faithfully mimic the original morphology. Reefs which are permanently emerged by tectonic uplift do have sufficient time to produce

a range of karst landforms including cave systems, solution holes, and gorges. Their surfaces are an intricate network of holes and pinnacles either completely exposed or buried under soil and in some instances, commercially valuable phosphate deposits. Numerous local terms have been applied to this rocky relief (*makatea*, *feo*, *ironshore*, *champignon*, *platin*, *pavé*).

Emerged reef erosion may also take place horizontally with processes of bioerosion aided by some solution and physical abrasion creating an intertidal notch with overhanging visor (Fig. 2) at rates of up to 7 m/ka (Hopley 1982). Undercuts of several meters may occur and remain etched into the sides of emerged reefs as they are uplifted, providing erosional evidence of former relative sea levels, in addition to the reefs themselves (Fig. 3).

The Value of Emerged Reefs for Scientific Research

The accessibility of an ecological community and landform which is normally underwater, the relatively tight constraint in the relationship of a raised reef surface to a former sea level, and the accuracy of radiometric dating of constituent corals make emerged reefs highly valuable to several branches of scientific research. For example, they have been used in the development, calibration, and refining of dating techniques (e.g., Grün et al. 1992) but they have made particularly prominent contributions to the understanding of past eustatic sea levels, including those of the last glacial period. In some areas, at least, average uplift rates have remained constant over the latter part of the Pleistocene and are calculated from the present level of reefs for which the original eustatic level is well constrained (e.g., the mid-Holocene or the last

Coral Reefs, Emerged,

Fig. 2 Notch and visor cut into the emerged barrier reef of Sabari Island, Louisiade Archipelago, Papua New Guinea. The reef is probably last interglacial in age

**Coral Reefs, Emerged,**

Fig. 3 The back of the 83,000-year reef terrace with sea stack displaying fossil notch and visor cut into older Pleistocene reef limestone, north-east coast of Barbados



interglacial). The uplift rate is then applied to other terrace levels which are dated by radiocarbon or uranium series methods to obtain a past sea level. These techniques were developed originally in Barbados (Broecker et al. 1968; Mesollela et al. 1969) and the Huon Peninsula in Papua New Guinea (Bloom et al. 1974; Chappell 1974). Results are continually being refined (e.g., Ota et al. 1993) and the degree of agreement between results from different areas justifies the basic assumptions.

Evidence from emerged reefs has also been used in geophysical modeling to develop theories of seafloor spreading, and processes associated with hot spots or melting anomalies,

including lithosphere loading, cooling subsidence, and asthenospheric bumps (McNutt and Menard 1978; Scott and Rotondo 1983). They have also aided in the understanding of upper earth rheology, flexure of the ocean lithosphere, and global isostatic responses.

Accessibility allows complete species inventories to be made and corals in particular may remain well preserved in even the oldest elevated reefs. From the data collected, the evolution of genera and species can be followed (Veron 1995) or the changes in ecological communities deduced from time-scales ranging through millions of years to the Holocene (Veron and Kelley 1988; Budd 2000).

Cross-References

- Atolls
- Coral Reefs
- Coral Reefs, Emerged
- Notches
- Seismic Displacement
- Tectonics and Neotectonics
- Uplift Coasts

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Cross-Shore Sediment Transport

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Cross-Shore Transport

Cross-shore transport refers to the cumulative movement of beach and nearshore sand perpendicular to the shore by the combined action of tides, wind and waves, and the shore-perpendicular currents produced by them. These forces usually result in an almost continuous movement of sand either in suspension in the water column or in flows at the surface of the seafloor. This occurs in a complex, three-dimensional pattern, varying rapidly with time. At any moment, some sand in the area of interest will have an onshore component while other sand is moving generally offshore. The separation of the total transport into components parallel and perpendicular to the shore is artificial and is done as a convenience leading to a simpler understanding of a very complex environment (see entry on *Longshore Sediment Transport* for shoreparallel transport).

Cross-shore transport has been well studied because of its importance to beach erosion. Unlike longshore transport, which is difficult to observe, cross-shore transport can result in large and highly visible changes in the beach configuration over intervals as short as one day. Because of the complexities associated with cross-shore transport, its magnitude and direction is usually inferred from changes in the elevation of the beach and the adjacent ocean bottom. Elevation profiles across the beach and through the surf zone are obtained with conventional surveying techniques, augmented as necessary with water depth measurements from a boat or other floating platform. A reduction in the amount of sand above mean sea level and a corresponding increase in the volume below this reference, for example, would be interpreted as offshore transport. If the beach volume increases at the expense of submerged sand levels, this implies shoreward transport. If the measurements were errorfree and the transport was entirely in the cross-shore mode, the amount of sand in the system should remain constant. Gradients in long-shore

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transport, however, can result in local accumulations or losses that make the precise assessment of cross-shore transport difficult.

Small cross-shore adjustments occur during each tide cycle along coasts with significant tide ranges. Much larger cross-shore transport events are associated with major storms, which may last for several days. These changes can be expressed in substantial beach or dune erosion or in rapid formation or migration of bars or other submerged formations. Intervals of smaller waves, and particularly swell of low steepness, can reverse the effects of storms and repair eroded beaches or move bars shoreward. These reversals have much longer timescales, perhaps as long as weeks or months. The accumulated effects of many storms during winter months and long intervals of low waves in summer, for example, can produce seasonal effects through cross-shore transport, as illustrated in Figure C87. Winter beaches may be characteristically lower and narrower with pronounced bars near the location of the largest breakers, while summer beaches are wider and with smaller, less-distinct bars closer to shore. Local wave climates, tide ranges, and sediment size influence these seasonal types, but the changes are a direct result of cross-shore transport (Fig. 1).

A number of investigations have been made in an attempt to predict what wave conditions would result in a shift between offshore and onshore transport. Much of the early work was done in laboratory wave tanks, using simple waves (Johnson 1949). Johnson associated the shift between the formation of summer and winter waves with a critical incident wave steepness in deep water. Other investigators found that the absolute size of the wave changed the critical steepness ratio. A second dimensionless parameter, the ratio of the wave height to the mean sediment diameter, was added by Iwagaki and Noda (1963) to account for scale. Seymour and Castel (1989) review the performance of six simple predictors for the direction of cross-shore transport, which involve some non-dimensional combination from among wave height and period, sediment size and fall speed, and beach slope. The best of these was successful only about two-thirds of the time. Bailard and Inman (1981) introduced physics into the study of

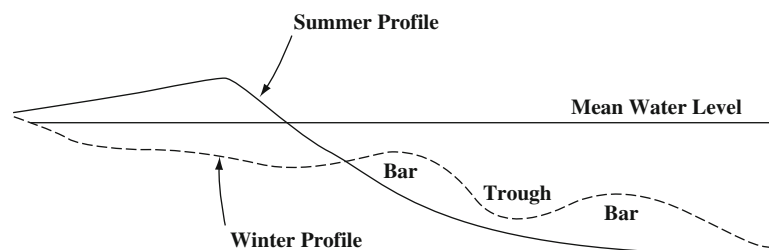
cross-shore transport with a predictive model that was based upon energetics, the balance between gravitation tending to move sediment downslope and the upslope stress produced by wave asymmetry. More recently, numerical models have been developed to predict storm wave erosion (Larson and Kraus 1990).

One of the earliest concepts about cross-shore transport concerned the tendency for waves to selectively move coarser sediment sizes shoreward (Cornish 1898). This results from the asymmetry of waves in shallow water (see entry on *Waves*). As waves approach the beach, the crests become steeper and the troughs longer. The shoreward flows under the crests are higher and for shorter duration than the offshore-directed flows under the troughs. Because larger sediment sizes require higher flow speeds to initiate movement, they tend to be preferentially moved by the shoreward flows. This was later refined to the null point hypothesis in which there is a stable offshore distance for a given sediment size, with a continuous gradation from coarse on the beach to fine in deep water (Ippen and Eagleson 1955). A number of field investigations have shown that sediment sizes are sorted in the cross-shore direction in nature, although bars and other features tend to produce locally mixed sizes as exceptions to regular gradation.

The study of cross-shore transport has resulted in the concept of an equilibrium underwater profile (Dean 1991). The shape of the offshore profile is approximated by

$$h(y) = Ay^{0.67},$$

where $h(y)$ is the depth at distance (y) and A is a scale factor related to the grain size distribution of the sand forming the beach. This is an engineering simplification that assumes, under the constant application of a given incident wave, a particular beach will evolve toward a certain profile shape. The incident wave can be further generalized to include all of the seasonal variation, resulting in a single equilibrium profile configuration for that beach. The actual profile may never achieve this contour because the incident waves are constantly changing. However, it is useful to coastal engineers



Cross-Shore Sediment Transport, Fig. 1 Cross-shore transport. The figure shows schematically how seasonal changes in wave climate affect beach profiles. During winter storms, the dry beach is eroded and seaward cross-shore sediment transport results in the formation of one

or more offshore bars. Typically, the envelope within which the shoreline ranges retreats seaward in winter and shoreward under the milder wave climates of summer

and scientists concerned with understanding the natural variability of beaches. The simple exponential model for the equilibrium beach does not include any provision for bars or other contour complexities such as rock outcroppings.

Cross-References

- [Beach Erosion](#)
- [Beach Processes](#)
- [Beach Profile](#)
- [Cross-Shore Sediment Transport](#)
- [Depth of Closure on Sandy Coasts](#)
- [Dynamic Equilibrium of Beaches](#)
- [Waves](#)

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Cross-Shore Variation of Grain Size on Beaches

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Definition

The distinctive sorting and cross-shore distribution of clastic beach particles is often evident even to the casual observer. The position at which a sediment particle comes to rest on a beach profile and the net patterns of particle size variation resulting from sediment transport processes have long been of interest to scientists and engineers attempting to understand

the fundamental relationships governing process and form on beaches.

Introduction

Cross-shore sorting and distribution of sediment in the coastal environment under waves, currents, and the effects of gravity may take several forms and have a broad range of manifestations in the sediment record. Sediment sorting on a range of scales is the basic process which leads to the formation of stratified layers of finer and coarser sediment (facies), which depend on and reflect the particular combinations of hydrodynamic processes, grain size and density distribution, beach slope and morphodynamic state, and sediment transport regime. The study of grain size variation and sorting of sediment on beaches is of significance to characterizing the morphodynamic behavior of beaches, to the interpretation of facies in the historical stratigraphic (geological) record, and has practical application in the design of beach nourishments, dynamic revetments, and other forms of coastal protection, as well as the assessment of environmental impacts of dredging and construction in coastal environments. Coastal studies involving estimation of beach slopes, equilibrium beach profile definition or sediment transport analysis by numerical or analytical modelling, depend on knowledge of the sediment grain size and its variation at a given site.

This entry reviews the long history of scientific fascination with the natural sorting of sediments which occurs on beaches, provides a brief review of methods of investigation, followed by summaries of the processes of selective sediment transport which result in cross-shore sorting and the observed grain size variation of different beach types from both field and laboratory investigations. Finally, several practical applications are provided with examples drawn from the literature.

Historical Perspective on Beach Particle Equilibrium and the Null-Point Hypothesis

The null-point concept was introduced to express stability of a beach profile and the associated distribution of sediment on the profile as a reflection of the equilibrium between asymmetric fluid forces (including wave asymmetry, steady currents, and the effects of percolation), bottom slope (gravity force), and the size and weight of sediment (Cornaglia 1889). Details of the original null-point hypothesis as expressed by Cornaglia did not predict particle and beach morphological behavior as observed, however. Bowen (1980) and Bailard (1981) improved on the original null-point hypothesis by applying Bagnold's (1966) sediment transport model. According to Bowen, sediment coarser than the equilibrium grain size on the local slope has a smaller value of β/W_s (ratio

of local slope and particle settling speed, descriptive of suspended load) and $\beta/\tan \phi$ (ratio of local bottom slope and grain friction angle, descriptive of bedload). Therefore, assuming a shoreward asymmetry in the velocity field, the term involving gravity is reduced, and coarser particles move shoreward to a position having the limiting slope. Similarly, finer particles should move offshore, as observed (e.g., Zenkovitch 1946).

Sediment transport and profile formation are more complex than are described by the theoretical models of Bowen, Bailard, and others. The concept of a sediment grain being in equilibrium with forcing requires the choice of appropriate temporal and spatial scales for analysis. Time scales as short as a single or even half wave cycle may be involved in the sorting and concentration of mineral grains according to size and density within beach laminae in the swash zone (Clifton 1969). At somewhat longer timescales, antecedent conditions, such as recent storms or periods of swell, spring/neap variations in tide level, water table variations, or a locally significant change in sediment supply or exposure can alter the shape and location of beach profile features and influence the distribution of sediments comprising the beach (e.g., Narra et al. 2015; Chen et al. 2012). At still longer timescales the relative sea level, sediment sources, prevailing climatology, and offshore bottom slope comprise the framework on which beaches and their particle size distributions develop.

Methods of Investigation of Grain Size Variation on Clastic Beaches

Studies of the grain size variation on beaches have most often been conducted by direct in-situ sampling followed by laboratory analysis to determine the grain-size distribution typically using mechanical (sieving) or optical (laser diffraction or laser in-situ scattering transmissometer (LISST)) or photographic analysis methods. Field samples for laboratory analysis can be obtained by a variety of methods that include grab sampling and coring. Short (1984) for example, has investigated the morphology, texture and facies sequence on seven sand beaches, located in low, moderate and high wave energy, microtidal environments in southern Australia using box coring and Scuba observations. Similar work was conducted in non-tidal barred beaches in the Great Lakes and tidal beaches of Atlantic Canada by Greenwood and Davidson-Arnott (1972).

Sediment profile imaging (SPI) techniques have been used to photograph the interface between the sediment bed and overlying water column. The technique is used to measure or estimate biological, chemical, and physical processes occurring in the first few centimeters of sediment, pore water and the benthic boundary layer. Time lapse imaging (tSPI) is used to examine physical and biological activity over natural

cycles, including tides, storms and daylight or anthropogenic variables such as thin layer dredge disposal and feeding loads in aquaculture (see review by Germano et al. 2011).

Wang et al. (1998) reviews various direct methods of measuring sand size distribution under active transport in the nearshore. Conventional suspended-sediment sampling for grain size analysis has generally relied on mechanical pumping and suction techniques (e.g., Nielsen 1983, 1984). The acoustic backscatter and LISST techniques developed by Hay and Sheng (1992) and Agrawal et al. (1996), respectively, are generally capable of estimating vertical profiles of sediment concentration and mean grain size simultaneously with higher temporal and spatial resolutions than pumping, suction, and trapping techniques and have been fairly widely applied in nearshore experimental studies. Limitations with acoustic and laser scattering techniques occur in the breaker zone due to the presence of bubbles. Therefore, Wang et al. (1998) obtained suspended sediment samples throughout the water column in the surf zone using the streamer-trap technique (Kraus 1987) to examine the vertical and temporal variation of sand size distribution under breaking waves.

Sorting and selective transport processes can also be measured effectively using tracer techniques. Sand size sediment tracers include optical detection of magnetic or fluorescent tracers with both natural and synthetic materials (e.g., Smith et al. 2007; Black et al. 2007). Larger particles such as pebbles, gravel and cobbles may be monitored using magnetic particles (Hatfield et al. 2010; Osborne and Chen 2005) or Radio Frequency Identification (RFID) Tags embedded in natural particles (e.g., Osborne et al. 2010).

Laboratory studies involving mobile bed physical models provide a controlled approach for studying sorting processes and resulting morphological development of beach and nearshore environments (e.g., Jiang et al. 2015; Chen et al. 2012; Greenwood and Xu 2001; Tanaka 1989).

Various integrated numerical models for nearshore sediment transport and morphology change now incorporate the effects of variable grain size fractions, sorting, bed armoring and the effects of the vertical structure of bed sediments (e.g., Deltares 2016). Lagrangian particle tracking methods coupled with 1D, 2D, or 3D hydrodynamic models can be used to evaluate grain sorting and selective transport processes in a wide range of environments (e.g., MacDonald et al. (2006)).

Processes of Selective Sediment Transport

The equilibrium profile depends on the sorting (standard deviation) and shape of the sediment as well as its size. Poor sorting results in less percolation and flatter profiles compared with well-sorted sediments of the same size (Krumbein and Graybill 1965; McLean and Kirk 1969). The sorting

processes themselves are dependent on the particle properties as well as the hydrodynamics. Sand-sized particles are sorted according to differences in densities, sizes, and shapes of the sediment particles. Four basic processes of dynamic equivalence or selective sorting act on beach sediments (Komar 1989; Hughes et al. 2000). Settling equivalence, hydraulic equivalence (Rubey 1933), or suspension sorting is the principle that permits a larger grain of lower density to be deposited simultaneously with smaller grains of higher density. Entrainment equivalence or selective entrainment (Komar and Wang 1984) refers to the difficulty of entraining smaller and heavier particles within a matrix of comparatively larger and lighter particles. Smaller particle size can be a direct factor in reducing particle mobility in terms of both relative exposure to flow and larger size of pivoting angle. Selective transport refers to the differential rate at which particles are carried by a flowing fluid by virtue of their position in the flow. Dispersive pressure equivalence or shear sorting refers to the process in which coarser grains tend to migrate upward to zones of lower shear while finer and denser grains move downwards to the zone of maximum shear (Bagnold 1954). The operation of such selective processes in the swash zone may lead to heavy mineral enrichment particularly in erosional episodes.

Part of the cross-shore selective sorting of sediment can be attributed to the patterns of orbital motions under waves as they shoal. Orbital motions under a Stokes-type wave, for example, consist of a forward orbital motion of short duration but high velocity under the crest while the return flow under the trough is slower but of longer duration. Cornish (1898) hypothesized that Stokes asymmetry might explain the shoreward movement of coarse particles and the seaward movement of fine particles. The idea has been verified in experiments (Bagnold 1940) and modeled with some success (Horn 1992). A grain may move incrementally shoreward by falling back to the bed before a wave nearbottom orbital trajectory has completed its half-cycle. A grain that is thrown into suspension and falls to the bed in a longer time than onehalf the wave period may be deposited seaward of its initial location (Dean 1973; Nielsen 1983). Ahrens and Hands (2000) compared sediment stability values with the maximum near-bottom velocity under the breaking wave trough, U_t , calculated with Stream Function Wave Theory to reliably discriminate onshore and offshore sediment transport. Large absolute values of U_t are associated with high suspended load and bedload transport offshore by undertow, and small absolute values of U_t with net onshore bedload transport. Although most of the attention has been focused on the dominant incident waves that arrive at the outer edge of the surf zone, the relative roles of waves of different lengths or frequencies in the selective sorting process in the nearshore zone has not been investigated in detail.

Both field and laboratory experiments have been used to examine the vertical and temporal coupling between the sediment in transport and the near bed fluid motion to gain further insights regarding the fractionation process (see review by Greenwood and Xu 2001). Osborne and Vincent (1996) demonstrate the importance of the phase coupling between the unsteady flow at different elevations and the vertically varying instantaneous suspended sediment concentrations in determining the direction and magnitude of sediment transport in combined waves and currents over rippled beds. Prototype scale wave flume measurements document distinct horizontal fractionation of sediment by size (Greenwood and Xu 2001). Sediment became progressively coarser onshore and progressively finer offshore in the wave flume from the same homogeneous source suggesting that the sorting is explained by suspension transport alone and is controlled by: (1) an increase in the proportion of fines and a decrease in mean size with increasing elevation above the bed and (2) vertically stratified sediment transport with net onshore transport near the bed and net offshore transport at higher elevations.

Cross-Shore Variation in Grain Size and Sorting

Mean grain size may decrease both shoreward and seaward from the plunge point at the base of the swash zone at sandy shores. A secondary coarsening of sediments has been observed on the offshore bar, where wave breaking is initiated (e.g., Fox et al. 1966; Greenwood and Davidson-Arnott 1972). Gravel and coarse sand lags as well as heavy mineral placers can also develop in the troughs or embayments between nearshore bars due to the interaction of waves with strong longshore currents and rips (Komar 1989). Recent laboratory experiments by Çelikoğlu et al. (2006) seem to mimic patterns observed in several field studies showing that finer material tends to be deposited on the crest of bar-type beach profiles, with coarser grains accumulated on the foreshore, trough, ridge, and toe sections. Finer materials settled on the foreshore and crest of step-type profiles while coarser grains were deposited on the ridge and toe sections.

On sandy beaches, mean grain size typically decreases up the foreshore, as a reflection of the diminishing swash intensity. Sediment sorting is poorest (standard deviation of sizes is largest) where energy dissipation is greatest. Skewness is most positive (or least negative) also at the plunge point, indicating a mixing of an over-abundance of fines with coarser sediments, relative to a lognormal distribution of sizes. Bascom (1951) observed a similar pattern, with the additional finding of the second coarsest beach material on the summer berm, and hypothesized that deposition occurred by wave uprush that had overtopped the berm crest and subsequent winnowing of finer fractions by wind. Edwards

(2001) has examined size variation and sorting of sandy beach sediments from more than 274 beaches in five regions of the globe. Edwards found that typically average grain size decreases from inshore (zone of breakers and currents) towards the backshore with foreshore and backshore sands better sorted than inshore sands. Edwards also found that terrigenous beaches are finer than carbonate beaches. Wider coastal plains exhibit finer and better sorted sediments.

In one of the more comprehensive and systematic studies of the relationship between size distribution, beach facies and morphodynamics, Short (1984) examined the variation in sediment texture and facies both within and between seven beaches in South Australia. Short found that low-energy reflective beaches are limited in lateral and vertical extent and in facies to beach laminations separated by coarse step deposits from finer nearshore cross-lamination facies. Moderate-energy intermediate beaches characterized by rip circulation possess increasingly wider surfzones with ridge and runnel and bar-trough facies separating the beach and step facies from the more extensive nearshore sequence. High-energy dissipative beaches may have 500 m wide surfzones containing multiple bar-trough topography. Fine beach laminations with backwash structures grade into 4–5 m thick bar-trough sequences then the extensive nearshore facies. As wave energy increases from low ($H_b < 1$ m) to high ($H_b > 2.5$ m) the vertical extent of the beach to nearshore sequence increases from < 10 m to approximately 30 m, and the width from 100 m to several kilometers. Consequently one would expect higher-energy paleo-beach sequences to be represented more by diagonal than vertical facies sequences.

The overall range in grain-sizes present on a beach has significant implications for both the cross-shore distribution of sediment and the form and behavior of the beach. Beaches with a broad mixture of sediments from silts, sands, gravels, cobbles and boulders often have a composite profile with a distinct break in slope between the upper foreshore where coarse sediment prevails and the low-tide terrace where finer sediment is more prevalent. Sediment size distribution on mixed beaches is responsible for beach hydraulic conductivity, which in turn, controls the beach profile and groundwater flow and also for modifying swash zone asymmetry (e.g., Mason and Coates 2001). High beach permeability leads to asymmetry of swash, such that swash is stronger during the up-rush than during the backwash. This effect tends to roll both gravel and sand shoreward during the up-rush, but mobilizes finer material seaward during the backwash, effectively stranding gravels higher on the beach with each successive wave. Mixed sediment beaches behave differently than either pure sand or pure gravel beaches. For example, a mixed-sediment profile will reflect more energy than both a sand beach (due to a steeper gradient) and a gravel beach (due to less energy dissipation through infiltration) (Mason and Coates 2001). Mixed beaches also exhibit distinctive

armoring in which gravel and pebbles form a thin surface layer over a more homogeneous mixture of sand, pebbles and gravel (Finlayson 2006).

On lower energy beaches with a broad distribution of sediments including cobbles and boulders, or beaches where sediments are simply too large to be re-distributed by the wave climate (such as Puget Sound or in sheltered estuaries), the beach profile may become reverse graded with the presence of lag stones on the lower profile (Finlayson 2006; Jackson et al. 2002; Nordstrom 1992). Lag stones are large boulders that erode from adjacent coastal bluffs and are deposited on the beach face rolling down the steep beach face due to gravity and forming a beach armor or lag deposit at or near the break in slope. Sediment size decreases both landward and seaward of the lag armor.

Beach sand sampled at transects across the active profile at Terschelling (North Sea barrier island of The Netherlands) showed consistent cross-shore pattern of grain size distribution over a period of 19 months (Guillén and Hoekstra 1996). The proportion of sample in each grain size fraction (50 micron increments) showed a distinctive cross-shore pattern, unique for each size fraction and termed the equilibrium distribution curve. The distribution of the fraction across the profile, relative to the total presence of the fraction, expresses the probability concept in cross-shore sorting.

Field studies of the spatial/temporal relationships between sand size distribution and morphology have been conducted by Sonu (1972) and Katoh and Yanagishima (1995) and others. More recently, Narra et al. (2015) conducted field studies of the spatial (cross-shore) and temporal (seasonal) variation in d_{50} by weekly sampling of sediments on Barra Beach, Aveiro. They found that sediments were typically coarser in the intertidal zone and breaker zone while generally decreasing offshore. A distinct coarsening in d_{50} was observed in foreshore sediments after storms and generally a greater variation occurred in winter months (storm season). Greater variance in d_{50} was also observed in the mid portion of the profile, whereas the extreme ends (both onshore and offshore) showed a tendency for less variation. Similar studies have been conducted by Medina et al. (1994) at “El Puntal” Spit, Santander, Spain. Principal component analysis of sediment samples and beach profile variability was used to conclude that spatial variability of both sediment and profile data are strongly related and each sediment size responds to the same hydrodynamics differently. Beach profile recovery following a storm started with the movement of fine material from the bar crest.

Application

Geomorphologists, engineers, and planners need to reliably project structural and beach fill performance, in terms of

offshore and longshore losses of beach material, or the change in beach profile or character of the beach face, after the material is sorted by natural processes acting over a period from years to decades.

Dean (1990) illustrates the application of an understanding of equilibrium profile behavior is useful to design and evaluation of coastal engineering projects including design of beach nourishments using empirical correlations between the profile parameter and the dimensionless fall velocity which is dependent on sediment size. Applications include analysis of shoreline position change, profile evolution, sediment redistribution and the effects of structures, under changes in water level, wave heights, and sand size variation.

Sediment supplied in a nourishment to Terschelling, The Netherlands, caused a coarsening (20–40 μm) of the sediment in the zone directly affected by the nourishment (Guillén and Hoekstra 1997). Six months after the nourishment, the grain size distribution across the profile was nearly the same as the original, and no significant effects of the nourishment could be recognized in the median grain size. Individual grain size fractions displayed a temporal evolution more complex than the median size, and significant changes, unrelated to the sand supplied, were observed. Overall results of Guillén and Hoekstra (1997) indicate that the shoreface nourishment only had a short-term and localized impact on the sediment distribution. Some months after the nourishment, the former grain size distribution was reestablished. On a yearly perspective, the natural variability of the sediment was higher than the changes caused by the nourishment.

Understanding sediment size distribution and its variability is also of relevance to assessing the impacts of construction in the marine environment (e.g., Dean 1990), the effects of sand mining (e.g., Hilton and Hesp 1996), and dredging and disposal at sea (e.g., Germano et al. 2011).

Studies of grain sorting also have relevance to understanding the formation of placers, which may have economic or intrinsic value (e.g., Komar and Wang 1984). Understanding the processes that cause size variation of sediments across a beach profile and the impact of the material composition on shore morphology is key to wise management of coastal resources.

Large and often devastating tsunamis as well as hurricane related storm surges can mobilize and redistribute substantial amount of coastal sediments, changing coastal morphology and sorting sediment in the process. Chen et al. (2012) conducted laboratory experiments in a wave flume to investigate changes to beach profile and mean grain size caused by tsunami-like waves. Their results showed that sand suspension and erosion were caused mainly by sheet flows when water retreats from the beach, whereas offshore sandbars, composed mainly of coarser sand, formed mainly from

the run-down hydraulic jumps caused by interactions between retreating water with the waves in the undulating tail of the tsunami. Similar studies (e.g., Chen et al. 2012; Lario et al. 2010) can assist in the interpretation of tsunami and major storm deposits in the historic record and contribute to the understanding the potential impacts of future events.

Summary

The cross-shore sorting of noncohesive sediment often results in distinctive visual patterns as well as the formation of a variety of sedimentary structures (or facies) and morphological features. Beaches exhibit a wide range of variation in sediment type and distribution which is often reflected in their morphological characteristics and behavior. Generally, average grain size decreases from inshore (zone of breakers and currents) towards the backshore with foreshore and backshore sands better sorted than inshore sands. Sediments are typically coarser in the intertidal zone and breaker zone while generally decreasing offshore. As wave energy increases from low ($H_b < 1$ m) to high ($H_b > 2.5$ m), the vertical extent of the beach to nearshore sorting sequence increases from < 10 m to approximately 30 m and the width from 100 m to several kilometers. The sorting process is controlled by a range of hydrodynamic processes (e.g., waves of different frequency and of variable symmetry and nearshore currents) as well as the grain size and density distribution of available sediment, beach slope and morphodynamic state of the beach.

Methods for investigation of grain size variation on beaches have advanced from direct sampling and physical/mechanical laboratory analysis to include optical (laser diffraction or laser in-situ scattering transmissometer (LISST)) or photographic methods; the latter techniques also including in-situ methods that avoid significant disturbance of the sediment structure. Numerical models for sediment transport now incorporate many of the effects of variable grain size fractions and can be used to evaluate sediment sorting along transport pathways.

Understanding sorting and distribution of sediment has significance with respect to characterizing morphodynamic character and behavior of beaches and for the interpretation of facies in the geological record. The topic also has practical application to the design of beach nourishment, shore restoration, and coastal protection, as well as in the assessment of environmental impacts of dredging and construction in coastal environments. Coastal studies involving estimation of beach slopes, equilibrium beach profile definition, or sediment transport analysis by numerical or analytical modelling also depend on knowledge of the sediment grain size and its variation at a given site.

Cross-References

- Beach Nourishment
- Beach Processes
- Cross-Shore Sediment Transport
- Dynamic Equilibrium of Beaches
- Sandy Coasts
- Sediment Suspension by Waves
- Surf Zone Processes

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Cusplate Forelands

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Cusplate Forelands are accretionary features, which occur on many coastlines of the world; Cape Kennedy (Cape Canaveral) on the east coast of the United States and Dungeness on the south coast of Britain being but two of

the more well-known examples (King 1972). While they vary considerably in shape and size they are generally triangular in outline with the base attached to the coastline and the apex facing the open sea or ocean. Although similar in overall shape, cusplate forelands must not be confused with cusplate spits; the former are cusplate forms which have grown by progradation, usually with nested shore-parallel beach ridges. Cusplate forelands are related to beaches, tombolos, spits, and barriers but whereas the related accretionary features tend to smooth out the coastal outline cusplate forelands accentuate coastal irregularities.

Formation

A number of theories have been forwarded to account for their origin. They may develop in the lee of an offshore island that shelters a segment of the coast in the shadow of the predominant oncoming wave front (Bird 1984). Waves are refracted to either side of the offshore island and concentrated wave energy affects the mainland shore to either side. Immediately behind the protective island diminished wave energy allows longshore moving sediment to accumulate. If the offshore island is sufficiently near the mainland the cusplate foreland may extend to the island itself in which case a tombolo develops. This has been observed in South Australia at Gabo Island. Here a cusplate foreland developed, it then joined the offshore island to form a tombolo and subsequent storm breaching made it a cusplate foreland once more. Formation in the lee of a protective offshore island however is only one way in which they can form.

More difficult to explain are cusplate forelands which develop on an open coastline with no protective offshore islands (Steers 1964). The east coast of England affords many examples of this type. Lowestoft Ness (the most easterly point of the British mainland) is one such feature. Whereas the cusplate forelands formed in the lee of protective islands remain relatively stationary, forelands with no such protection can, and very often do, migrate along the coast (Russell 1958). Surveys in the British Isles have tracked such features migrate along a coastline many kilometers. Banacre Ness in Suffolk has moved northwards by some 2 km over the last 100 years in spite of the dominant longshore drift being to the south (Robinson 1965). In some cases, the longshore migration has oscillated with periods where a foreland has migrated in one direction followed by periods where reverse migrations have occurred (Hardy 1966). Occasionally two adjacent cusplate forelands on an open stretch of coast have migrated in opposite directions. Such variations and contrary movement defy simplistic explanations relating foreland development merely to longshore drift.

While beach drifting and dominant waves are important, offshore current residuals and changing seafloor topography

are significant contributory factors. Offshore from the British east coast examples can be found of bank and channel complexes developed by the ebb and flood of the tide. Ebb- and flood-dominated channels tend to be separated by sand banks and adjacent to the apex of each foreland can be found an offshore sandbank trending toward the coast (Robinson 1964). Over time any change in the position of the offshore bank has resulted in a change in the position of the foreland, albeit after a short time lag. The supply of sediment arriving on the beach from the offshore bank is greater than the removal of the sediment along the beach and hence the local build-up of sediment at the landward termination of the adjacent sand bank. The movement of sand onshore from the adjacent sandbank together with wave refraction caused by the offshore sand bank itself accounts for the creation of the foreland and can account for its movement along the coast, which may be in a contrary direction to the main longshore drift.

Surface gravel, which is often present on the British cuspate forelands, is usually found in ridges—it is the tracing of these shingle ridges, which gives a clue to both their formation and migration, (Lewis and Balchin 1940). The shingle moves along the surface of the beach and accumulates in the area of lower wave energy. While the offshore sandbanks offer less protection than adjacent offshore islands the lesser influence on wave refraction is made up for by their contribution of sediment to the adjacent coastline.

Cross-References

- [Barrier](#)
- [Beach Ridges](#)
- [Gravel Barriers](#)
- [Gravel Beaches](#)
- [Offshore Sand Banks and Linear Sand Ridges](#)
- [Spits](#)

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Dalmatian Coasts

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Definition

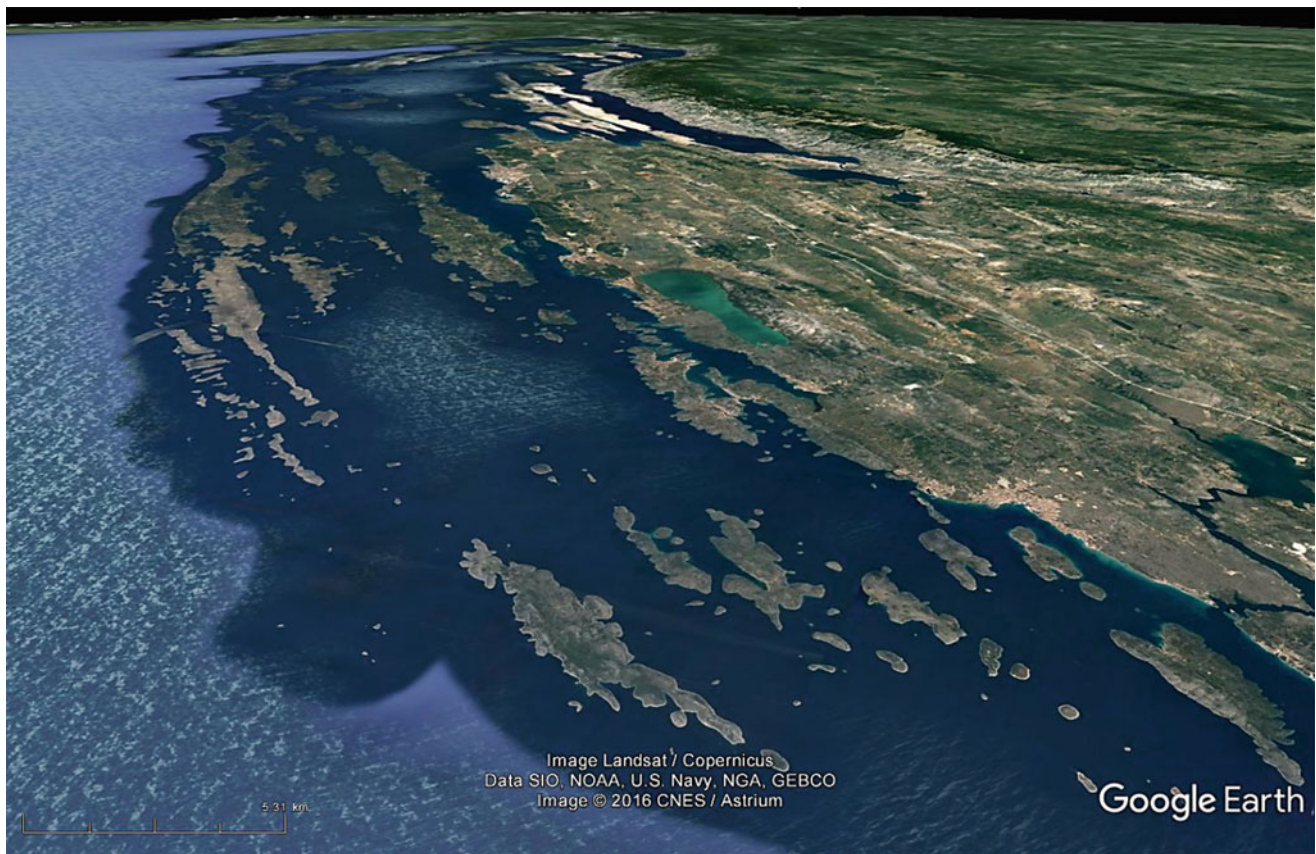
The term “Dalmatian Coasts” refers to a prototype of a primary coast by ingression of the rising postglacial sea into a relief of coast-parallel anticlines and synclines from a young orogenesis, named after the landscape of Dalmatia (Croatia, former Yugoslavia, Adriatic Sea, Mediterranean; Charlier 2010). In the eastern Adriatic Sea, this coastal landscape extends for 400 km from Rijeka in the north to Dubrovnik in the south. Single islands are up to 60 km long but only a few km wide.

Elongated cuestas may build the island chains instead of anticlines. Narrow channels between long islands are called “vallone” or “canale” (from the Italian word for channel or valley), and the Dalmatian Coast therefore is named canale or vallone coast, as well (Fig. 1). The coastlines of the central part of Croatia, built up by Mesozoic limestones, show only very little forming by true littoral processes (Pikelj and Juračić 2013). Beaches are missing as well as extended cliffs. The first is because of the small discharge or limited discharge area in a limestone environment with

strong karstification and the second because of a weak surf energy along the Adriatic Sea or inside of the canale and vallone with small fetch. Therefore, the coastlines exactly follow a contour line of an older relief, formed during longer periods of terrestrial denudation. Partly drowned karst features (dolines, uvala, poljes; Suric et al. 2005, 2014) decorate some of the coastlines, in particular along the mainland coast, where fresh water karst springs at the coast or even under the sea are numerous. The overall picture of the Dalmatian Coast gives the impression of submergence and down warping. Indeed higher coastal terraces or deposits from earlier interglacial sea-level highstands are missing. This picture will change very little during a further sea level rise or sea level fall. In the first case, the ingression will drown another depression, while the outermost islands will disappear under a rising sea level (Furlani et al. 2014), and in the second case, new island chains will emerge, while canale or vallone transform into inland depressions.

Conclusion

This type of coastline is rare; another large example is the Island of Sumatra with the Mentawai Islands parallel in the Indian Ocean. The terms “Dalmatian Coasts” as well as “vallone” and “canale” are not generally in use outside of the Adriatic Sea, although they exist in different dimensions worldwide.



Dalmatian Coasts, Fig. 1 Canale – and vallone – coastline type in the eastern Adriatic Sea (Credit: Google Earth)

Cross-References

- [Karst Coasts](#)
- [Submerged Coasts](#)
- [Submerging Coasts](#)

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Dams, Effect on Coasts

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People have constructed dams, weirs, or barriers on rivers to utilize and manage water resources for thousands of years. The utilization of water energy also has a long history, such as waterwheels and watermills, etc. The initial dams were used to control water resources and to supply water for irrigation and domestic purposes. Later, the energy of rivers was harnessed behind dams to power primary industries directly, and more recently still, hydroelectricity has been generated using water held in reservoirs, allowing the impoundment and regulation of river flow. The modern era of major dams dates from the 1930s and began in the United States with the construction of the 221 m-high Hoover Dam on the Colorado River. But in the last 50 years there has been a marked escalation in the rate and scale of construction of large dams

all over the world, made possible by advances in science and technology. Some major rivers in the world have been intensively manipulated in this way.

Today, approximately one-third of the countries of the world rely on hydropower for more than half of their electricity supply. The hydropower provides 20% of all electricity globally. If this electricity were produced by the combustion of fossil fuel, carbon dioxide emissions would increase by 2 billions tons a year. The construction of dams on rivers, as an infrastructure of national economy, plays an important role in controlling flood, hydroelectricity generation, irrigation, navigation, water supply of industry and agriculture, etc. There is no doubt that the construction of large dams has been successful in achieving their socioeconomic benefits. However, dams alter river hydrological regimes, accordingly induce structure and function changes of ecosystems, and inevitably have marked effects on nature and ecological environment of river drainage basins. An estuary is the only pathway of materials from the drainage basin into the sea. The coastal area far away from dams is the sink of riverine freshwater, sediments, and nutrients. Dams on rivers alter their downstream material discharge into the sea, accordingly induce impacts on estuaries and the coastal zone, such as estuarine channel evolution, the actions between runoff and tide, coastline changes, saltwater intrusion in estuarine and land salinization in delta areas, and ecosystem variations in estuaries and seas. Thereby, damming has positive and negative socioeconomic effects, but in the coastal area it has more negative effects.

Status of Dam Construction

The International Commission on Large Dams (ICOLD) defines *large dams* as usually >15 m from foundation to crest. Dams of 10–15 m can also be defined as large dams if they meet the following criteria: crest length of 500 m or more; reservoir capacity of at least 1 million m³; maximum flood discharge of at least 2,000 m³/s; “specially difficult” foundation problems, or “unusual design.” *Major dams* meet one or more of the following criteria: at least 150 m-high; havin a volume of at least 15 million m³; reservoir capacity of at least 25 km³; or generation capacity of at least one gigawatt.

Based on the definition above, there are some 45,000 large dams, most of which were built in the past 50 years, that now obstruct the world’s rivers. There are 306 major dams in the world today and 57 are planned in the near future (Revenge et al. 1998). The most active phase of large dam construction was in the period between 1950 and the mid-1980s, when 885 dams were completed on average each year. At the end of 1960, there were 7,408 such dams

registered. By 1986, the total had reached 36,562. Of those constructed, 64% were in Asia, with no less than half of the world total in China. Excluding China, most of the new structures have been built in the temperate zone, but tropical and subtropical countries such as Brazil, Mexico, India, Thailand, Indonesia, Zimbabwe, Nigeria, Cote d’Ivoire, and Venezuela have also become prominent in dam construction. Of these dams 79% are less than 30 m in height, and only 4% exceed 60 m. In total, the world’s major reservoirs were estimated to control about 10% of the total runoff from the land in the 1970s. That figure had risen to 13.5% by the early 1990s (Postel 1996) and must be nearing 15% today. The world’s largest impoundment, the 8,500 km² Volta Reservoir behind Ghana’s Akasombo Dam, flooded 4% of that nation’s land area. There are 10 watersheds with more than 5 major dams (Table 1).

To date, China has built almost half (22,000–24,000) of the world’s estimated 45,000 large dams, and remains one of the most active dambuilding countries today. Over 22,000 of the estimated 85,000 significant reservoirs and dams in China are considered as large dams, though there are perhaps a few million small-scale and localized water diversions, check dams and weirs built by local collectives or individual farmers that are unrecorded. China has thus built more than three times the number of large dams in the United States and over five times the number in India. Virtually, all these dams were built since the founding of the People’s Republic of China in 1949, as only 22 large dams existed before that time. The water resources controlled about 100 billion m³, risen to 565 billion m³ today. At least 90 dams over 60 m are under construction, including the Three Gorges dam on the Yangtze (175 m) and Xiaolangdi on the Yellow River. The utilization rate of river water resources increased to 19.9% in 1997, while the Yellow, Huaihe, and Haihe River are all over 50%, and the Haihe River is nearing 90%. In the United States, 5,500 large dams make it the second most dammed country in the world.

Dams, Effect on Coasts, Table 1 Watershed with more than five major dams

Watershed	Number of dams
Paraná (South America)	14
Columbia (North America)	13
Colorado (USA)	12
Mississippi (USA)	9
Volga (Russian)	9
Tigris and Euphrates (Mid-East)	7
Nelson (Canada)	7
Danube (Mid-Europe)	7
Yenisey (Russian)	6
Yangtze (China)	6

Effects of Dams on Coast

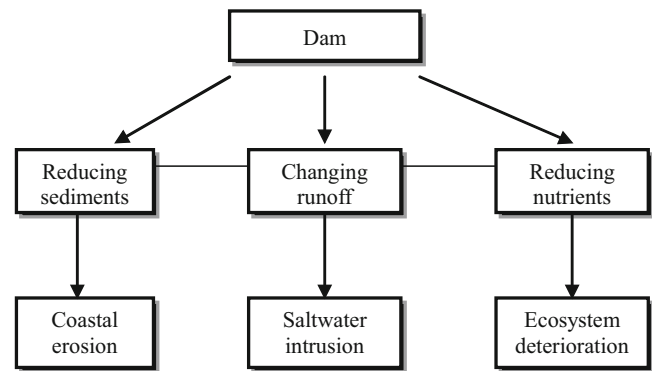
Alternation of Material Discharge into the Sea

Dam construction on a river changes the flow regime of almost the entire river, both changing its seasonal variations and/or reducing its overall volume, and further change and/or reduce sediment and nutrient outflow. Consequently, almost all parts of the estuarine and coastal environment can be impacted by changes to flow, sediment, and nutrients.

First, damming and reservoir construction change water discharge into the sea. The nature of the impacts depends on the design, purpose, and operation of the dam, among other things. For instance, following construction of the Aswan High Dam on the Nile River, Egypt and Lake Powell River dam on the Colorado River, USA not only altered the seasonal variations of natural flow but also reduced its overall water volume into the sea. One well-documented example of additional stresses created by dams pushing an ecosystem toward collapse has been documented in San Francisco Bay. Prior to 1850, approximately 34 km³ was discharged on average into San Francisco Bay from the San Joaquin and Sacramento River systems. Today, there exists more than 20 km³ of water storage capacity in the Central Valley. Approximately 40% of the flow in these Central Valley rivers are removed for local consumption and 24% is pumped from the San Francisco Bay Delta and exported to Southern California and the Central Valley. Less than 40% of the river flows are now discharged to the Bay (Nichols et al. 1986). In the extreme drought of 1977, releases dropped to 100 m³/s, from the customary dry season discharges of 400 m³/s.

Second, because of damming, sediments are intercepted in reservoirs behind dams and reduce sediment discharge into the sea. Before the Aswan High Dam, the Nile River carried about 124 million tons of sediment to the sea each year, depositing nearly 10 million tons on the floodplain and delta. Today, 98% of that sediment remains behind the dam. The construction of the Three Gorges Dam on the Yangtze River, China, will not change total water discharge into the sea, just change seasonal distribution of runoff, but will reduce sediment outflow by approximately 20%. Otherwise, besides reducing sediment discharge, dams may also change the grain size of sediments, large amounts of wash load are intercepted in reservoirs, reducing the supply of fine-grained sediments to estuaries.

Dams not only alter the pattern of downstream flow and sediment (i.e., intensity, timing, and frequency) they also change nutrient regimes. Nutrients carried by runoff and sediments discharge changes with water and sediment outflows. The Colorado, Nile rivers have weak nutrient levels, probably because of high abstraction rates from large reservoirs, leaving little discharge into the sea.



Dams, Effect on Coasts, Fig. 1 Effects of dam construction of rivers on coast

Dams alter the conditions of water, sediment, and nutrient discharges into sea, leading to huge impacts on the environment and ecosystems in estuaries and on coasts (Fig. 1), such as coastal erosion, saltwater intrusion, and position changes of mouth bars in estuaries, nursery ground for many valuable fishes. Dam construction on an upstream river, although it does not alter overall volume of outflow, changes its seasonal variations and reduces sediment discharge, which all have profound effects on a river mouth and its adjacent coast.

Coastal Erosion

A coastline is dynamic and constantly changing due to coastal erosion and accretion, and its change rate is determined by natural processes, such as rough seas, sea-level rise, high tides, nearshore currents, runoff, as well as human developments that can restrict or accelerate the volume of sediment available for shores. A constant supply of sediment is necessary for depositional coasts to form and be maintained along this coastline. Many human activities reduce the supply of sediment that reaches the sea and, in turn, deprive coasts of replenishment. These activities include dam construction and other developments. Lack of sediment replenishment creates greater vulnerability for coastlines that have always been subject to varying levels of erosion. A principal coastal concern today and in the foreseeable future is coastal erosion. It is estimated that 70% of the world's sandy shoreline are eroding (Bird 1985). In the United States, the percentage may approach 90% (Leatherman 1988). This worldwide extent of erosion suggests, that eustatic sea-level rise is an important underlying factor, while many other processes, especially reduced sediment discharge into the sea, are important causes. Dam construction on rivers which intercepts sediments in reservoirs is one of the causes for reduced sediment discharge into sea. Because of dams, the sediment supply to an estuary and its adjacent coast is reduced, thus accelerating coastal

erosion. Even in a coastal delta of rich sediment supply, such as the Yangtze Delta, dams on the river may slow its accretional rate and gradually transfer to erosion.

Reduced sediment discharge directly affects coastal delta progradation, leading to loss of beach. For example, the slow accretion of the Nile delta was reversed with the construction of the delta barrage in 1868. Today, other dams on the Nile, including the Aswan High Dam, have further reduced the amount of sediment reaching the delta. As a result, much of the delta coastline is eroding by up to 125–175 m/year. Deltas below impoundments tend to shrink, reducing habitats, because of the capture of sediments by impoundments. The Nile Delta, Egypt, has shrunk at a rate of 125–175 m/year (Rozenfurt and Haydock 1993). The consequence of reduced sediments also extends to long stretches of coastline where the erosive effect of waves is no longer sustained by sediment inputs from rivers. Another example of this problem is along the mouth of the Volta River in Ghana. Akosombo Dam has cut off the supply of sediment to the Volta Estuary, affecting also neighboring Togo and Benin, whose coasts are now being eaten away at a rate of 10–15 m/year. The break-off water discharge in the lower Yellow River, China results in a sharp decrease of sediment discharge into the sea and its delta erosion. The major causes are damming and water transfer. The delta erosion of the Yellow River is an issue that needs to be urgently solved due to water and soil preservation and construction of Xiaolangdi Reservoir.

Accelerated coastal erosion has been detected as a result of regulated flows and dams reducing sediment discharge to the littoral cells. For example, local coastal regions along the Egyptian Mediterranean coast have experienced erosion rates that were three times greater than in the period prior to the Aswan High Dam (Smith and Abdel-Kader 1988). The construction of the Yangtze Three Gorges Dam will reduce sediment discharge and increase sediment transport capacity of flow in its estuary. Although it is of benefit to the sedimentation of navigation channel, it increases the action of bed sediments and leads to channel instability. Furthermore, it will increase coastline erosion and slow tidal flat accretion seaward (Chen and Xu 1995).

Saltwater Intrusion

Estuarine salinity varies with runoff, wind wave, and tidal current, and has complex spatial and temporal changes. Generally, estuarine salinity is higher during a dry season and is lower during flood season. There is a negative correlation with water discharge. Therefore, the water discharge is an important cause of saltwater intrusion. The change of water discharge leads to the alternation of the saltwater intrusion pattern: distance and intensity. Decreased

discharge rates can result in saltwater intrusion along estuaries, and pollute estuarine water and groundwater. It will affect the development of freshwater resources, especially domestic and agricultural water. In the delta area, because of impounding upstream, the aquifers underground becomes saline, and leads to a shortage of fresh groundwater. Besides, the alternation of seasonal distribution of runoff also affects seasonal saltwater intrusion, thus altering the position of estuarine balance zone.

Reduced water discharge due to dams undoubtedly intensifies saltwater intrusion in estuaries. However, seasonal regulation of dams may mitigate saltwater intrusion due to increased water outflow during the dry season. For instance, salinity intrusion is a problem in the Yangtze Estuary because of the vast area using water from the Yangtze. The water stored by Three Gorges at the end of the flood season will be released to increase the dry season flow and may help to alleviate the salinity problem in the estuary.

The intrusion of seawater into delta areas when river flows have declined due to reservoir impoundment is another common downstream effect. Construction of the Farakka Barrage on the River Ganges in India in 1975 has significantly increased the amount of cropland plagued by salinity problems in Bangladesh, as dry-season discharge has declined. Nationally, nearly 350,000 ha of land in Bangladesh was affected by salinity problems before the barrage was built, but by the mid-1990s this total had more than doubled to 890,000 ha. Downstream changes in salinity due to construction of the Cahora Bassa Dam in Mozambique are also threatening mangrove forests at the mouth of the Zambezi. Besides, reduced sediment discharge due to dams not only affects delta erosion but also intensifies land salinization.

Effects of Ecosystem and Biodiversity

A river's estuary, where freshwater meets the sea, is a particularly rich ecosystem. Some 80% of the world's fish catch comes from these habitats, which depend on the volume and timing of nutrients and freshwater. The riverine nutrients exist in water flow and sediments, carried and transported by flow and sediments, increase the nutrient level in estuaries and the adjacent sea. The decrease and alteration of the flows and nutrients reaching estuaries have marked effects on coastal and marine ecosystem and biodiversity, such as the nursery ground of fishes in estuaries and marine fish catch. The alteration of the flows reaching estuaries because of dams is a major cause of the precipitous decline of sea fisheries in many estuaries of the world.

The influence of diminished freshwater outflow to bays, estuaries, and coastal wetlands has become generally accepted during the past decade. One well-documented example of additional stresses created by dams pushing an

ecosystem toward collapse has been documented in San Francisco Bay. This reduction in freshwater inflow contributed to a drop in phytoplankton biomass to less than 20% of normal, and zooplankton was also significantly reduced. These conditions resulted in striped bass, one of the key indicator species of the health of the ecosystem, being reduced to the lowest recorded levels (Nichols et al. 1986). Water withdrawal on the North Caspian had the following effects (Rozengurt and Hedgpeth 1989): (1) the mean salinity increased from 8 to 11 ppt; (2) the estuarine mixing zone was compressed and moved up to the delta; (3) the nutrient yield, especially phosphorus, and sediment load were reduced by as much as 2.5 and 3 times, respectively; (4) biomass of phytoplankton, zooplankton, and benthic organisms were decreased by as much as 2.5 times; and (5) a substantial part of the Volga flood plain that served as a nursery ground for many valuable fishes was transformed into drying swamps or deserts. The regulation of the Volta River in Ghana by the Akasombo and Kpong dams has led to the disappearance of the once-thriving clam industry at the river's estuary, as well as the serious decline of barracuda and other sport fish. The effects of increased salinity on fishes of the Nile Delta have been documented that out of 47 commercial fish species in the Nile prior to the construction of the Aswan High Dam, only 17 were still harvested a decade after its completion. The annual sardine harvest in the eastern Mediterranean has dropped by 83%, probably the effect of a reduction in nutrient-rich silt entering that part of the sea. The effect of lowered nutrient input is generally the greatest in the first year of life of the fishes.

Decreased discharge rates can result in an increase in salinity in estuaries and change the composition of species in this zone. The Danube Delta, central Europe, coastline is receding at a rate of up to 17 m/year, threatening benefits from tourism down to bird life. The delta supports large populations of bird species that are generally widespread over Europe; some 170 species of birds breed in the delta, including pelicans, herons, ibises, and terns. Impoundments upstream, including seven major dams, on the 2,860-km-long river, channelization and the loss of the nutrient absorption capacity of upstream floodlands has meant that nutrients and other pollutants are affecting delta water quality. Bird populations are at a fraction of their historical numbers. So, although reservoirs may provide new habitat upstream their impacts on birds may be negative in the long-term.

The effect of dams on the ocean might be expected to be significant. The Three Gorges Dam on the Yangtze River in China is expected to reduce the productivity of the East China Sea, one of the largest fishing grounds in the world. It has been estimated that if the Yangtze River outflow is cut back by 10%, the cross-shelf water exchange will be reduced by 9%

with a similar reduction in the onshore nutrient supply. Primary production and fish catch in the East China Sea is expected to diminish by a similar proportion (Chen 2000).

Conclusion

Dam construction on rivers is a great act that human conquers nature and modifies nature. It has twofold effects: positive and negative. The positive roles are irrigation, power, flood control, and water supply, etc. In the meantime, it also leads to negative effects on environment, involving entire river basin, estuary, coast, and sea. The effects of dams on coast include coastal erosion, saltwater intrusion, ecosystem, and biodiversity, etc., and these effects are profound and mostly negative.

Cross-References

- [Asia, Eastern, Coastal Geomorphology](#)
- [Beach Erosion](#)
- [Deltas](#)
- [Erosion Processes](#)
- [Estuaries](#)
- [Hydrology of the Coastal Zone](#)
- [Submerging Coasts](#)

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Deltaic Ecology

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A System Overview

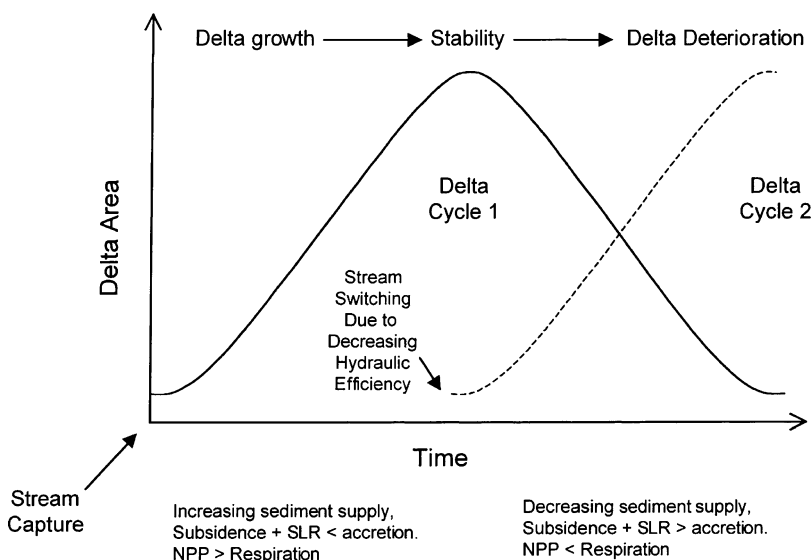
Deltas are depositional areas where rivers meet the sea, often forming extensive, coastal, alluvial fans. Although there is much variability, deltas tend to develop where high-energy rivers deposit their sediment loads into shallow, low-energy marine systems or, more generally, where river energies exceed marine energies. For example, the Mississippi River discharges an average of 6.2×10^{11} kg of sediments per year into the shallow, low-energy waters of the Gulf of Mexico, allowing for the development of an extensive delta complex spanning 30,000 km². Although deltas can dominate the coastal landscape, they are by no means static; characteristically undergoing a cycle of growth, channel abandonment, and deterioration (Fig. 1). Thus, the ecological populations and communities associated with deltas are unique in that they are subject to powerful allogenic riverine and marine forces, and to the overarching pattern of the deltaic cycle itself. For this reason, an introductory discussion of deltaic ecology is best approached at an ecosystem level, which considers the interactions between the biota and the abiotic forces that simultaneously drive, stress, and subsidize them.

Ecosystem Types

Since an entire deltaic complex can span hundreds of kilometers, a wide variety of ecosystem types can be associated with them. Although natural upland habitats can be found along the landward edge of the delta and on the levees that form alongside stream channels, the systems that characterize and dominate deltas are wetlands. Wetlands are distinguished by; (1) the presence of water, either above the surface, or in the rooting zone, (2) hydric soils that develop under anaerobic conditions, and (3) vegetation that is adapted to wet conditions. Since deltas are ecotones where rivers meet the sea, concentrations of salts in the waters associated with these wetlands can range from hypersaline (>35 ppt salt) to fresh (0 ppt salt). In general, the hydroperiod (the pattern of water levels and flooding in a wetland) and the salinity control the types of wetland communities that can be found within a given delta.

In temperate zones, salt and brackish water marshes (marshes are wetlands dominated by herbaceous vegetation) characterize the saline habitats. Freshwater tidal marshes and swamps (swamps are wetlands dominated by trees) characterize either the inland regions of the delta or those areas dominated by freshwater inputs from the active delta lobe. Between latitudes 25°N and 25°S, forests called mangrove swamps characterize the saline habitats. The term “mangrove” is not a taxonomic term, but rather describes a group of 12 genera of woody halophytes adapted to growing in intertidal regions of the tropics. Mangroves show remarkable adaptations to salt and flooding stress, but not to freezing temperatures, therefore they are replaced by salt marshes in the colder temperate zones.

Deltaic Ecology, Fig. 1 The delta cycle. (Modified from Roberts 1997)



Primary Producers and Net Primary Productivity

For several reasons, wetlands associated with deltaic systems are among the most productive in the world. First, several types of primary producers (organisms that assimilate the energy from light to synthesize organic compounds) can be found within one wetland including; photosynthetic bacteria, algae (e.g., phytoplankton, periphyton, and macro algae), bryophytes, and vascular macrophytes ranging from sea grasses to trees. Second, delta systems receive multiple subsidies such as riverine inputs that supply nutrients, sediments, and freshwater, and tides that flush plant toxins from the sediments and maintain moderately aerobic conditions. Finally, although the vascular plants that dominate most wetlands are subject to the stress of periodic flooding and salt, they are still capable of relatively high rates of production because they have evolved numerous strategies that allow them to adapt to, or avoid entirely, anaerobic stress and high concentrations of salt. These strategies include the development of specialized tissue (aerenchyma) that transport oxygen down to the roots, and root-level micro-filters that prevent salt from entering a plant.

Ecologists typically quantify the productivity of a system by measuring its net primary productivity (NPP). NPP is defined as the rate at which biomass is accumulated by photosynthesizers, after accounting for losses due to respiration. NPP is most often expressed in units of carbon, or dry weight, per some unit of time. For example, primary producers in deltaic systems can have rates of NPP exceeding $2.5 \text{ kg m}^{-2} \text{ year}^{-1}$ (above ground dry weight). This is equivalent to rates of NPP measured in tropical rain forests (Whittaker and Likens 1973). Ecologically, the concept of NPP is relevant because it represents the total amount of energy available to heterotrophs, or secondary producers, in the system, assuming no other inputs from outside the system. In deltaic wetlands, rates of NPP are controlled by (among other things), solar radiation, temperature, nutrients, salinity, and the availability of oxygen.

Secondary Production

The biomass accumulated by primary producers in a wetland can be consumed directly by either herbivores, organisms that feed on dissolved organic matter, or detritus feeders (detritivores), and indirectly by carnivores. Secondary production is the rate at which biomass is produced by these consumers of primary production. Due to the inefficiencies of energy transfer, secondary production within a marsh system is much less than primary production, unless there are significant sources of organic matter coming into the wetland from outside the system. This can be the case in some riparian wetlands, for example.

It is estimated that almost 75% of primary production in marsh ecosystems is broken down by bacteria and fungi. Early studies suggested that secondary production in saltmarsh wetland systems was driven, in large part, by these detritivores feeding upon the vascular plants growing within a marsh (Odum 1980). However, more recent research suggests that a large fraction of secondary production in salt marshes may be directly linked to fauna that consume phytoplankton and benthic algae (Kreeger and Newell 2000).

Ecological Forcing Functions

Definition

Forcing functions are the exogenous variables that drive an ecosystem. For example, the sun is the ultimate forcing function that drives primary production, but of course, it is not unique to deltaic ecosystems. As mentioned earlier, deltaic ecosystems are unique because they are driven by powerful forcing functions that can either subsidize or stress them. For example, river flooding contributes to anoxic conditions in a wetland (a stress), but also brings nutrients to the system (a subsidy). The following section outlines several types of forcing functions that characterize and define deltaic ecosystems.

Pulsed Forcing Functions

Perhaps more than most systems, deltaic ecosystems are subject to pulses of energy, ranging from daily tides to once-a-century category 5 hurricanes, that subsidize the ecosystem. These energetic events, which occur at varying timescales (Table 1) affect primary productivity, material fluxes, and

Deltaic Ecology, Table 1 Hierarchy of pulsing events in deltaic systems. (Modified from Day et al. 1995)

Pulsing event	Time scale	Impacts
Avulsion	1000+ years	New delta lobe formation
Major river floods	50–100 years	Major sediment deposition Enhanced productivity
Hurricanes/ tropical storms	5–20 years	Major sediment deposition
Seasonal river floods	Annual	Sediment deposition Nutrient inputs Decrease salinity Enhance production Organism transport
Spring tides	Bimonthly	Organism transport
Storm events (frontal passages)	Weekly	Sediment deposition Organism transport
Daily tides	1–2 cycles per day	Drainage Enhanced marsh production Flushes accumulating salt and sulfides Organism transport

the evolution of the delta itself (Day et al. 2000). Daily tides flush accumulated salts and sulfides that inhibit plant growth, thus enhancing primary productivity. Higher tides, which occur during full and new moons, allow fish to utilize interior marshes for feeding. The sustained winds associated with cold fronts can profoundly affect the hydrodynamics and associated sediment resuspension and deposition in deltaic regions. Annual river floods distribute sediments, nutrients, and freshwater to deltaic systems and have been shown to be positively correlated with primary and secondary productivity. On a longer term, major storms such as hurricanes can deposit several centimeters of sediment onto deltaic marshes and can be a critical process for maintaining a balance between rates of sea-level rise and rates of marsh accretion. Major river floods, which occur only once or twice a century are also associated with substantial sediment deposition and even the channel switching that starts a new delta cycle.

River Switching, the Ultimate Pulsing Event

The initial growth of a deltaic lobe starts as river sediments fill in areas along the coast and marine shelf. Eventually, as the delta extends seaward, the existing river becomes longer, the channel slope to the sea decreases and becomes hydraulically inefficient, and the entire river switches to a new channel. This switching event (an avulsion) often occurs abruptly, during a major flood, for example, when an upstream levee is breached and the river captures a new, shorter, and steeper course to the sea. In the final phase of the delta cycle, water and associated sediments are directed to the new channel, and the old deltaic lobe deteriorates as rates of subsidence become greater than the rate that sediments accumulate (Fig. 1). Of course while the old delta is deteriorating, a new delta, associated with the new river channel, is forming. The rate of channel switching is governed by the characteristics of the river and the receiving basin. For the Mississippi River delta, perhaps the most studied in the world, the river changes course approximately every 1000 years. Although pulsing events like the ones described above control ecosystem dynamics on the short term, it is the ecosystem's temporal position within the delta cycle that ultimately controls it.

Subsidence and Sea-Level Rise

Natural rates of subsidence in many deltaic systems can be quite high due to the compaction, consolidation, and downwarping associated with the rapid deposition of alluvial sediments. This subsidence, plus the global eustatic sea-level rise (currently $0.15 \text{ cm year}^{-1}$), defines the rate of relative sea-level rise (RSLR). Some of the highest rates of RSLR ($1.1\text{--}1.3 \text{ cm year}^{-1}$) have been recorded in the Mississippi River delta of the United States. In wetlands, high rates of RSLR can lead to increasingly long periods of inundation that are associated with decreased primary productivity and

plant mortality. Therefore, RSLR can be thought of as a forcing function that stresses, rather than subsidizes a wetland ecosystem.

Deltaic wetlands can persist in the face of RSLR when vertical accretion equals or exceeds the rate of subsidence. The rate at which a wetland accretes depends upon sediment and nutrient supply, tidal range, vegetation, and climate. Current models suggest that coastal wetlands can remain stable against a RSLR rate as high as 1.2 cm year^{-1} (Morris et al. 2002). Since this rate is equal to or higher than the current rate of RSLR in many deltaic systems, this would suggest that these systems are sustainable. However, best estimates suggests that, due to the impacts of global warming, the eustatic sea levels could rise 48 cm by the year 2100 (Church et al. 2001). Recent studies have suggested that few deltaic wetlands would be able to keep pace with this predicted increase in sea-level rise (Day et al. 1999). Exacerbating this problem, reductions in sediment delivery have led to rapid wetland loss in many deltaic regions worldwide. This is discussed further in the next section.

Anthropogenic Disturbances

Although it is technically correct to position humans within an ecosystem, it is not out of order to consider some anthropogenic actions as allogenic forcing functions. For example, as previously discussed, seasonal river flooding deposits sediments onto wetlands that contribute directly to elevation gain, and counterbalance the effects of RSLR.

However, many of the world's major rivers that are responsible for the creation of deltas in the first place have now been dammed. This has not only trapped large amounts of sediments in the reservoirs behind the dams, but has also prevented the seasonal flooding that once distributed sediments onto the delta. For example, due to upstream dams, the sediment load has been reduced by 95% in Nile, Indus, and Ebro rivers, by 75% in the Po and Mississippi rivers, and by over 50% in the Rhone river (Day et al. 2000). Additionally, because of the creation of extensive levee systems along these rivers, the remaining sediment that does reach the coast often discharges directly into deeper waters of the receiving basin, rather than accreting on the deltaic plain. Finally, many wetlands in deltaic regions have become hydrologically isolated by dense networks of canals and spoil banks constructed during the past several centuries. Spoil banks impede drainage and often physically impound wetlands, thus preventing the overland flow of any remaining sediments into coastal wetlands. As an example, because of the combined effect of high rates of RSLR and decreased sediment delivery, wetlands in the Mississippi River delta complex are currently being lost at a rate of $100 \text{ km}^2 \text{ year}^{-1}$.

While the hydrologic alternations that reduce pulses are perhaps the greatest anthropogenic forcing function threat

to deltaic ecosystems, they are by no means the only ones. Other anthropogenic disturbances include eutrophication, organic enrichment, thermal pollution, toxic inputs, over harvest, and exotic species introduction. Also, because of their position in the landscape, deltas are, and always have been, desirable places to live, farm, and conduct commerce, and humans continue to encroach on these systems.

Toward the Conservation and Restoration of Deltaic Ecosystems

Deltaic wetlands provide numerous valuable services such as fisheries production, disturbance regulation, and water quality improvement to a global population that is increasingly concentrated near the coast and dependent upon its resources (Nuttall et al. 1997). Scientists and resources managers have become aware that, to prevent the further loss of these systems, landscape scale solutions directed toward the restoration of sediment supplies are required. Some of the more successful projects have included: (1) creating gaps in the artificial levees that line most distributory streams, (2) removing sea-side dikes to restore tidal flushing, and (3) creating large-scale diversions of river water to re-introduce nutrients, freshwater, and sediments to wetlands that have been hydrologically isolated.

The Caernarvon Freshwater River Diversion project in coastal Louisiana, USA, which has been operating since 1991, is one of the best examples of this third type of restoration. The project consists of a structure containing 5 m²-gated culverts that divert water from the Mississippi River into a part of the delta that was losing approximately 400 ha year⁻¹ due to high rates of RSLR. Construction was completed in February 1991 at a cost of US \$26.1 million, and the estimated net benefit for fish, wildlife, and recreation is over US \$9 million per year. In a 1998 project summary, the Army Corps of Engineers reported a net increase in marshland of 165 ha within a monitored area that originally contained 930 ha of marsh. Oyster and fish productivity, and waterfowl usage has increased dramatically as well.

Cross-References

- [Changing Sea Levels](#)
- [Coastal Subsidence](#)
- [Dams, Effect on Coasts](#)
- [Deltas](#)
- [Dikes](#)
- [Estuaries](#)
- [Eustasy](#)
- [Mangroves, Ecology](#)

- [Salt Marsh](#)
- [Sea-Level and Climate Change](#)
- [Wetlands](#)

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Deltas

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Deltas are coastal landforms comprised of subaerial and subaqueous packages of fluvial-transported sediments that have

Shea Penland: deceased.

formed an alluvial landscape by deposition at the mouth of a river. Deltas form at the coastal interface where riverine sediment supplied to the coastline is not removed by tides or waves. The term delta derives from Herodotus who, in the fifth century BC, noted a geometric similarity between the tract of land at the mouth of the Nile River and the Greek letter “Δ” with its apex directed landward (Moore and Asquith 1971). Although this distinctive morphology is absent in many river-mouth landscapes, the term has nonetheless been accepted to describe the geographical region near a river mouth and the sedimentary package that develops at a fluvial entrance into a depositional receiving basin.

Globally, deltas can be found on all continents and in all climates (Fig. 1). In a general sense, the locations of deltas are similar; at the terminus of a catchment basin that provides sediment load into an ocean, gulf, lagoon, estuary, or lake. Although a delta may form regardless of the size of the fluvial system or receiving basin, some tectonic settings are more conducive than others to the development of major deltaic landscapes. Using the Inman and Nordstrom (1971) tectonic classification of coasts, trailing-edge coasts typically have the largest drainage basins followed successively by marginal-seacoasts and leading-edge coasts. Of the 58 major river systems in the world with drainage areas greater than 10^5 km^2 , 56.9% are on trailing edge coasts, 34.5% along marginal-seacoasts, and 8.6% on leading-edge coasts.

Trailing-edge coasts provide a geologic setting favorable to delta development because of their tectonic stability and geologically old, low-relict terrains with extensive river systems that provide an abundant supply of sediment. Moreover, trailing-edge coasts often border broad continental shelves,

which provide ideal shallow-water platforms for the formation of deltas. Marginal-seacoasts typically provide a low-energy setting with many trailing coast characteristics that are conducive to delta development. However, the mountainous and immature drainage systems of tectonically unstable, leading-edge coasts with small catchment basins and low sediment supply generally limit the likelihood of extensive deltaic deposition. Table 1 lists the sizes and locations of the some of the world’s major deltaic systems.

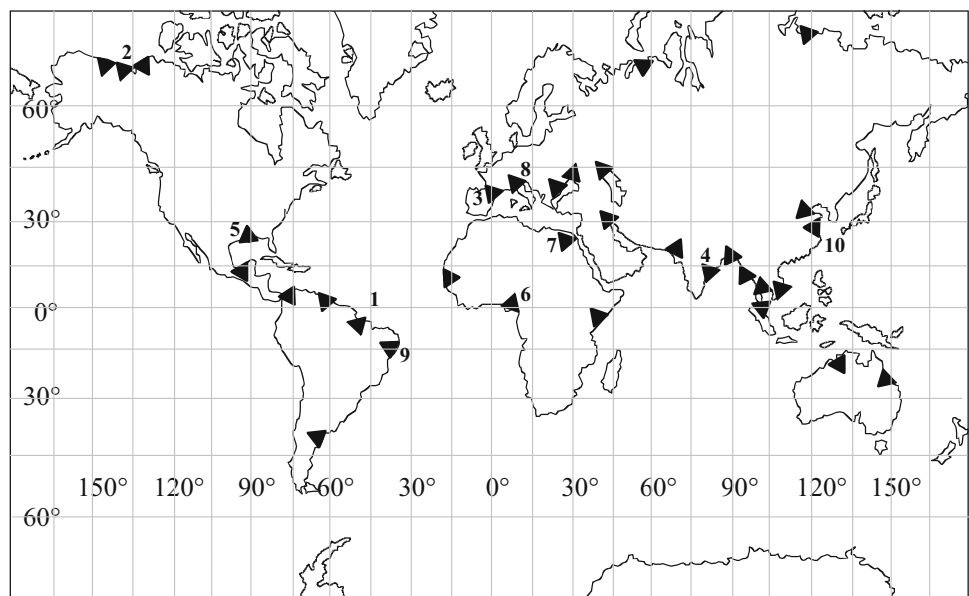
Delta Types and Formation

Previous summaries of the world’s deltas have primarily focused on the major large deltas (e.g., Coleman and Wright 1975). A review of the recent literature on late Quaternary deltaic systems reveals that deltas can be classified into four basic types depending on sea-level stage (highstand or lowstand) and/or sediment supply (high or low) (Kindinger 1988; Boyd et al. 1989; Nichol et al. 1996). The four basic types of deltas are lacustrine, bayhead (or lagoonal), continental shelf, and continental margin. In an evolutionary sense, these four types of deltas can represent a developmental continuum. Thus, during a sufficiently long sea-level stillstand a fluvial system with adequate sediment supply could result in a lacustrine deltaic system that evolves into a continental-shelf margin deltaic system through progradational outbuilding.

Lacustrine Deltas

A lacustrine delta consists of deposits located landward of the coast or paralic zone within an inland lake or sea. The

Deltas, Fig. 1 Map showing the global distribution of major deltas. Note that most of the major deltaic systems are located along trailing-edge coastlines close to the equator. (Modified from Coleman and Prior 1980). Numbered locations refer to deltas listed in Table 1



Deltas, Table 1 Names, locations, and sizes of major deltas. (From, data compiled from Coleman and Wright 1975)

River	Location	Receiving body of water	Deltaic area (km ² × 10 ³)
Amazon (1)	South America	Atlantic Ocean	467.1
Burdekin	Australia	Coral Sea	2.1
Chao Phraya	Asia	Gulf of Siam	11.3
Colville (2)	North America	Beaufort Sea	1.7
Danube	Europe	Black Sea	2.7
Dneiper	Asia	Black Sea	NA
Ebro (3)	Europe	Mediterranean Sea	0.6
Ganges–Brahmaputra (4)	Asia	Bay of Bengal	105.6
Grijalva	North America	Gulf of Mexico	17.0
Hwang Ho	Asia	Yellow Sea	36.3
Indus	Asia	Arabian Sea	29.5
Irrawaddy	Asia	Bay of Bengal	20.6
Klang	Asia	Straits of Malacca	1.8
Lena	Asia	Laptev Sea	43.6
Mackenzie	North America	Beaufort Sea	8.5
Magdalena	South America	Caribbean Sea	1.7
Mekong	Asia	South China Sea	93.8
Mississippi (5)	North America	Gulf of Mexico	28.6
Niger (6)	Africa	Gulf of Guinea	19.1
Nile (7)	Africa	Mediterranean Sea	12.5
Ord	Australia	Timor Sea	3.9
Orinoco	South America	Atlantic Ocean	20.6
Paraná	South America	Atlantic Ocean	5.4
Pechora	Europe	Barents Sea	NA
Po (8)	Europe	Adriatic Sea	13.4
Red	Asia	Gulf of Tonkin	11.9
Sagavanirktok	North America	Beaufort Sea	1.2
São Francisco (9)	South America	Atlantic Ocean	0.7
Senegal	Africa	Atlantic Ocean	4.3
Shatt-al-Arab	Asia	Persian gulf	18.5
Tana	Africa	Indian Ocean	3.7
Volga	Europe	Caspian Sea	27.2
Yangtze-kiang (10)	Asia	East China Sea	66.7

environmental setting in such cases is typically a shallow-water basin with a low-energy regime. Along the Gulf of Mexico, the best example of a lacustrine delta is the Grand Lake delta within the Atchafalaya drainage basin of southcentral Louisiana. At the start of the nineteenth century

Grand Lake within the Atchafalaya Basin was a shallow, low-energy lake (Tye and Coleman 1989). However, by the 1950s a large lacustrine delta had filled this lake and started bypassing into the Atchafalaya Bay as a result of the progressively increasing Mississippi River flow into the Atchafalaya basin (van Heerden and Roberts 1988). Historically, if left uncontrolled, the Mississippi River would have avulsed from its modern course into the Gulf of Mexico past New Orleans to the Atchafalaya course located farther west. Other lacustrine delta examples are present within Lake Geneva and Lake Constance (Switzerland) (Müller 1966; Reineck and Singh 1973).

Bayhead/Lagoonal Deltas

Bayhead/lagoonal deltas are located in the upper reaches of the coastal zone. In many cases they represent an evolutionary step from a lacustrine delta during a sea-level stillstand or withdrawal, which allows the delta to build into the coastal or paralic zone. Bayhead-delta settings are typically protected by barrier-island systems or remnants of preexisting antecedent topography, resulting in a low-energy environment. Examples of bayhead deltas are found within the northern part of Mobile Bay Alabama (Kindinger et al. 1994), Trinity Bay, and San Antonio Bay Texas; (McEwen 1969; Donaldson et al. 1970; van Heerden and Roberts 1988).

Continental-shelf Deltas

Continental-shelf deltas develop during prolonged sea-level stillstands and/or a sea-level fall. These deltas build out onto the continental shelf. Trailing edge or marginal sea continental-shelf deltas typically are characterized by an abundant sediment supply and their geomorphology reflects the wave, riverine, and tidal conditions in which they develop. Examples of continental-shelf deltas include the Mississippi (US, Louisiana), the Ebro (Spain), and the Colville (US, Alaska) (Frazier 1967; Maldonado 1975; Naidu and Mowatt 1975; Alonso et al. 1990).

Continental-shelf Margin Deltas

Continental-shelf margin deltas are typically most well-developed during sea-level lowstands or formed by rivers with very large sediment loads during prolonged sea-level stillstands. The majority of shelf-margin deltas in the recent geologic record have developed during falling sea-level or sea-level lowstand conditions as river valleys lengthen, incise across subaerially exposed continental-shelf sediments, and discharge their sedimentary load at the shelf margin (Suter 1994). With sufficient sediment load and/or during lowstand conditions these deltas continue to build seaward, creating an apron of deltaic sediments along the shelf edge that extends the preexisting continental margin into deeper basinal waters. Numerous lowstand shelf-margin deltas, formed during the late Wisconsin sea-level fall and lowstand that culminated at

approximately 18,000 year BP, are found seaward of the Rio Grand and the Mississippi River deltas (Morton and Price 1987; Suter et al. 1987; Suter 1994). Sufficiently large rivers carrying substantial sediment loads may however, construct shelf-margin delta deposits during sea-level highstand conditions. An example of a shelf-margin highstand delta is the depocenter of the main stem of the modern Mississippi River delta, the representative “Bird Foot” delta (Boyd et al. 1989).

Delta Processes

Deltas result when sediment-laden fluvial water decelerates upon entering a receiving basin and sediment transport competence of the river is lost. Consequently, fluvial characteristics and river-mouth processes constitute an important controlling factor in the nature of deltaic deposition (Bates 1953). A consideration of fluvial processes is particularly important in deltaic environments where receiving basin processes such as wave and tide regimes are minimal. In such environments the primary fluvial factors, influencing the nature of deltaic deposition, include the volume of riverine discharge and transported sediment, as well as the textural character of the sediment load. Generally, the forces associated with these factors that dictate the character of fluvial-margin deposition are: (1) the riverine effluent inertia and diffusion, (2) friction between the sediment load and the floor of the receiving basin, and (3) density contrasts between the effluent and the receiving basin waters (Bates 1953; Wright 1977; Orton and Reading 1993).

Homopycnal Conditions

Homopycnal flow describes river discharges where the density contrast between riverine and basinal waters is small. Discharge patterns are most heavily influenced by inertia of the riverine water, which allows the effluent to radially spread into the receiving basin. As a result of deltaic progradation, fine-grained bottomset sediments are overlain by relatively coarser grained topset beds and river mouth-bar deposits, creating the classic coarsening upward depositional units common to deltaic systems. Gilbert (1884) first described this type of deltaic system in lakes of the western United States; hence, they are termed Gilbert-type deltas.

Hyperpycnal Conditions

Hyperpycnal discharge conditions occur when the riverine inflow is denser than the waters of the receiving basin. Consequently, the sediment-laden effluent moves along the receiving basin bottom as a density current. This type of discharge process is rare and typically restricted to deltaic systems rich in silt and coarse-grained sediments such as in the Bella Coola fjord delta of British Columbia (Kostaschuk 1985).

Hypopycnal Conditions

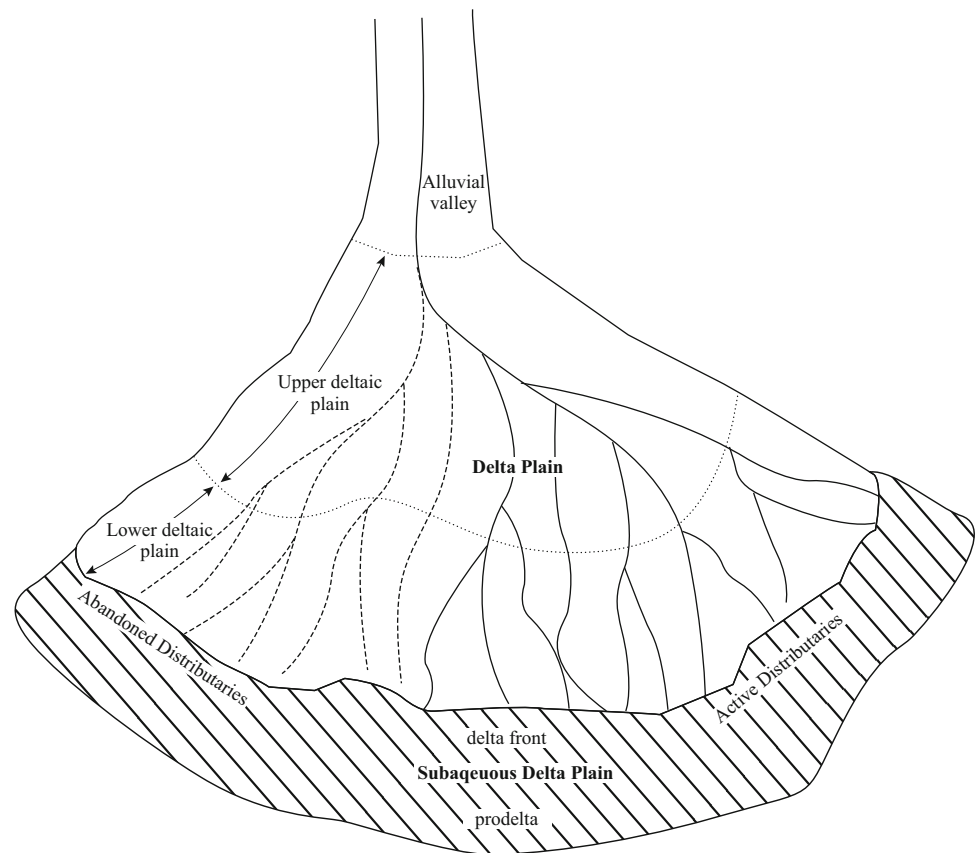
Hypopycnal discharge conditions occur when the riverine inflow is less dense than the waters of the receiving basin, a condition commonly present in lakes and inland seas. Most of the world's deltas form under hypopycnal conditions. In hypopycnal settings, as the riverine inflow enters the receiving basin it undergoes turbulent mixing resulting in the deposition of the sediment load. Finer-grained sediments are transported well into the receiving basin as the freshwater sediment plume disperses above relatively more dense basinal waters, whereas the relatively coarser grained sediments are deposited in close proximity to the river mouth. In shallow-water hypopycnal discharge conditions, friction between the receiving basin bottom and the riverine leads to multiple bifurcations of the distributary system. Conversely, in deep water during hypopycnal discharge conditions, buoyancy dominates during the dispersal of the effluent, producing elongated distributaries such as present at the modern Mississippi River depocenter (Suter 1994).

Delta Components

The delta plain is the subaerial and subaqueous zone where sediments are discharged from the alluvial valley and accumulate in the receiving basin. Typically, a delta plain is subdivided into an upper delta plain and a lower delta plain, each of which are characterized by different vegetation, morphology, and depositional processes (Fig. 2) (Reineck and Singh 1973; Coleman 1981; Elliot 1986). Many of the features of the upper delta plain are inherited from its alluvial valley. In the upper delta plain deposition is primarily fluvial and as the main stem of the river leaves the alluvial valley it discharges onto the upper delta plain and undergoes channel bifurcation. The vegetative landscape of the upper delta plain is a freshwater environment, generally beyond the influence of marine incursions created by the tidal regime of the receiving basin. The major landform includes distributary channels with their subaerial natural levees, point bars, and crevasse splays, as well as interdistributary environments characterized by low-lying swampy areas.

Seaward of the upper delta plain is the relatively lower relief of the lower delta plain that is within the realm of tidal incursion. The lower delta plain is the site of more active deposition through overbank flooding processes. Because the elevations of the natural levees are lower, channel crevassing and distributary shifting is a common and very dynamic process characterizing the lower delta plain. These processes disperse sediment from areas within the upper delta plain into brackish to saline environments of the lower delta plain. Distributary shifting common in the lower delta plain results in a complex distribution of multiaged distributary channels in all stages of evolution (Fig. 2). The lower delta plain

Deltas, Fig. 2 Schematic diagram showing the primary physiographic components and depositional environment of a delta. (Modified from Coleman and Prior 1980)



contains active and abandoned distributary channels in all stages of regressive and transgressive evolution (Penland and Suter 1989). Because the lower delta plain is a transitional environment influenced by fresh-water fluvial and saline marine waters the vegetative landscape generally contains species tolerant of both brackish and saline water.

Deltaic Sediments

In general, deltaic sedimentary packages contain a coarsening upward vertical stratigraphy that reflects seaward progradation of the delta into its receiving basin. At any given time there is typically a well-developed gradient of grain sizes extending seaward from the subaerial to subaqueous portions of a delta. The relatively coarsest grained material is deposited proximal to the river mouth where transport competence is highest, whereas the finest-grained material is carried farther seaward and deposited in more distal locations. Thus, as sediments are delivered to the receiving basin from the river mouth they accumulate in a subaqueous, fine-grained depositional zone referred to as the *prodelta*. The prodelta forms the platform across which deltas prograde and subsequently aggrade. The prodelta sedimentary package is a widespread laterally continuous interval primarily composed of the finest

sediment fraction transported by the fluvial system. Thus, the overall finest sediment is found at the base of the prodelta sequence and the whole sequence coarsens upwards. Moreover, the initial prodelta deposits are the thinnest at the base and the beds thicken upward as a result of deltaic progradation. Bioturbation of prodelta sediments is limited by the rate of deposition.

The delta front is located between the zone of fine-grained prodelta deposition and the more landward located, coarser-grained distributary mouth deposits of a progradational deltaic system. The relatively coarser-grained sediments of the delta front generally consist of interbedded clays, silts, and sands. This zone is dominated by the interaction of fluvial and marine processes, resulting in sand-rich accumulations landward of the advancing prodelta. The delta front also represents a zone of transition between deposits representative of progradation and aggradation.

In a progradational deltaic sequence, distributary channel deposits overlie the relatively finer grained delta front and prodelta deposits. The framework of distributary channels consists of distributary mouth bars overlain by natural levee deposits. Distributary mouth bar deposits are primarily subaqueous deposits that grade laterally into relatively finer grained deposits; locally, mouth-bar deposits may contain fine-grained beds within the generally sandy matrix of the

mouth bar. The presence of fine-grained deposits primarily represents deposition during low-flow conditions. During flood conditions, natural levee ridges can be over-topped and breached creating crevasse splays. These splays within the distributary network provide conduits for sediment dispersal into interdistributary bays and the origin of crevasse-splay deposits.

Interdistributary bay and marsh sediments are generally as volumetrically significant as prodelta sediments. The interdistributary area is a low-energy depositional environment and less dynamic than areas of the delta plain characterized by multiple channels, channel bifurcations, and channel avulsions. Interdistributary deposits characteristically consist of fining upward deposits consisting of clay-rich bay sediments overlain by organic-rich marsh deposits. This fine-grained depositional environment is punctuated by sandy depositional events associated with overbank flooding and crevasse splay events.

Delta Morphology

Although sediment delivery to a delta is by a fluvial system, it is ultimately the dynamic interaction of riverine and basinal processes that control the morphologic, stratigraphic, and sedimentologic variability of a deltaic accumulation. Water depth and basinal configuration, tidal range, wave climate, and coastal currents are the primary basinal processes that control delta morphology. The dynamics of these processes and complex interaction between them leads to a highly variable array of deltaic configurations (Coleman and Wright 1975).

One of the first and simplest attempts to describe deltaic variability as a function of processes involved depicts the morphology of river deltas as a function of sediment influx and marine processes such as waves and longshore currents (Fig. 3). The tidal regime of the receiving basin was not

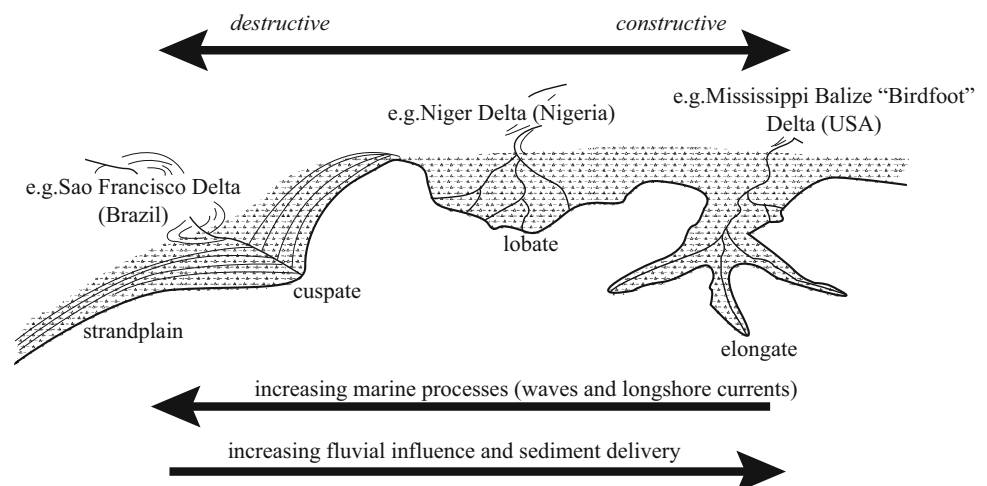
considered in formulation of this model. Fundamentally, this model describes the processes leading to deltaic deposition as either constructive or destructive (Davis 1983). Deltas dominantly influenced by the nearshore wave climate are termed destructional, characteristically cusped shaped with poorly developed distributaries and well-developed strand plains such as in the Sao Francisco delta of Brazil. Intermediate between destructional and constructional forms, is the lobate-shaped delta, typified by the Niger delta of Africa. Lobate deltas possess well-developed distributary networks and a smooth coastal outline. Deltas that are strongly influenced by sediment influx are essentially constructional and are characterized by distinctive elongate distributaries. An example of this type of delta is the modern “Birdfoot” depocenter of the Mississippi River.

In the 1970s, new process-response approaches to understanding morphologic variability of deltas incorporated features of the constructional/destructional approach but also considered the effects of the receiving basin tidal regime. In the Galloway (1975) scheme a variety of previously documented delta morphologies (e.g., Fisher et al. 1969; Wright and Coleman 1973) were classified within the framework of a ternary diagram that effectively differentiated delta morphology as a function of sediment input, wave-energy flux, and tidal energy flux (Fig. 4).

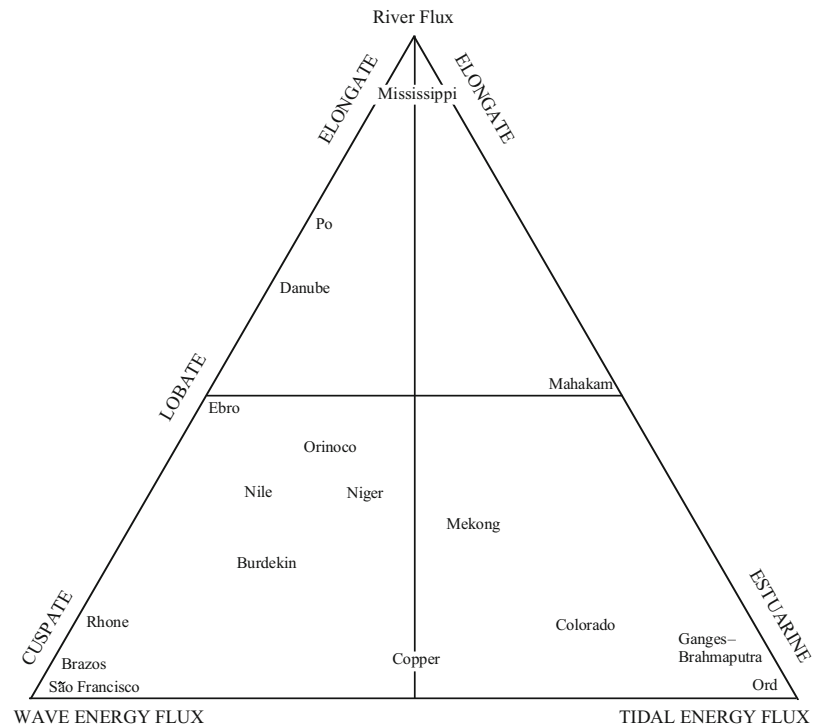
Deltas dominated by sediment input are supplied by a large well-developed drainage system, capable of transporting large volumes of sediment to the deltaic coastline. Distributary switching is an important process, resulting in a highly constructive elongate to lobate deltas with straight to sinuous active and abandoned channels. The bulk composition of the sediment supply is muddy to mixed and the framework facies include distributary mouth bar, channel-fill sands, and delta-margin sheet sands (Table 2).

Deltas shaped primarily by waves are usually characterized by a small catchment basin with little sediment influx. The high-energy wave environment leads to winnowing

Deltas, Fig. 3 Schematic diagram showing the variation in deltaic geomorphology as a function of the relative influences of fluvial and marine processes. (Fisher et al. 1969, with permission of University of Texas, Bureau of Economic Geology)



Deltas, Fig. 4 Ternary classification of deltas on the basis of dominate processes and the resulting morphology. (After Galloway 1975)



of fine-grained sediments and leaves behind a relatively coarser-grained sediment fraction. Consequently, the bulk composition of wave-dominated deltas is typically sandy. The primary distributary channels are meandering and the framework facies are coastal barrier and beach-ridge sands (Table 2).

Deltas shaped primarily by tides are similar to wave-dominated deltas in that tidal currents winnow away much of the fine-grained sediments delivered to the coast. However, a fundamental difference is that tidally formed sand bodies are oriented shore normal. The distributary channels flare seaward and in some cases are straight and oriented parallel with the tides. The bulk composition of tide-dominated deltas is highly variable. The primary framework facies include fine-grained estuarine fill and coarse-grained sand ridges (Table 2).

Modern Deltas

Globally, modern deltaic systems are geologically young features that have formed since the initiation of late Wisconsin deglaciation at approximately 18,000 year BP. Although the development, existence, and evolution of deltas along modern coastlines is contingent upon sea-level change, sediment delivery, and marine processes, the timing of formation for modern deltas suggests that sea level has been a primary factor controlling delta development and stability (Stanley and Warne 1994).

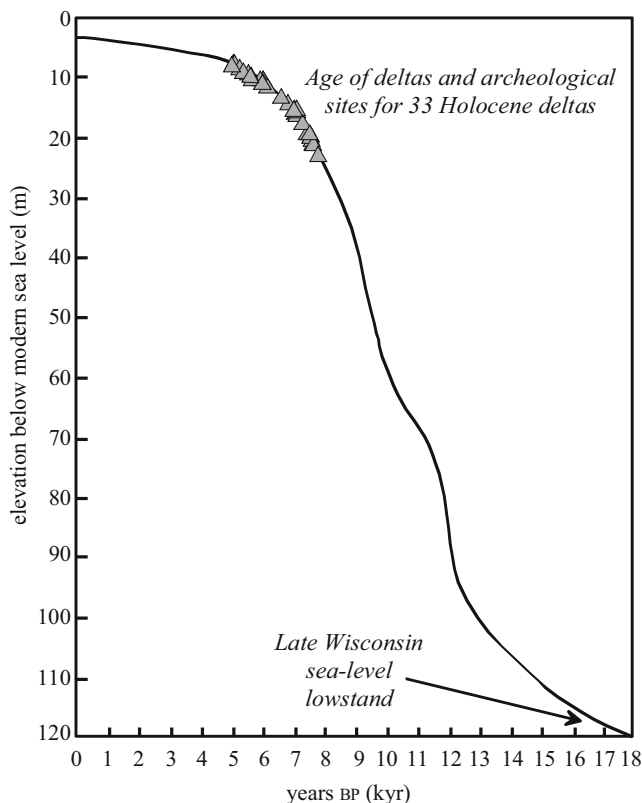
Deltas, Table 2 Morphologic, stratigraphic, and sedimentologic characteristics of delta depositional systems that result in fluvial, wave, and tide-dominated systems. (Modified from Galloway 1975)

Characteristics	Fluvial-dominated	Wave-dominated	Tide-dominated
Geomorphology Channels	Elongate to lobate Straight to sinuous distributaries	Arcuate Meandering distributaries	Estuarine to irregular Flaring straight to sinuous distributaries
Sediments Framework facies	Muddy to mixed Distributary mouth bar and channel fill sands, delta margin sand sheet	Sandy Coastal barrier and beach-ridge sands	Variable Estuary fill and tidal sand ridges
Framework geometry	Parallels depositional Slope	Parallels depositional strike	Parallels depositional slope

At the height of the late Wisconsin glacial period approximately 18,000 year BP, sea level was approximately 120 m lower than present sea level (Fairbanks 1989). During this interval of lower sea level rivers crossed subaerially exposed continental shelves within incised channel systems. Large volumes of sediment were debouched along bordering continental margins and locally, thick and extensive shelf-margin deltas developed (Morton and Price 1987; Suter et al. 1987). Subsequently, as the late Wisconsin glaciers melted and sea

level rose, exposed shelves were flooded and transgressed with marine waters. Initially, sea level rose so rapidly that the global paralic zone was forced tens of kilometers landward. Marine processes at the leading edge of the transgressing marine waters reworked the formerly exposed continental shelf leaving a marine truncation surface and transgressive coastal-zone deposits in the form of ravinement surfaces and transgressive coastal marine deposits, inner shelf shoals, sand sheets, and shoreface ridges.

Following the initially high eustatic rise rates associated with deglaciation the rate of sea-level rise dropped below a critical threshold rise rate allowing global landward migration of the paralic zone. Beginning approximately 7000 year BP, the rate of sea-level rise appears to have slowed enough to allow for deltaic progradation and a relatively synchronous development of the world's deltas (Fig. 5). The temporal uniformity in delta development suggests a decline in the rate of sea-level rise, following the melting of late Wisconsin ice sheets, strongly influenced Holocene deltaic formation (Stanley and Warne 1997).



Deltas, Fig. 5 Graph showing the time of formation for latest Quaternary deltaic systems and ages of deltaic archeological sites in relation to deceleration in the rate of sea-level rise following the late Wisconsin sea-level lowstand. (Ages from Stanley and Warne 1997; sea-level curve from Fairbanks 1989)

Deltas and Society

Deltaic environments have played an important role in our global society. Deltaic environments served as the culture hearth for early civilizations throughout the world because of their tremendous variety of food resources such as fish and wildlife. Moreover, the rich alluvial soils of deltaic landscapes allowed for the establishment of bountiful crops vital to the establishment and expansion of early cultures (Stanley and Warne 1997). Today, deltaic wetlands that were more recently viewed as uninhabitable wasteland are once again considered to be of utmost ecological value. Additionally, ancient deltaic deposits are actively explored as they often contain a wealth of fossil fuels such as hydrocarbons and coal, resources that helped spur the industrial revolution and continue to be critically important to our modern technologically driven society.

However, as humans have populated the world's deltas, harnessed the rivers that built them by flood-control structures, and altered natural deltaic coastlines the delicate balance of these dynamic systems has been compromised. Deprived of water and sediment, most of the world's deltas are undergoing dramatic coastal land loss and habitat change as a result of anthropogenic induced as well as naturally occurring subsidence and rising sea level along coastal zones. As a result, a new multidisciplinary field of coastal restoration science is emerging as humans embark to rescue and effectively manage deltaic landscapes. The future of modern deltas will be dependent on our ability to balance environmental protection and economic development, while developing new restoration technologies for their maintenance and restoration.

Cross-References

- [Changing Sea Levels](#)
- [Coastal Sedimentary Facies](#)
- [Deltaic Ecology](#)
- [Ingression, Regression, and Transgression](#)
- [Late Quaternary Marine Transgression](#)
- [Sea-Level Fluctuations Over the Last Millennium](#)
- [Sea-Level and Climate Change](#)
- [Sedimentary Basins](#)
- [Sequence Stratigraphy](#)
- [Submerging Coasts](#)
- [Wetlands](#)

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Demography of Coastal Populations

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Definition

Demography of coastal human populations attempts to quantify the current and future number of residents living within 100 km of the shore country by country, and region by region.

Introduction

To most observers who live along a coast it seems clear that the resident human population is growing rapidly, but how fast is population growing? Will future growth be concentrated in the coastal zone? Do such demographics portend well for sustainable development?

To answer these questions we have combined the recent work of Burke et al. (2001), who reported the fraction of a country's population living within 100 km of the coast, with United Nations (2017) population figures to calculate the 2020 and 2035 coastal populations for 129 countries. Burke et al. relied on the work of the Center for International Earth Science Information Network (CIESIN, see reference in Burke et al.) at Columbia University, to obtain the coastal population fraction. In our work, we did not attempt to reanalyze the work of CIESIN. Rather country-by-country we multiplied the fractional coastal population reported in Burke et al. by United Nations 2020 and 2035 population values (United Nations 2017).

The United Nations based its calculated demographic projections for each country on birth, death, fertility rates, and age structure of the population. We assumed that the 2035 coastal population fractions would remain unchanged from the 2020 values reported by Burke et al. (2001). It is likely, however, that for some countries, such as the United States, coastal populations will grow faster than noncoastal areas (Culliton et al. 1990), as aging populations will live longer and will seek to live near marine environments. Thus the 2025 populations reported herein should be considered minimum values.

Coastal Properties and Demographics by Country and Region

Table 1 lists the results for the 129 countries (column 1) in thousands of people. The countries are grouped within eight geographic regions (Burke et al. 2001), with each country listed beneath the region. Coastal lengths and claimed Exclusive Economic Zone are from Burke et al. (2001), and the total population values (column 4) are from United Nations (2017). Each country's demographics are broken down into the 2020 and 2035 total populations, the 2020 and 2035 coastal populations (persons living within 100 km of the coast), the difference between 2020 and 2035 coastal populations (column 9), and the relative increase or decrease over the 2020 population (%). The footnotes at the base of Table 1 specify the calculations.

Surprisingly India not China has the largest coastal population, 364 and 411 million people for 2020 and 2035, respectively, or 13% growth in 15 years. Bulgaria, Latvia, and Lebanon will experience 11%, negative growth in their coastal population from 2020 through 2035. Fifteen other countries in Europe plus Japan will experience decreasing coastal populations 2035. All in all, only 23 countries are projected to have declining coastal populations in the next two decades or so.

Figures 1–4 show the overall 2020 and 2035 demographic trends for the eight geographic regions. Figure 1 shows the

total *coastal population* estimated for the year 2020, and that projected for 2035 in millions of people. Figs. 2 and 3 compare overall total population, by region, with the corresponding total *coastal population*, for 2020 and 2035, respectively. In Fig. 4 we present the *population increase per kilometer of coastline* for the eight regions summarized in Table 1.

Middle East and North Africa and Sub-Saharan Africa coastal populations will grow by 18% and 42%, respectively, from 2020 to 2035 (Fig. 1). Combined, these changes represent an addition of approximately 143 million people living in the coastal area by 2035 over the 2020 coastal population of approximately 496 million people.

The Asian region (which excludes the Middle East), which includes heavily populated China, India, and Indonesia, will experience relatively smaller percentage growth: 8%, by 2035. However, in absolute terms the Asian coastal region will be home to 1.7 billion people in 2035, an increase of 135 million over 2020.

North America, Central America and the Caribbean Islands, South America, and Oceania will experience increased growth by 10%, 13%, 22%, and 18%, respectively, or an increase of 81 million people by 2035; the calculated total coastal population in 2035 for these four regions is 572 billion (Table 1).

Europe is the only region where coastal populations may decrease rather than increase. In 2020 the European coastal population is estimated at 308 million people (Table 1); by 2035 we calculate that the coastal population to be 311 million, an increase of only 3 million people. The reason for the small increase in coastal population (1%) is due to population stability (zero growth rate) or in some cases negative population growth rates.

Figure 4 shows the 2020 and 2035 coastal populations normalized to coastal length, a factor that can indicate “environmental stress” (Culliton et al. 1990). By 2035 the regions of Asia, Middle East/North Africa, and Sub-Saharan Africa will contain more than 5000 persons per km of coastal length. The impacts of such a high density of people along a coast will include marine pollution, over-fishing, and loss of natural habitat.

Consequence of Increasing Coastal Populations

Much of humanity lives in the coastal zone, and probably has done so for millennia. That a 12% increase in coastal population, compared with a 13% increase overall by 2035, should be no surprise, given the uncertainty of such calculations. We estimate then that globally, the coastal populations will grow slightly faster than the overall population, but that the difference is not statistically significant. Clearly, however, Sub-Saharan Africa will grow their coastal populations much

Demography of Coastal Populations, Table 1 Coastal properties and demographics by country and region

	Coastal length ^a	Claimed exclusive economic zone ^b	Total 2020 population ^c	Population within 100 km from coast ^d	2020 population within 100 km from coast ^e	Total 2035 population ^f	2035 population within 100 km from coast ^g	Population change, 2020–2035, within 100 km from coast ^h	Increase
	km	000 km ²	000 people	%	000 people	000 people	000 people	000 people	% ⁱ
1. Asia									
Azerbaijan	871		10,100	55.7	5626	10,861	6050	424	8
Bangladesh	3306	39.9	169,775	54.8	93,037	191,601	104,997	11,960	13
Cambodia	1127		16,716	23.8	3978	19,724	4694	716	18
China	30,017		1,424,548	24	341,892	1,433,509	344,042	2151	1
Georgia	376	18.9	3899	38.8	1513	3663	1421	–91	–6
India	17,181	2103.4	1,383,198	26.3	363,781	1,564,570	411,482	47,701	13
Indonesia	95,181	2915	272,223	95.9	261,062	304,759	292,264	31,202	12
Japan	29,020	3648.4	126,496	96.3	121,815	118,500	114,115	–7700	–6
Kazakhstan	4528		18,777	3.6	676	20,950	754	78	12
Korea, Dem People's Rep	4009	72.8	25,841	92	23,774	26,972	24,814	1041	4
Korea, Rep	12,478	202.6	51,507	100	51,507	52,806	52,806	1299	3
Malaysia	9323	198.2	32,869	98	32,212	38,381	37,613	5401	17
Myanmar (Burma)	14,708	358.5	54,808	49	26,856	60,431	29,611	2755	10
Pakistan	2599	201.5	208,362	9.1	18,961	261,093	23,760	4799	25
Philippines	33,900	293.8	109,703	100	109,703	132,668	132,668	22964.225	21
Singapore	268		5935	100	5935	6480	6480	544.807	9
Sri Lanka	2825	500.8	21,084	100	21,084	21,492	21,492	407.848	2
Thailand	7066	176.5	69,411	38.7	26,862	69,200	26,780	–82	0
Turkmenistan	1289		6031	8.1	489	7073	573	84	17
Uzbekistan	1707		33,236	2.6	864	38,059	990	125	15
Vietnam	11,409	237.8	98,360	82.8	81,442	108,988	90,242	8800	11
Sums and Increase, %ⁱ	283,188		4,142,879		1,593,068	4,491,779	1,727,648	134,580	8
2. Europe									
Albania	649		2942	97.1	2857	2895	2811	–45	–2
Belgium	76		11,620	83	9645	12,160	10,093	448	5
Bosnia and Herzegovina	23		3498	46.6	1630	3336	1555	–75	–5
Bulgaria	457	25.7	6941	29.2	2027	6157	1798	–229	–11
Croatia	5663		4116	37.9	1560	3790	1436	–123	–8
Denmark	5316	80.4	5797	100	5797	6118	6118	321	6
Estonia	2956	11.6	1301	85.9	1117	1225	1052	–65	–6
Finland	31,119		5580	72.8	4062	5785	4212	149	4
France	7330	706.4	65,721	39.6	26,026	68,861	27,269	1243	5
Germany	3624	37.4	82,540	14.6	12,051	81,730	11,933	–118	–1
Greece	15,147		11,103	99.2	11,014	10,624	10,539	–475	–4
Iceland	8506	678.7	343	99.9	343	374	374	31	9
Ireland	6437		4888	99.9	4883	5374	5369	486	10
Italy	9226		59,132	79.1	46,773	57,534	45,510	–1264	–3
Latvia	565	15.6	1893	75.2	1424	1682	1265	–159	–11
Lithuania	258	3.6	2852	22.9	653	2639	604	–49	–7
Montenegro	293	87.5	629	8.1	51	619	50	–1	–2
The Netherlands	1914		17,181	93.4	16,047	17,694	16,526	479	3

(continued)

Demography of Coastal Populations, Table 1 (continued)

	Coastal length ^a	Claimed exclusive economic zone ^b	Total 2020 population ^c	Population within 100 km from coast ^d	2020 population within 100 km from coast ^e	Total 2035 population ^f	2035 population within 100 km from coast ^g	Population change, 2020–2035, within 100 km from coast ^h	Increase % ⁱ
	km	000 km ²	000 people	%	000 people	000 people	000 people	000 people	% ⁱ
Norway	53,199	1095.1	5450	95.4	5199	6193	5908	709	14
Poland	1032	19.4	37,942	13.5	5122	35,692	4818	–304	–6
Portugal	2830	1656.4	10,218	92.7	9472	9696	8988	–484	–5
Romania	696	18	19,388	6.3	1221	17,974	1132	–89	–7
Russian Federation	110,310	6255.8	143,787	14.9	21,424	138,076	20,573	–851	–4
Slovenia	41		2082	60.6	1262	2035	1233	–28	–2
Spain	7268	683.2	46,459	67.9	31,546	45,861	31,139	–406	–1
Sweden	26,384	73.2	10,122	87.7	8877	10,942	9597	720	8
Ukraine	4953	86.4	43,579	20.9	9108	39,896	8338	–770	–8
United Kingdom	19,717		67,334	98.6	66,392	71,897	70,891	4499	7
Sums and Increase, %ⁱ	325,989		674,440		307,582	666,862	311,133	3550	1.15
3. Middle East and North Africa									
Algeria	1557		43,333	68.8	29,813	51,070	35,136	5323	18
Egypt	5898	185.3	102,941	53.1	54,662	128,264	68,108	13,446	25
Iran	5890	129.7	83,587	23.9	19,977	90,479	21,624	1647	8
Iraq	105	0.7	41,503	5.7	2366	59,821	3410	1044	44
Israel	205		8714	96.6	8417	10,628	10,267	1849	22
Jordan	27		10,209	29	2961	11,841	3434	473	16
Kuwait	756		4303	100	4303	5111	5111	808	19
Lebanon	294		6020	100	6020	5336	5336	–684	–11
Libya	2025	222.4	6662	78.7	5243	7604	5984	741	14
Morocco	2008	328.4	37,071	65.1	24,133	42,407	27,607	3474	14
Oman	2809	487.4	5150	88.5	4557	6129	5424	866	19
Saudi Arabia	7572		34,710	30.2	10,482	41,317	12,478	1996	19
Syrian Arab Republic	212		18,924	34.5	6529	28,885	9965	3436	53
Tunisia	1027		11,903	84	9999	13,161	11,055	1057	11
Turkey	8140	176.6	83,836	57.5	48,206	90,915	52,276	4071	8
United Arab Emirates	2871	21.2	9813	84.9	8331	11,642	9884	1553	19
Yemen	3149	465	30,245	63.5	19,206	39,962	25,376	6170	32
Sums and Increase, %ⁱ	44,545		538,924		265,205	644,573	312,476	47,271	18
4. Sub-Saharan Africa									
Angola	2252		32,827	29.4	9651	51,665	15,189	5538	57
Benin	153		12,123	62.4	7565	17,568	10,963	3398	45
Cameroon	1799	10.9	25,958	21.9	5685	36,884	8078	2393	42
Congo	205		5687	24.5	1393	8277	2028	635	46
Congo, Dem Rep	177		89,505	2.7	2417	138,153	3730	1313	54
Cote d'Ivoire	797	157.4	26,172	39.7	10,390	37,411	14,852	4462	43
Equatorial Guinea	603	291.4	1406	72.3	1017	2105	1522	505	50
Eritrea (Red Sea)	3446		5432	73.5	3993	7421	5454	1462	37

(continued)

Demography of Coastal Populations, Table 1 (continued)

	Coastal length ^a	Claimed exclusive economic zone ^b	Total 2020 population ^c	Population within 100 km from coast ^d	2020 population within 100 km from coast ^e	Total 2035 population ^f	2035 population within 100 km from coast ^g	Population change, 2020–2035, within 100 km from coast ^h	Increase
	km	000 km ²	000 people	%	000 people	000 people	000 people	000 people	% ⁱ
Gabon	2019	180.7	2151	62.8	1351	2821	1771	420	31
Gambia	503	20.5	2293	90.8	2082	3383	3072	990	48
Ghana	758	216.9	30,734	42.5	13,062	40,719	17,306	4244	32
Guinea	1614	97	13,751	40.9	5624	19,789	8094	2469	44
Guinea-Bissau	3176	86.7	2001	94.6	1893	2755	2607	714	38
Kenya	1586	104.1	53,492	7.6	4065	74,086	5631	1565	39
Liberia	842		5104	57.9	2955	7271	4210	1255	42
Madagascar	9935	1079.7	27,691	55.1	15,258	39,891	21,980	6723	44
Mauritania	1268	141.3	4784	39.6	1894	6766	2679	785	41
Mozambique	6942	493.7	32,309	59	19,062	48,242	28,463	9400	49
Namibia	1754	536.8	2697	4.7	127	3521	165	39	31
Nigeria	3122	164.1	206,153	25.7	52,981	297,323	76,412	23,431	44
Senegal	1327	147.2	17,200	83.2	14,311	24,861	20,685	6374	45
Sierra Leone	1677		8047	54.7	4402	10,568	5780	1379	31
Somalia	3898		16,105	54.8	8826	24,700	13,535	4710	53
South Africa	3751		58,721	38.9	22,843	66,880	26,016	3174	14
Sudan	2245		43,541	2.8	1219	60,996	1708	489	40
Tanzania, United Rep	3461	204.3	62,775	21.1	13,245	95,862	20,227	6981	53
Togo	53	10.8	8384	44.6	3739	11,663	5202	1462	39
Sums and Increase, %ⁱ	59,363		797,043		231,050	1,141,582	327,359	96,309	42
5. North America									
Canada	265,523	3006.2	37,603	23.9	8987	41,888	10,011	1024	11
United States	133,312	8078.2	331,432	43.3	143,510	365,034	158,060	14,550	10
Sums and Increase, %ⁱ	398,835		369,035		152,497	406,922	168,071	15,574	10
6. Central America and Caribbean									
Belize	1996	12.8	398	100	398	507	507	109	27
Costa Rica	2069	542.1	5044	100	5044	5554	5554	510	10
Cuba	14,519	222.2	11,495	100	11,495	11,411	11,411	–85	–1
Dominican Rep	1612	246.5	11,108	100	11,108	12,497	12,497	1389	12
El Salvador	756		6479	98.8	6401	6887	6804	403	6
Guatemala	445	104.5	17,911	61.2	10,961	22,772	13,937	2975	27
Haiti	1977	86.4	11,371	99.6	11,326	13,036	12,984	1658	15
Honduras	1878	201.2	9719	65.5	6366	11,779	7715	1349	21
Jamaica	895	234.8	2913	100	2913	2908	2908	–5	0
Mexico	23,761	2997.7	133,870	28.7	38,421	153,061	43,928	5508	14
Nicaragua	1916		6417	71.6	4594	7318	5240	646	14
Panama	5637	274.6	4289	100	4289	5151	5151	862	20
Trinidad and Tobago	704	60.7	1378	100	1378	1361	1361	–17	–1
Sums and Increase, %ⁱ	58,165		222,393		114,696	254,241	129,996	15,301	13

(continued)

Demography of Coastal Populations, Table 1 (continued)

	Coastal length ^a	Claimed exclusive economic zone ^b	Total 2020 population ^c	Population within 100 km from coast ^d	2020 population within 100 km from coast ^e	Total 2035 population ^f	2035 population within 100 km from coast ^g	Population change, 2020–2035, within 100 km from coast ^h	Increase % ⁱ
	km	000 km ²	000 people	%	000 people	000 people	000 people	000 people	% ⁱ
7. South America									
Argentina	8397	925.4	45,510	45.1	20,525	51,028	23,014	2488	12
Brazil	33,379	344.5	213,863	48.6	103,937	229,203	111,392	7455	7
Chile	78,563	3415.9	18,473	81.5	15,055	20,059	16,348	1293	9
Colombia	5874	706.1	50,220	29.9	15,016	54,055	16,162	1146	8
Ecuador	4597		17,336	60.5	10,488	20,554	12,435	1947	19
Guyana	1154	122	791	76.6	606	833	638	33	5
Peru	3362		33,312	57.2	19,055	38,299	21,907	2852	15
Suriname	620	119.1	578	87	503	630	548	46	9
Uruguay	1096	110.5	3494	78.5	2743	3629	2848	105	4
Venezuela	6762	385.7	3629	73.1	2652	38,266	27,972	25,320	955
Sums and Increase, %ⁱ	143,804		387,206		190,580	456,555	233,266	42,686	22
8. Oceania									
Australia	66,530	6664.1	25,398	89.8	22,808	29,526	26,515	3707	16
Fiji	4637	1055	925	99.9	924	985	984	60	7
New Zealand	17,209	3887.4	4834	100	4834	5370	5370	535	11
Papua New Guinea	20,197	1613.8	8756	61.2	5358	11,362	6953	1595	30
Solomon Islands	9880	1377.1	647	100	647	839	839	191	30
Sums and Increase, %ⁱ	118,453		40,560		34,572	48,082	40,661	6089	18
Grand Totals^j			7,131,919		2,854,678	8,062,514	3,209,949	355,271	

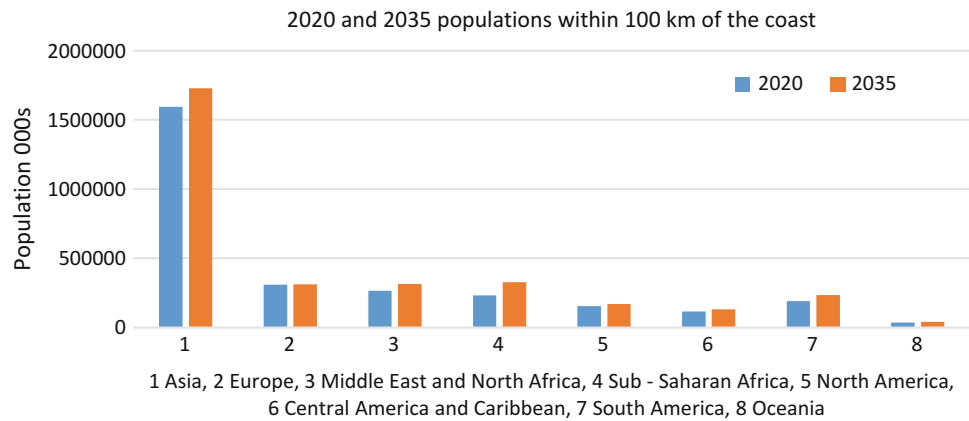
^aFrom Burke et al. (2001)^bFrom Burke et al. (2001)^cFrom United Nations (2017)^dFrom Burke et al. (2001)^eTotal 2020 population x 0.01 x % Population Within 100 km from Coast.^fFrom United Nations (2017)^gTotal 2020 population x 0.01 x % Population Within 100 km from Coast^h2035 population within 100 km from the coast minus 2020 population within 100 km from Coastⁱ((2035 coastal population divided by 2020 coastal population) minus 1)x100)^jGrand Totals, sums for all countries

more rapidly than the global average (42%), while Europe (1%) will be much less.

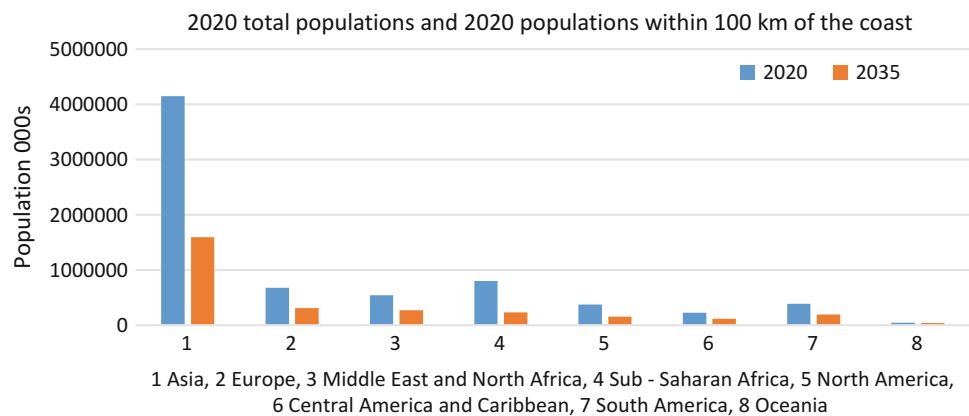
In terms of gross numbers, however, it is Asia, with a projected 8% increase translating to 1.3 billion more persons living in the coastal zone by 2035, that is of greatest potential environmental impact. Sub-Saharan Africa's 42% increase equates to 1 billion more people in their coastal zones. These two regions, totaling 2.3 billion more coastal dwellers between 2020 and 2035, account for 64% – almost two-thirds – of all population growth. Socioeconomically, it might well be argued that these two regions also represent the lowest per capita income of all Earth's people.

Rapid coastal population growth coupled with limited financial and natural resources is a well-documented recipe arguing against sustainable development (see Maul – *Small Islands* – this volume). The weathered mantra “Give a man a fish – feed him for a day; teach a man to fish – feed him forever” will not withstand the scrutiny of the population statistics in Table 1. Humankind requires resources to survive, infrastructure to provide protection from natural hazards, and a society to provide essentials of public health and governance. Coastal resources are finite, as is Earth's ability to provide in extra-coastal regions. Overfishing is perhaps only the first sign that coastal-dwelling humankind is growing beyond sustainable limits.

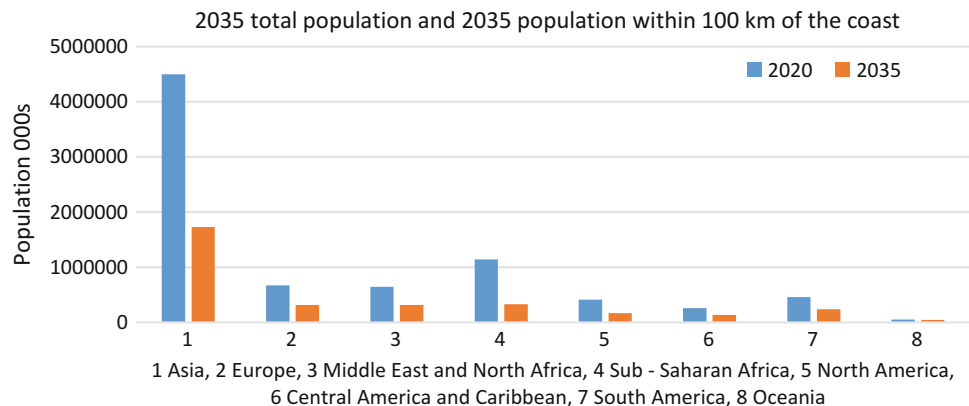
Demography of Coastal Populations, Fig. 1 2020 and 2035 coastal populations within 100 km of coast by region



Demography of Coastal Populations, Fig. 2 2020 total population (countries with a coastal zone) and 2020 population living within 100 km of the coast, by region



Demography of Coastal Populations, Fig. 3 2035 total population (countries with a coastal zone) and 2035 population living within 100 km of the coast, by region



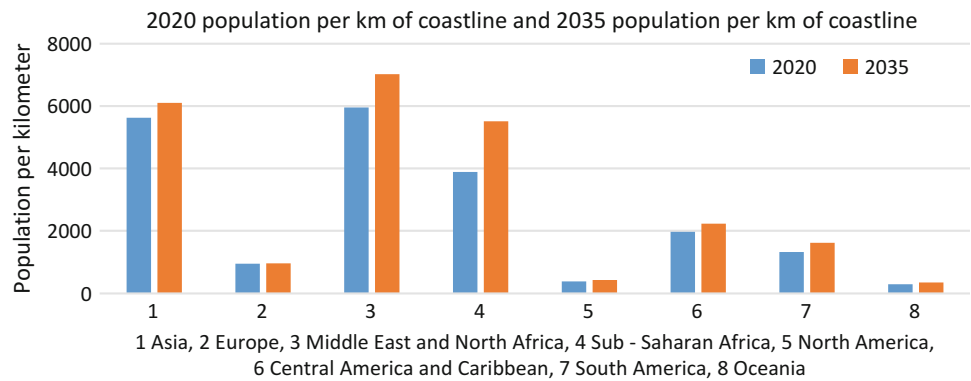
While feeding burgeoning populations is an essential enterprise, “sustainable development” must also include some other aspects of quality of life. In many tropical regions, the great storms variously called hurricanes, typhoons, or cyclones have been the cause of more loss-of-life than any other natural hazard. In the twentieth century, an estimated 1,000,000 souls have been lost to this one hazard than any other. With 64% increase in coastal populations in Asia and Sub-Saharan Africa by 2035, the risk is substantially

increased. In extra-tropical environments, these storms are still extremely destructive, not to mention the intense winter storms of upper mid-latitude flailing against coastal populations. Modern weather forecasting with satellite and radar observations in developed countries clearly is mitigating the risk to coastal communities of such storms.

Another coastal natural hazard, tsunami waves, knows no latitudinal boundaries, as do tropical storms. Although often thought of as a natural hazard of the Pacific Basin, tsunamis

Demography of Coastal

Populations, Fig. 4 Population per 1 km of coastline, 2020 and 2035, living within 100 km of the coast, by region



are in fact a source of great risk to coastal dwellers in all oceans, especially the Mediterranean and the Caribbean Seas. The great Lisbon earthquake and tsunami of 1755 would cause orders-of-magnitude more loss of life and socioeconomic disruption today if repeated in the North Atlantic Ocean.

Sea level rise due to thermal expansion of the upper layer of the ocean and to the melting glaciers will affect many countries, especially small island countries. Coastal flooding and saltwater intrusion are two consequences for countries with sloping shorelines. There is considerable variation in the overall calculated rate of global sea level rise, ranging from 1.7 mm/year to a high of 3.17 mm/year (IPCC 2017). The current public takes little notice, as the rise in sea level is hardly perceptible; however, future coastal populations will surely suffer depending on location, geography, and increased rates of sea level rise.

Beyond the natural hazards, those of geophysical origin, are the biogeographical hazards of intense coastal population growth and development. Humankind produces domestic and industrial wastes. These wastes can be hazardous to others of our species – producing numerous and well-known toxicities or communicable diseases. Given that the greatest growth of our species is in two coastal regions – Asia and Africa – it is reasonable to project increased reports of water-borne public health outbreaks. Implantation of acceptable waste management practices for example, alone, is potentially of a magnitude to overcome all financial resources of intergovernmental banking agencies in the next quarter-century.

Finally, we briefly compare future coastal population with future tourist population. According to Goldberg (1994) world tourism in 1970 grew from 160 million to 341 million people in 1985, or an annual growth rate of 5%. If one projects this growth rate exponentially from 1985 through 2025, total world tourism will grow to 2.5 billion, or approximately four times the increase in world resident coastal population calculated (Table 1). While it is problematic whether world tourism can sustain such a large growth and surely not all tourists seek only the coastal zone, the calculation does demonstrate that some coastal countries may need to either limit tourism or develop new or more efficient infrastructure to accommodate

the growing numbers of people living and/or recreating on beaches or in water. For example, the economies for many of the countries in Central America and Caribbean Region, and Oceania are driven by tourism. Thus such countries will need to carefully consider their options in the areas of waste disposal, habitat, and pollution in order to optimize conditions leading to clean and natural environments to sustain both the local and national economy.

Conclusions

It is not our intention to paint a bleak picture of our coastal future – but of a realistic scenario of population growth and environmental and socioeconomic consequences. Earth is remarkably resilient to humankind's activities. Our planet's ability to dilute pollution is not infinite, and our coastal impacts will exacerbate the global problem. Surprisingly, the global population growth is not significantly different from what is reasonable to expect in the coastal zone – that is within 100 km of the shore. The environmental and socioeconomic impact however may be more. The essence of this increased potential impact is simply gravity.

The rivers and streams that bring mid-continent discharges to the coastal waters are essentially pressure gradient forces driven by gravity. Communities that discharge their wastes into coastal waters, whether by gravity or pumping, inject their wastes into littoral ecosystems. The 355,271,000 ($\pm$) more of us living within 100 km of the coast between 2020 and 2035 will impact the very small fraction of Earth's surface area we consider home to 40% of humankind. Surviving that percentage growth is a taxing but not insurmountable challenge.

No other challenge will tax the intellect of humankind more than population growth. It is comforting perhaps that the challenge in the coastal zone is not more daunting than in other regions. However, oceanographers have long known that coastal ecosystems are particularly at risk from human activity. These fragile natural communities are increasingly seen as endangered, not only because of those living within 100 km of the coast, but also because of all others, whose effluent reaches the coast.

Cross-References

- Beach Use and Behaviors
- Climate Patterns in the Coastal Zone
- Pressure Gradient Force
- Small Islands
- Tourism and Coastal Development
- Tourism, Criteria for Coastal Sites

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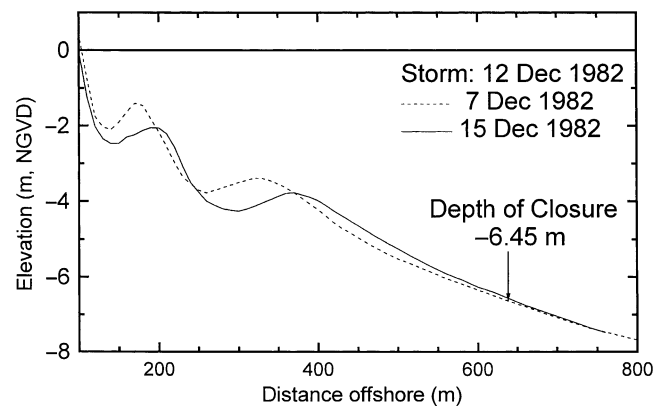
Depth of Closure on Sandy Coasts

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Kraus et al. (1999, p. 272) proposed the following definition of the *depth of closure* (DoC):

The depth of closure for a given or characteristic time interval is the most landward depth seaward of which there is no significant change in bottom elevation and no significant net sediment exchange between the nearshore and the offshore.

Figure 1 illustrates the concept by showing the change in the nearshore profile resulting from a single storm. In this case, the DoC is 6.45 m as there is little change in the bottom deeper than this. Thus, the DoC separates the active nearshore from a less-active offshore and therefore is an important parameter in many coastal engineering projects. For example, in order to build out the entire beach profile, nourishment quantities are often computed by multiplying the desired added beach width by the DoC. In another application, numerical models of beach change use the DoC as an offshore limit to their computations. Because of its importance, the



Depth of Closure on Sandy Coasts, Fig. 1 Example nearshore profile change caused by a single storm event with the DoC located at the deepest point of significant bottom change (in this case, <6 cm). (Data from the Corps of Engineers' Field Research Facility at Duck, NC, USA)

DoC is a topic of considerable research on how to predict and measure it, and on what it represents with respect to sediment transport.

Analytical Prediction

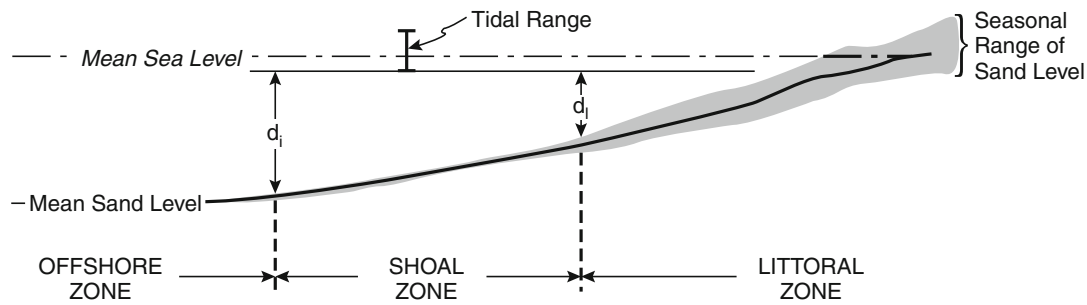
Hallermeier (1977, 1978, 1981a, b, c) developed the first predictive formula for the DoC. Using laboratory tests and limited field data, he hypothesized two limit depths, d_l and d_i (Fig. 2). He defined the inshore limit depth, d_l , as the seaward limit of the *littoral zone*, where “intense bed activity (is) caused by extreme near-breaking waves and breaker-related currents” (Hallermeier 1981c, p. 258). Seaward of d_l , the deeper limit, only insignificant onshore–offshore transport by waves occurs. Between these two limits, is the shoal zone, a buffer region zone where expected waves have neither a strong nor a negligible effect on the sandy bed during a typical annual cycle of wave action.

Hallermeier (1978) suggested an analytical approximation, using linear wave theory for shoaling waves, to predict an *annual* value of d_l :

$$d_l = 2.28H_e - 68.5 \left(\frac{H_e^2}{gT_e^2} \right) \quad (1)$$

where d_l is the annual DoC below mean low water (MLW), H_e is the nonbreaking significant wave height that is exceeded 12 h per year (0.137% of the time), T_e is the associated wave period, and g the acceleration due to gravity.

According to Eq. 1, d_l is primarily dependent on wave height with an adjustment for wave steepness. Hallermeier (1978, p. 1501) proposed using the 12 h exceeded wave height because it allowed sufficient duration for “moderate adjustment towards profile equilibrium.” Equation 1 is based



Depth of Closure on Sandy Coasts, Fig. 2 Definition sketch showing limits d_i and d_b , where d_i is the maximum water depth for nearshore erosion by extreme (12 h per year) wave condition (From Hallermeier 1981c)

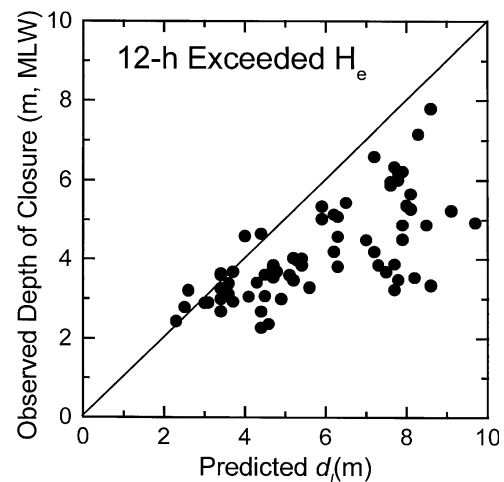
on quartz sand in salt water with a submerged density of $\gamma' = 1.6$ and a median diameter between 0.16 and 0.42 mm, which typifies conditions in the nearshore for many beaches. Because d_i was derived from linear wave theory for shoaling waves, d_i must be seaward of the influence of intense wave-induced nearshore circulation.

However, because of various factors, Hallermeier (1978, p. 1502) “proposed that the calculated d_i be used as a minimum estimate of profile close-out depth with respect to low (er) tide level.” Because tidal or wind-induced currents may increase wave-induced near-bed flow velocities, Hallermeier suggested using MLW as a reference level to obtain a conservative DoC. Although Hallermeier considered only an annual period, Stive et al. (1992) extended Eq. 1 to other time periods by defining H_e as the 12-h exceeded wave height over any time period.

Birkemeier (1985) used 2 years of profile data from Duck, North Carolina, to examine Eq. 1. He reasoned that since any storm could be the annual event, an event-dependent d_i could be computed for each storm. Taking this approach, he showed that the functional relationship proposed by Hallermeier appeared reasonable, but Eq. 1 overpredicted the observed DoC by about 25%. Nicholls et al. (1998) expanded on this analysis using 12 years of profile data to examine the parameters in Eq. 1. Using a depth change criteria of <6 cm between surveys as an indicator of the DoC, they confirmed that using the 12-h exceeded wave height and a datum of MLW provided a conservative estimate of the observed DoC (Fig. 3). They also examined the variation of the DoC with time and found that Eq. 1 became less capable at prediction as the time interval was increased beyond 4 years. They hypothesized that this indicates that, as the time interval increases, the DoC will be governed by other processes beside the highest wave height.

On eastern Lake Michigan, Hands (1983), using 9 years of profile surveys, determined that the closure depth was equal to twice the height of the 5-year return period wave height (H_5):

$$Z \simeq 2H_5 \quad (2)$$



Depth of Closure on Sandy Coasts, Fig. 3 Observed versus predicted DoC using Eq. 1 for erosional events measured in Duck, NC, USA (Modified from Nicholls et al. 1998). Except for a few cases of $d_i < 4$ m, Eq. 1 overpredicts, providing a conservative estimate of the actual observations

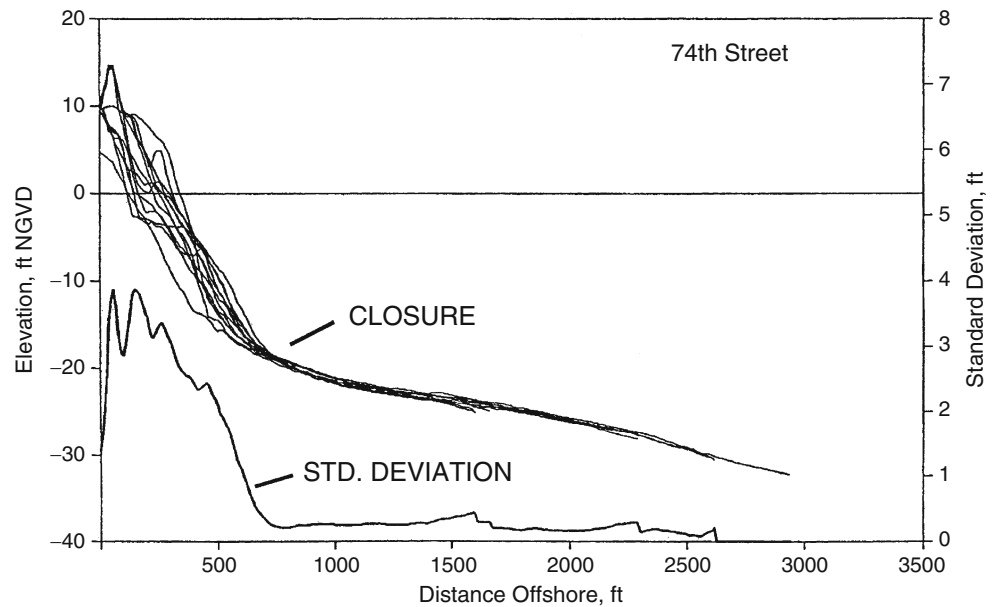
This relationship is similar to Eq. 1, being based on wave height. One requirement of both equations is that they assume sand-rich profiles and require knowledge of the near-breaking wave climate.

Empirical Prediction

The DoC can also be determined empirically by examining seafloor elevation changes measured by repetitive beach and nearshore surveys and identifying the depth at which changes become insignificant (see also Stauble et al. 1993; Grosskopf and Kraus 1994; Nicholls et al. 1998; Kraus et al. 1999). Since surveys integrate the impact of all processes acting on the profile, they inherently include the influence of storm and calm conditions along with important parameters not included in Eq. 1 such as grain size, sediment transport, and underlying geology.

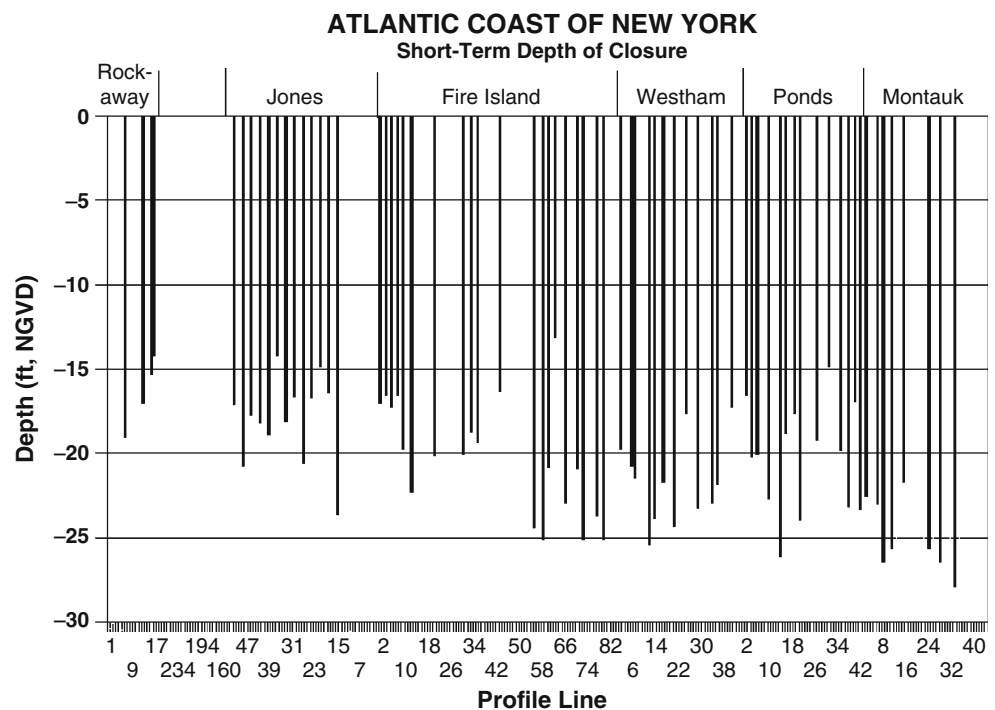
Depth of Closure on Sandy Coasts, Fig. 4

Profile surveys and standard deviation of seafloor elevation at 74th Street, Ocean City, MD (Adapted from Stauble et al. 1993). Surveys conducted from 1988 to 1992. Large changes above the datum were caused by beach fill placement and storm erosion



Depth of Closure on Sandy Coasts, Fig. 5

Variation in short-term closure depth along the south shore of Long Island, NY, computed for four survey dates in 1995 and 1996 (Adapted from Morang et al. 1999). Zones indicate south shore barriers or regions



In order to identify at what depth changes become insignificant, Kraus and Harikai (1983) examined the standard deviation in depth change computed from a series of surveys, and defined closure as the depth where the standard deviation decreased markedly to a nearconstant value. They interpreted the landward region of this depth to be the active profile, where the seafloor was influenced by gravity waves and storm-driven water level changes. The offshore region, of

smaller and nearly constant standard deviation, was primarily influenced by lower frequency sediment-transporting processes such as shelf and oceanic currents (Stauble et al. 1993).

An example of how closure was determined empirically at Ocean City, MD, is shown in Fig. 4. A clear reduction in the standard deviation occurs at a depth of about 5.5–6 m below the National Geodetic Vertical Datum (NGVD) for the 13 surveys conducted over a period of 5 years. For the 5.6 km of

shoreline surveyed at Ocean City, the DoC ranged between 5.5 m and 7.6 m. This longshore variation may reflect complex nearshore bathymetry resulting from shore-attached linear shoals. A similar longshore variation in the DoC is shown in Fig. 5 for 110 south shore Long Island profile lines. Generally, the depth increases toward the east, with Rockaway Beach averaging 5 m below NGVD and the Montauk zone averaging 7.6 m. These values were based on four high-quality surveys collected between 1995 and 1996 (Morang et al. 1999). The depth increase toward the east corresponds to an eastward increase in wave energy and further reflects the wave height dependency of Eq. 1. These data also demonstrate that there can be considerable variation in DoC along an 80 km, relatively straight stretch of shore.

While survey data can be used to determine the DoC, surveys have several limitations that require careful consideration. Most importantly, a repetitive survey program should cover at least the time interval of interest with enough frequency to resolve the natural variation which occurs. In general, the longer the program, the more different conditions will be monitored and the more robust the closure depth determination will be. To resolve small changes in depth and in standard deviation, highly accurate surveys ($\pm 2\text{--}3$ cm) are required and profiles must extend offshore far enough to include the entire active shoreface. In the United States east coast, surveys to 10 m water depth are normally sufficient.

Application

Closure depth is used in a number of coastal engineering applications such as the placement of mounds of dredged material, beach fill, placement of ocean outfalls, and the calculation of sediment budgets. The time dependent nature of the active portion of the shoreface requires a coastal engineer to consider return period. The closure depth that accommodates the 100-year storm will be deeper than one for a 10-year storm. Therefore, a closure depth must be chosen in light of a project's engineering requirements and design life. For example, if a berm is to be built in deep water, where it will be immune from wave resuspension, what is the minimum depth at which it should be placed? This is an important question because of the high costs of transporting and disposing material at sea. It would be tempting to use a safe criterion such as the 100- or 500-year storm, but excessive costs may force the project engineer to consider a shallower site that may be stable only for shorter return period events. The prediction techniques described above allow for both the determination of a closure depth and for the consequences of picking a shallower or deeper depth to be quantified.

Most research into the closure depth have been carried out on sandy beaches with a thick sand layer in wave-dominated

environments. Therefore, assumptions behind the concept hold best on open coasts far from structures and inlets and may be less valid for beaches where the morphology is largely controlled by underlying geology, for example, where the active sand is only a veneer above rock or clay. Also, little is known about the DoC for coasts where most sediment is moved by unidirectional currents, such as rivers and tidal channels or where there is long-term net loss of sediment.

Cross-References

- [Beach Profile](#)
- [Cross-Shore Sediment Transport](#)
- [Cross-Shore Variation of Grain Size on Beaches](#)
- [Dynamic Equilibrium of Beaches](#)
- [Littoral](#)
- [Net Transport](#)
- [Sandy Coasts](#)

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Depth of Disturbance

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Definition

Depth of disturbance, is the depth to which wave action can be expected to disturb the underlying bottom sediment. Sources of disturbance other than wind waves are not discussed. The sediment is considered to be unconsolidated particles in the sand size range.

Background

The approach adopted is to develop analytic expressions to estimate the depth of disturbance through the use of the velocity ratio,

$$U = u_{-h \max} / u_{\text{crit}}, \quad (1)$$

Where $u_{-h \max}$ is the maximum near-bottom orbital velocity of the wave and u_{crit} is the critical velocity required to initiate sediment movement under the wave. If U is greater than 1.0, then movement of the sediment would be expected; if U is less than 1.0, sediment movement would not be expected. Ahrens and Hands (1998) found the parameter U very effective for distinguishing between beach erosion/accretion events. For shallow water conditions, $u_{-h \max}$ will be estimated using Stream Function Wave Theory (SFWT), (Dean 1974). For deep water, $u_{-h \max}$ will be estimated using linear wave theory. The critical velocity will be estimated using the relation,

$$u_{\text{crit}} = \sqrt[3]{(8\Delta g d)}, \quad \text{for } 0.0625 \leq d \leq 2.0 \text{ mm}, \quad (2)$$

where g is the acceleration of gravity, d is the sediment grain diameter, and $\Delta = (\rho_s - \rho)/\rho$, where ρ_s is the density of the sediment and ρ is the density of water; Eq. 2 is based on research by Hallermeier (1980).

Concepts and Development

It is convenient to approximate the dimensionless near-bottom velocity as a function of relative depth, that is, $u_{-h \max} T/H = f(h/L_0) < \text{check}>$, where H is the wave height, h is the water depth, and L_0 is the deep water wave length given by $L_0 = gT^2/2\pi$, where T is the wave period. The general functional form used is:

$$f_i(h/L_0) = \exp[C_0 + C_1(h/L_0)](h/L_0)^{C_2}, \quad (3)$$

where C_0 , C_1 , and C_2 are dimensionless coefficients developed from regression analysis. Table 1 gives the values of the coefficients for the two wave theories and ranges of relative depth.

The breaker height, H_b , consistent with SFWT, can be estimated quite accurately as a function of relative depth using the relation,

$$H_b = 0.171 L_0 \tan h[0.73(2\pi h/L_0)], \quad (4)$$

(Ahrens and Hands 1998). In general, the wave height can scaled as a fraction of H_b , that is, $H/H_b/a$, where 1.0 or 4.0 in this article. When the above relations are brought together, the following equation is obtained,

$$U = 0.171 \tan h[0.73(2\pi h/L_0)] f_i(h/L_0) / a [16\pi\Delta(d/L_0)]^{1/2}, \quad (5)$$

where $f(h/L_0)$ can be any of the three functions indicated by Eq. 3 and Table 1. Equation 5 shows that U is a function of three variables, h/L_0 , d/L_0 , and Δ , however, typically Δ only varies over a small range. To obtain a visual impression of the influence h/L_0 , d/L_0 , on the depth of disturbance, U is set equal to 1.0 and $\Delta = 1.6$ to produce Fig. 1.

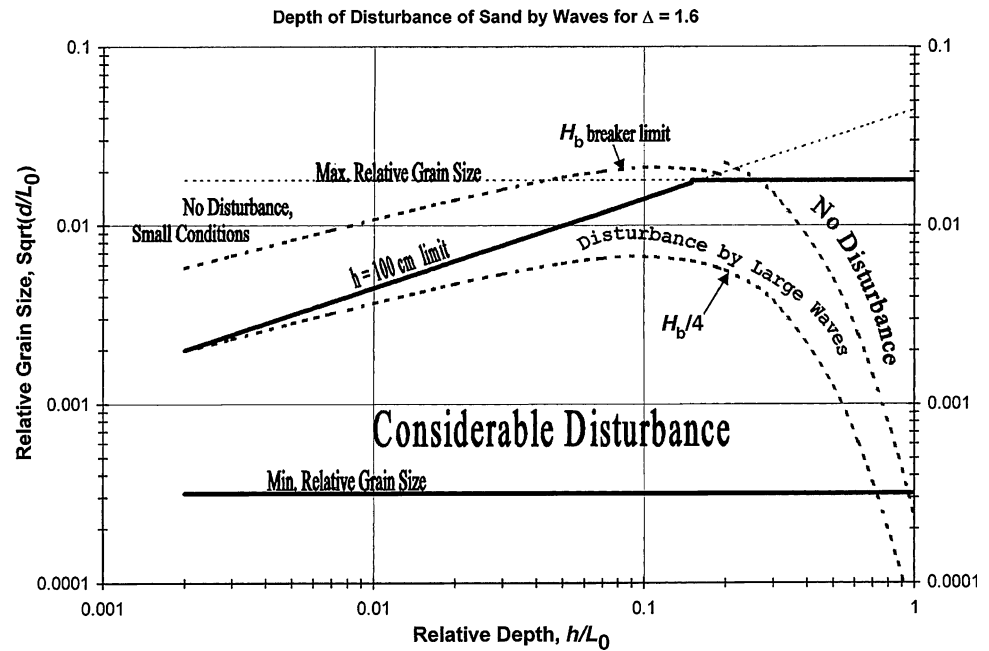
Two roughly parallel curves arch across Fig. 1 the upper curve is associated with H_b and the lower curve with $H_b/4$. There is a slight discontinuity in the H_b curve because linear wave theory and SFWT are not quite consistent at the transition depth, h/L_0 0.2. In Fig. 1 the area of primary interest is

Depth of Disturbance, Table 1 Coefficients used to estimate dimensionless bottom velocities

Function	Wave theory	Wave height	Relative depth range	Coefficients		
				C_0	C	C_2
f	SFWT	H_b	0.002 h/L_0 0.2	−0.227	−2.078	−0.600
f	SFWT	$H_b/4$	0.002 h/L_0 0.2	0.0835	−2.819	−0.601
f	Linear	All	0.2 h/L_0 2.0	1.930	−6.368	0.103

Depth of Disturbance,

Fig. 1 Depth of disturbance plane, relative grain size, $\sqrt[3]{(d/L_0)}$, versus relative depth, h/L_0



outlined by three heavy, straight lines. The maximum and minimum relative grain sizes are based on the grain size range of $0.0625 \leq d \leq 2.0$ mm and the range of wave periods considered, $2.0 \leq T \leq 20.0$ s. The “ h 100 cm limit” line indicates where the upper limit is 100 cm deep; above this line depths are less than 100 cm and the scale of the wave action is small.

Summary

Figure 1 is a map of the depth of disturbance plane for two wave heights; one equal to the breaker limit, H_b , and the other one quarter the breaker limit. On the right side of the figure, in deep water, is a region of no disturbance. Next to the no disturbance region is a region which is disturbed by large waves, that is, $H_b/4 \leq H \leq H_b$. Beneath this region is a region considerably disturbed by large waves. Above the h 100 cm limit is a region disturbed by relatively large waves and surprisingly a region of no disturbance. However, above the $h = 100$ cm limit wave conditions are small since the water depth is less than 100 cm.

Cross-References

- Beach Erosion
- Beach Processes
- Beach Sediment Characteristics
- Depth of Closure on Sandy Coasts
- Waves

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Desalination

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Water is a valuable natural resource, necessary for domestic, industrial, and agricultural activities. Because of the worldwide population growth, economic development, and erroneous management choices, humanity is nowadays facing an increasing scarcity of satisfactory quality water. This threat is stronger in the arid and semiarid areas (Kulga and Adanali 1990) but could extend to other regions in the context of a global climatic change which could modify the precipitation patterns at a worldwide scale (Pearce 1996). The possibility of utilizing saline underground and

seawaters, by removing soluble salts from water by desalination technology, is therefore considered today with an increasing attention.

Major Processes of Desalination

Important research programs had been put in place by the 1980s. Taking into account industrial requirements, three major processes have emerged. Distillation is the older method. Saline water is evaporated partially, leaving the salts behind. The desalinated vapor is then condensed in a separate container. Electrodialysis is based on the mobility of ions in an electrolyte submitted to an electric field. Membranes are placed between a pair of electrodes. Negative ions (anions) migrate toward the anode and positive ions (cations) toward the cathode, creating concentrated and diluted solutions between the alternating membranes. Reverse osmosis is the more recent, and today the most common, technology. The natural osmotic flow is reversed by applying pressure that forces the salt water through a membrane, permeable only for pure water.

A number of other methods may prove valuable under special circumstances or with further development. Freezing is based on the principle that water excludes ions when it crystallizes to ice. In the ion-exchange process, water passes through a bed of specially treated synthetic resin which extracts ions from the solution and replaces them with ions that form water.

Costs of Desalination

Total costs of desalination include initial technological investment and energy costs. The minimum energy requirement represents the energy necessary for obtaining freshwater from saline water in the case of an ideal reversible thermodynamic transformation (Nebbia 1968). In practice, the existing desalination methods consume an amount of energy which is at least 70 times greater than this theoretical minimum (Pearce 1996). This energy consumption is one of the causes for the high costs of desalination. The selection of an appropriate method depends thus on these economical considerations. Distillation is the more expensive process because of the great quantities of fuel required to vaporize salt water. Electrodialysis and reverse osmosis, which can desalt water much more economically, have therefore replaced distillation for the desalination of brackish water. Because of the high cost of membrane replacement however, these two processes are less suitable for desalinating seawater. Technological advancement has also allowed some reduction in the total desalination cost, by improving energy efficiency (multiflash distillation or hybrid systems), by facilitating

transfer processes, by energy recycling (process of cogeneration) or by using solar energy.

Ecological Impacts of Desalination

Wastewater effluents may have potentially adverse effects on the environment. They present excessive salt concentrations, increased temperatures, and increased turbidity levels and can be charged with chemicals, organics, and metals. In addition, distillation entails an oxygen depletion. These problems can be avoided by treating the effluents before releasing them in the environment but this constraint again increases the cost of desalination.

Conclusion

Desalination offers a nearly unlimited potential of freshwater, particularly in coastal areas. However, this technology is still not extensively developed, because the cost of freshwater production remains a problem. Some countries, which have opted for large-scale desalination, are now searching for more economical solutions (Nativ 1988). Desalination is essentially utilized to solve survival problems and development of arid and semiarid areas. The use of desalinated water in agriculture is by far less important, particularly because irrigation does not need high water quality (Rhoades et al. 1992). The limits of salinity are generally two to five times those tolerated for industrial and domestic water (Nebbia 1968). In the case of salinity sensitive crops such as citrus, however, partial desalination by mixing a part of desalinated water to saline underground water, is an interesting solution (Matz 1965). Because of industrial research and because of the rise in cost for conventionally treated water, desalination turns out to be more and more feasible, economic, and reliable and its importance is expected to increase in the future years. However, even if the problem of freshwater rarefaction can be, in the more or less long-term, technologically solved, we must wonder if an increasing demand for freshwater is compatible with the current will of preserving ecosystems and biodiversity. Excessive water utilization has serious consequences on the function of precarious ecosystems like wetlands (Pearce 1996). Better management options, centered on conservation and economy of resources constitutes a cheaper, more durable, and more ecological solution in face of water scarcity.

Cross-References

- [Demography of Coastal Populations](#)
- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Water Quality](#)

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Desert Coasts

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Very great environmental differences exist among the coastal deserts of the world. On an overall worldwide basis climate is the best criteria for subdivisions. Climatically, the coastal deserts fall into four principal types which can be called hot, warm, cool, and cold (Figs 1 and 2).

Hot Desert Coasts

The hot coastal deserts cover the largest areas (Fig. 3). Their approximate length is 10,000 miles (16,000 km). This truly tropical hot climate exists only along the southern half of the Red Sea and adjacent coasts of Arabia and the Somaliland horn of Africa (Figs. 4, 5, 6, and 7). The equatorward sections are warm even in winter with winter temperatures getting above 80°F (25°C) in the heat of the day and only dropping to 70°F (21°C) in the coolest part of the night. The poleward or subtropical type prevails along the Persian Gulf and the Indian Ocean margins of northwestern India, Pakistan, and Iran the northern Red Sea; the Gulf of California, and northwestern Australia.

Most of the hot coastal deserts have relatively little cloud or fog but very high solar radiation. Owing to shallow sill restrictions at the mouths of the Red Sea and the Persian Gulf the water does not mix freely with the colder water of the open

ocean, and as a result the water temperature is higher by several degrees. In summer the surface temperature in the southern part of the Red Sea is above 86°F (30°C) and in winter over 77°F (25°C). The Persian Gulf is hotter than the Red Sea in summer, but much cooler than the Red Sea in winter.

Warm Desert Coasts

The second type of coastal desert can be called warm rather than hot because summers are less hot, with mean temperatures, even in the warmest month being less than 86°F (30°C). The longest single stretch of this climate, about 1,600 miles (2,600 km), is along the Mediterranean Sea, where the Sahara Desert reaches the coast (Figs. 8 and 9). Though occasional short rainstorms occur during the winter season, most of the days are bright, mild, and sunny. It might be noted that the climate plus the nearness to densely populated Europe, makes this one of the important winter seaside resort areas of all coastal deserts.

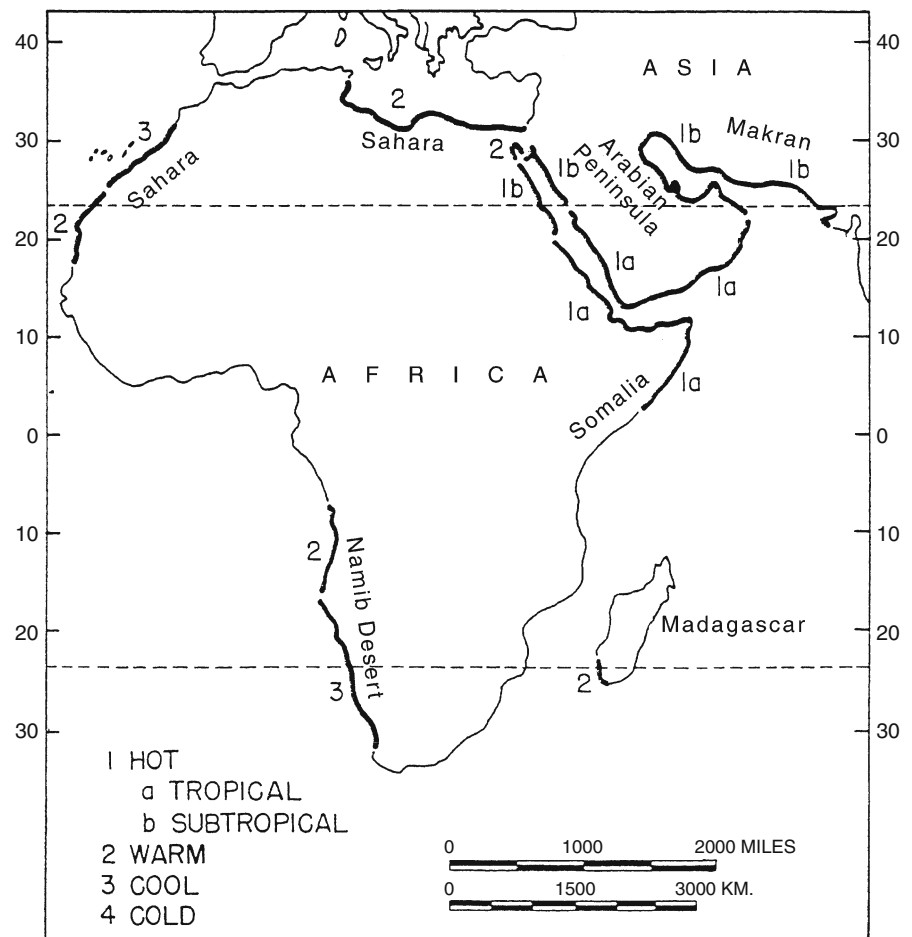
Warm coastal desert climates along exist along the west and south end of Baja California, Mexico (Fig. 12), with winter rainfall in the north, summer showers in the south, and non-seasonal very slight rainfall in the center. All the other west-coast deserts of the world also have warm climates in their equatorward positions, such as Angola, Maruritania, Peru, and Australia (Figs. 10, 11, 13, and 16). Like the southern end of Baja California (Fig. 12), these coastal areas all have their rainy season, such as that occurring, in summer, which is in harmony with their marginal tropical locations. A small but distinct area of this type of warm desert occurs at the southwestern coast of Madagascar (Fig. 11). Tourism is developing in some of these areas, especially the southern end of California where visitors from the United States and mainland Mexico come by air, sea, or some by road. Near their equatorial ends, these west-coast warm deserts are relatively free from cloud or fog. Toward their polar ends, especially in Peru, southwestern Africa, and Baja California (Figs. 11, 12, and 13), there is frequent low cloud or fog in summer, as the warm deserts merge into the cool deserts.

A particular tourist and biological resource of this part of Baja California is the gray whale, which travels thousands of miles, from the North Pacific to breed in Scammon's Lagoon, at 27°45'N latitude.

Cool Desert Coasts

Type 3, the cool desert coasts from the poleward portion of all five westdesert coasts. They occur nowhere else if the shores of the Great Bight of Australia (Fig. 17) is considered an extension of the west-coast desert. The longest stretches of

Desert Coasts, Fig. 1 Coastal desert types in Asia and Africa. (Adapted from Meigs 1966)



this cool desert coast are along the Peru-Chile coast, 2,100 miles (3,400 km) (Fig. 14).

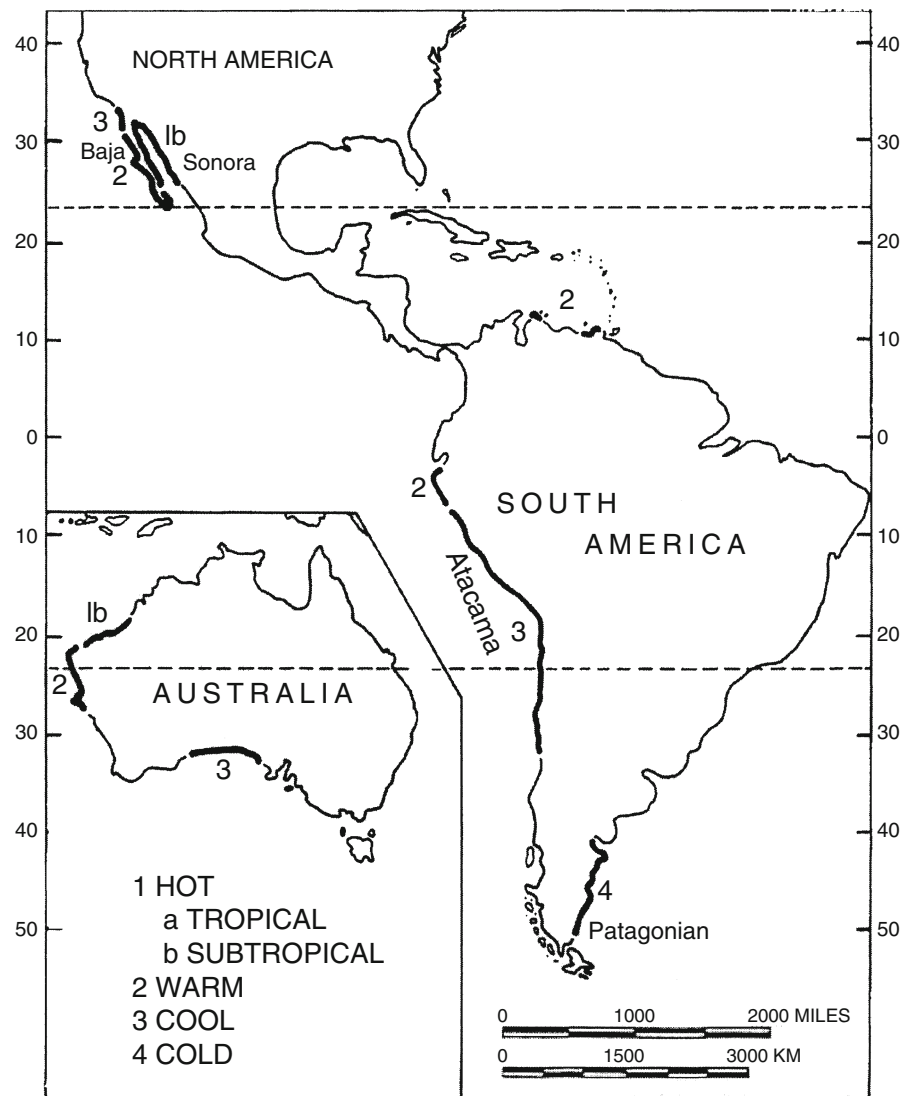
Nowhere else on earth is there a coastal desert of such intense aridity which spans so many degrees of latitude as the Chilean-Peruvian desert. The extreme dryness is all the more interesting when one considers that no part of this desert is far from a tropical ocean, for at a maximum it is never more than a few score miles (kilometers) in width, and the fact that the atmospheric humidity is high is attested to by the presence of a great deal of low stratus cloud, fog, drizzle, and heavy dew. Still it seldom rains. Obviously, the dynamic controls making for drought are amazingly well-developed.

Among the climatic anomalies of the Chilean-Peruvian desert two are particularly notable, (1) the unusual intensity of the dryness in conjunction with its latitudinal extensiveness and the abruptness with which it ends on both the north and south, so that a transitional steppe climate is only meagerly developed, and (2) the brief periods of heavy rainfall and high temperatures which vary occasionally affect mainly the northern parts of this desert area, this temporarily revolutionizing its weather and changing it to wet tropical conditions. In

addition, the Peruvian-Chilean desert has uncommonly well developed those special features of climate which are peculiar to dry western littorals bordered by cool currents. Such as relatively low annual temperatures, small annual and daily ranges of temperature, and the existence of much low stratus fog, cloud, and even drizzle.

Other locations where Type 3, the cool desert coasts, occur are the west coast of Namibia and South Africa, 1,200 miles in length. (2,000 km) (Fig. 11), while smaller sections include northern western Sahara and southern Morocco (Fig. 10). Baja California has a short segment just south of the California border (Fig. 12). These are foggy deserts with fog forming over the cool waters offshore and moving onto the adjacent land either as "high fog or low cloud," or at times directly along the ground. To the cooling effect of the cold water onshore can be added the reduction of solar radiation by the fog blanket. It can also be added that there are regional differences in the season of maximum fog. Along the Chile-Peru coast the season of great fog is in winter; while along the Baja California and northwest Africa coasts there is a summer maximum of the fog. Solar radiations is also affected by high

Desert Coasts, Fig. 2 Coastal desert climatic types. (Adapted from Meigs 1966)



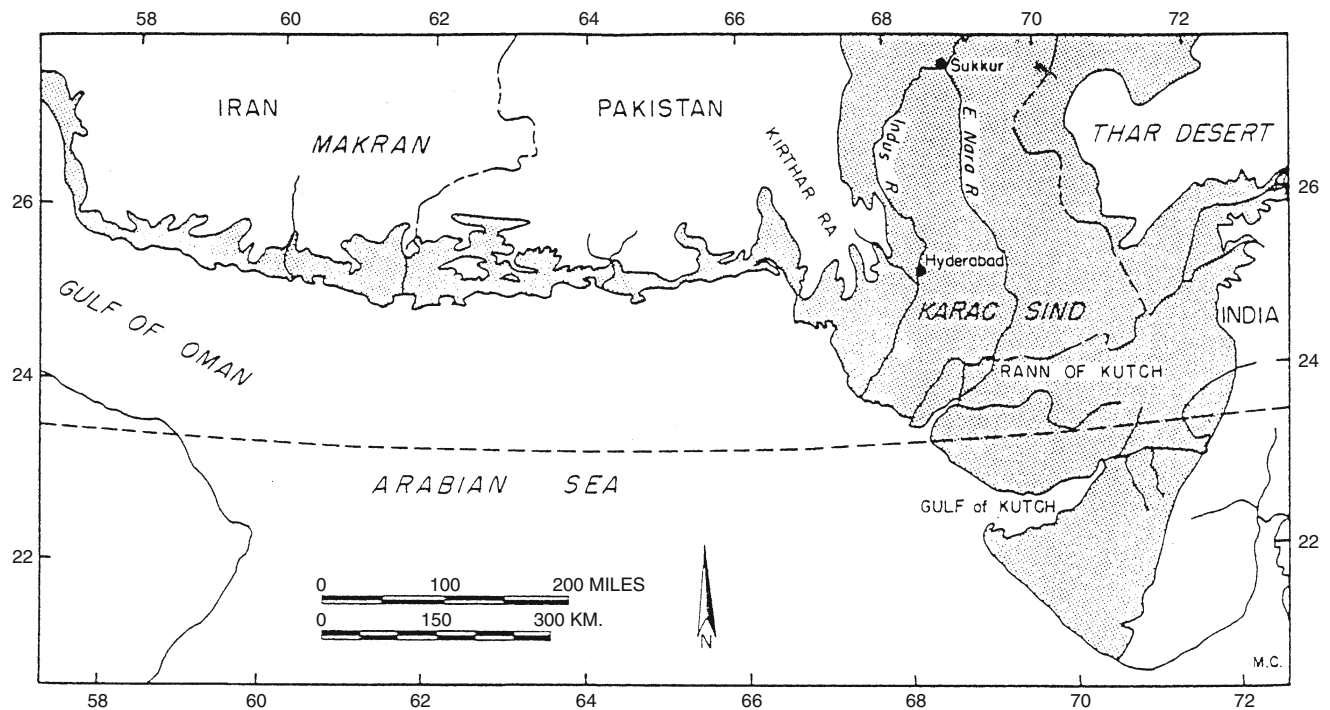
clouds associated with the seasonal rainfall regimes of the neighboring human-occupied areas: winter in most of the cool desert climates, while summer occurs in the main part of the Peruvian coast. However, it can be said, the deserts of the coast provide attractive relief for people from the hot interior valleys or deserts. It might be added that the special circumstances of cool to cold ocean waters offshore has resulted in great fishing grounds.

Cold Desert Coasts

The coldest coastal desert of all, Type 4 of the classification, occurs along the southeast coast of Patagonian Argentina (Fig. 15). Here a cool to cold ocean current reinforces the

dry rain-shadow effect of the Andes on the westerly storms that normally prevail in these latitudes. This desert extends to more than 50°S latitude, which is farther poleward than any other coastal desert in the world. The mean temperature of the coldest month is below 50°F (10°C), and the temperature of the warmest month is between 50° and 72°F (10° and 22°C). With such limited heat only the hardiest vegetation and crops grow in this region. Other resources must provide the principle economic basis for settlement.

This account considers only the connecting, not even the oceans. The inland lakes and seas are not considered, not even the large Caspian Sea. This sea has a climate quite different from those of other coastal deserts. Most of its climate (except along the southwest borders) is continental: colder in winter than the coldest of the coastal deserts (Type 4), with averages



Desert Coasts, Fig. 3 Kathiawar-Makran coastal deserts. (Adapted from Meigs 1966)

below freezing; and ranging with the warmest deserts (Type 2) (Figs 16 and 17) in summer. The sea affects the temperature of the land immediately adjacent to it, as occurs with other coastal deserts.

Coastal Desert Environments

A great variety of landscapes and landforms occur along desert coasts, in fact nearly every type of landform. Desert and semidesert coastal lands make up about 20–25% of all such world regions. Coastal dunes and desert plains predominate but long stretches have steep escarpments or plateaus which rise directly from the sea. In places such as the Makran coast of Pakistan, quite recent uplifted terraces predominate.

The source of sand for desert coasts comes mainly from the sand left on beaches which is moved inland by changing tides and waves plus coastal winds. Another major way for sand to reach desert coasts is by intermittent rivers and streams which carry the sand in short flash floods. The sand is either carried into the sea and then moved by waves and currents along the beaches or picked up by winds in river beds and moved to coastal dunes. There is a changing balance between wind transport of sand and by running water in river beds.

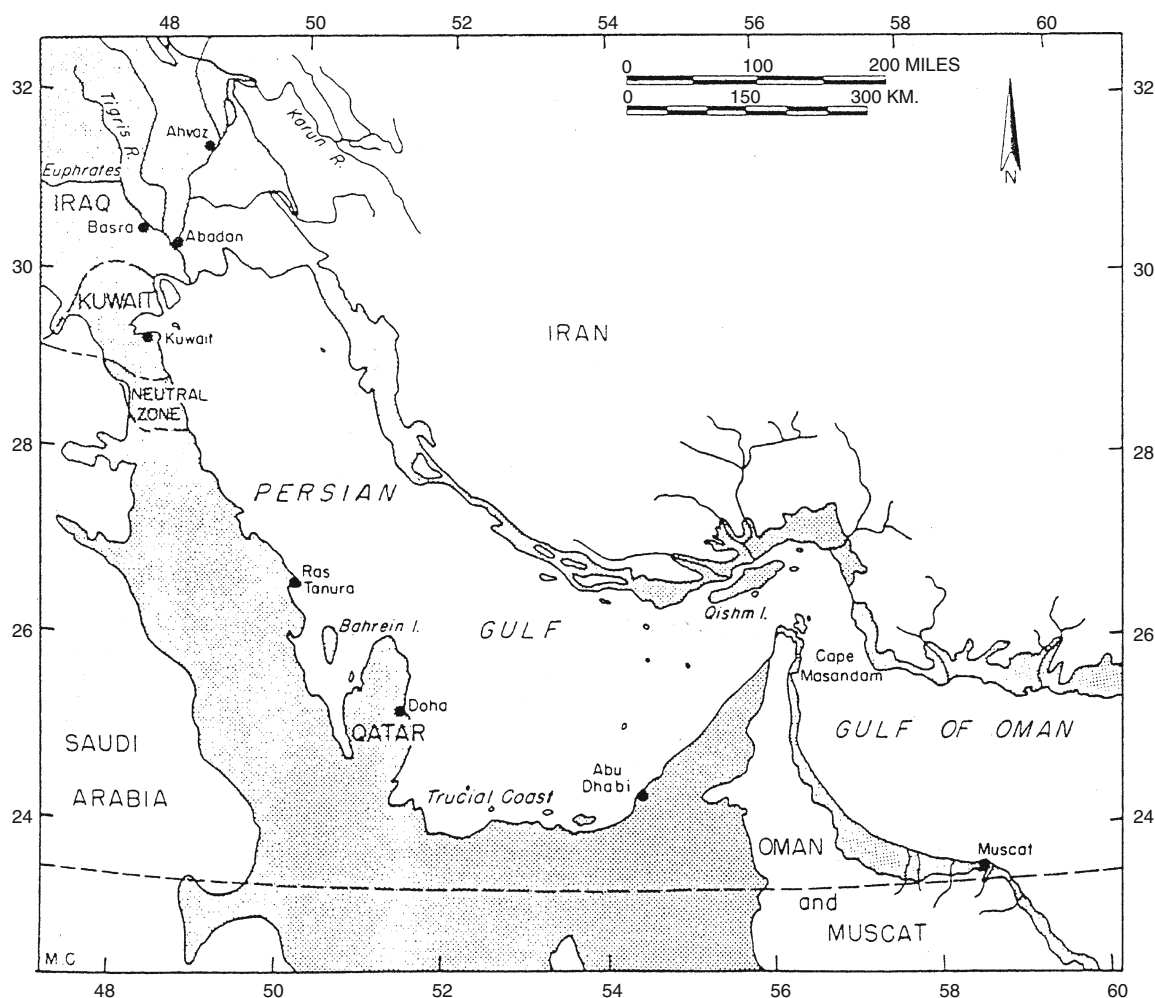
Most aeolian bodies occur in basins in the desert. The Pleistocene probably saw periods of more intense fluvial

erosion than the present. Many coastal ergs are fashioned out of deposits of Pleistocene or late Tertiary alluvium.

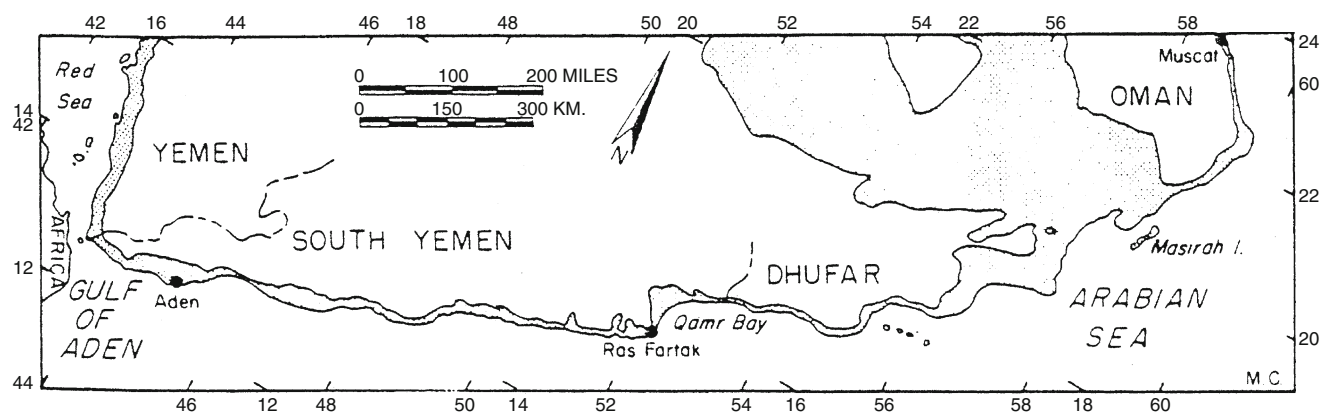
Clay dunes are perhaps surprisingly better authenticated than silt dunes and are more obviously dune like. They have been reported from the Gulf of Texas, northern Algeria, Australia, and the Senegal Delta.

Although it is not difficult to recognize a desert, opinions vary as to the physical features of the environment that should be used as criteria for defining a desert climate. Rainfall must obviously be taken into account, and desert conditions usually prevail in tropical regions when the average rainfall is under 5 in. (127 mm), with semidesert conditions occurring when the average rainfall is over 5 in. (127 mm) but under 15 in. (381 mm). But such figures can be only approximate, because what matters is not only the amount but the effectiveness of rain; and the effectiveness of rain depends on its seasonal occurrence, its rate of evaporation, and the nature of the soil (Cloudsley-Thompson 1975, p. 8).

The aridity of coastal deserts that results mainly from a lack of rain is often due to the way in which the atmosphere circulates, particularly in its lower layers, and this partly explains the geographical distribution of the world's coastal deserts: near latitudes 30°N and S, the direction of the winds tend to vary, and the air sinks. When this happens, it is warmed by the hot ground, and clouds are dispersed. At the same time, high-pressure belts that



Desert Coasts, Fig 4 Coastal deserts of the Persian Gulf area. (Adapted from Meigs 1966)

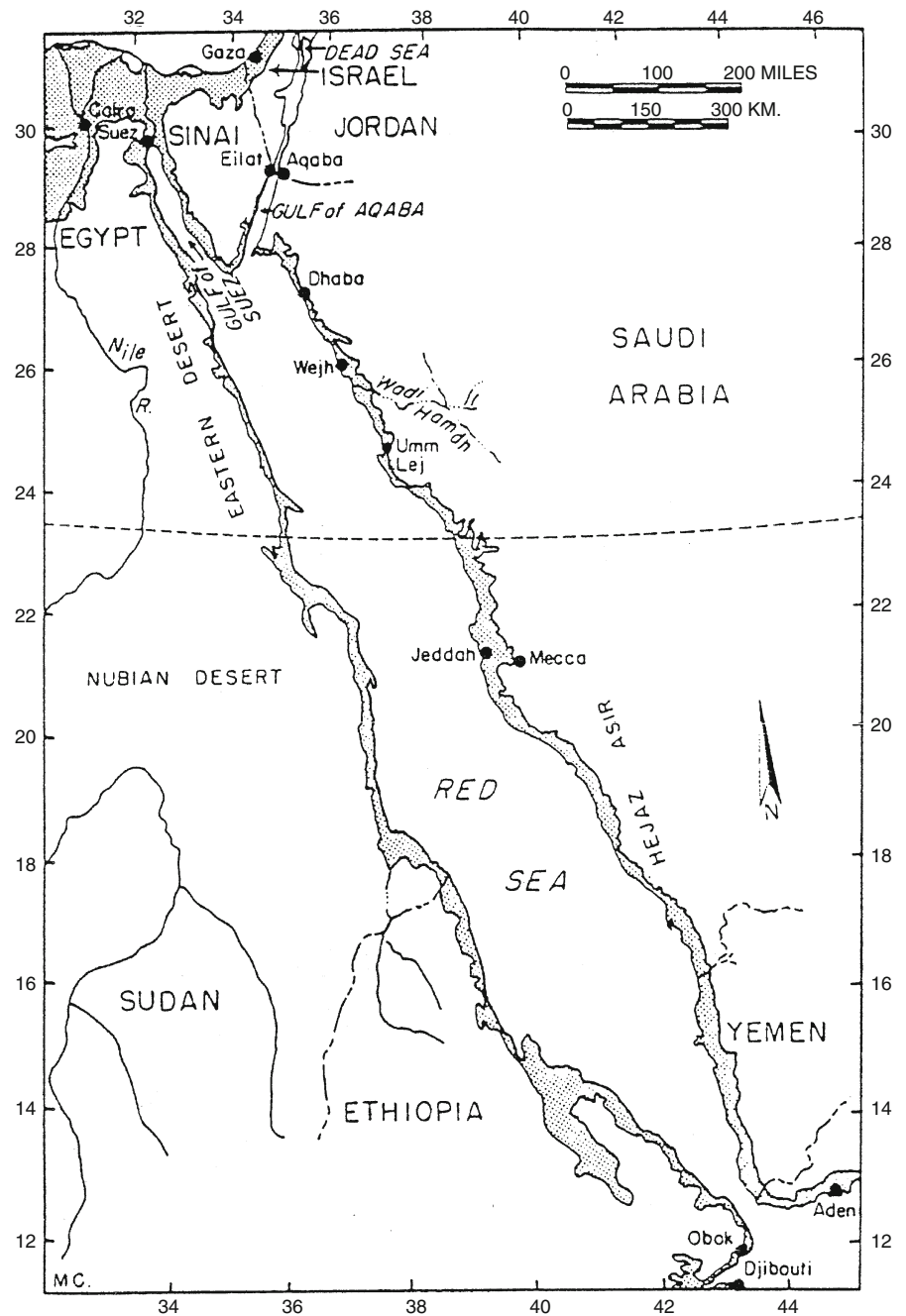


Desert Coasts, Fig. 5 Coastal deserts of southeastern Arabia. (Adapted from Meigs 1966)

separate the polar westerly winds and the tropical easterlies occur near these latitudes. Consequently, the disturbances associated with low pressure, which usually cause rain, are rare. Then, too, there are some local factors that quite often

result in low rainfall. These include the rain-shadow effects of mountain ranges, which cause the air to drop all its moisture as it crosses them. For example, the Patagonian desert including the Argentina coast, is the result of both

Desert Coasts, Fig. 6 Coastal deserts around the Red Sea. (Adapted from Meigs 1966)



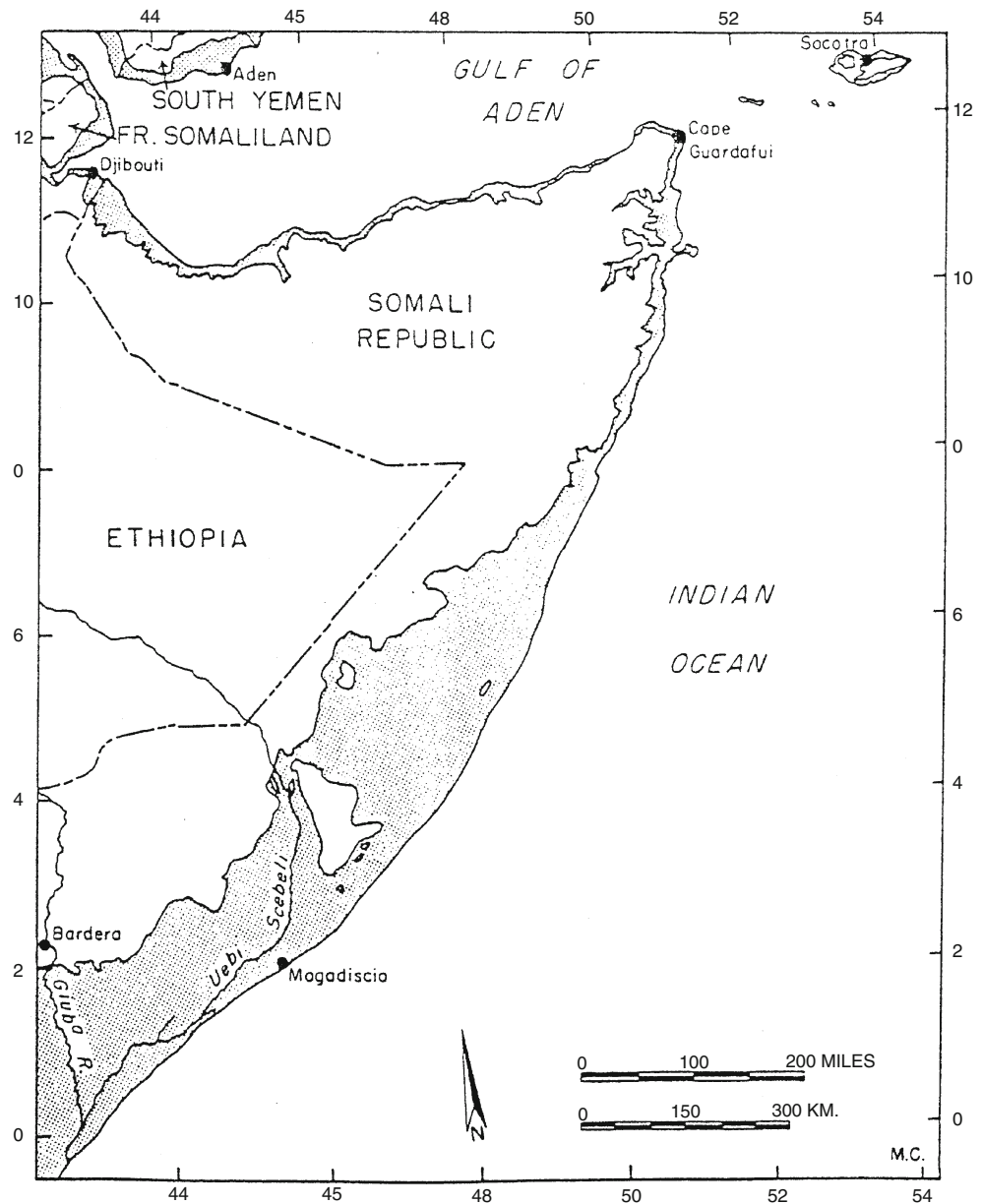
the rain-shadow effect of the Andes Mountains, which block the path of the westerlies, and the cool Falklands current alongshore.

Desert Coastal Vegetation

Most desert coasts support a *xerophytic* type of scrub vegetation which grows sparsely in most areas. A few scattered

trees, mostly the acacia type in tropical coastal deserts, can be found. In years with more than 8 in. (200 mm) of rainfall, a thick layer of grasses, herbs, and shrubs covers the ground. Climate is the major determinant of vegetation types, and rainfall is much more significant than temperature, however the amount of salt comes into play. In general, it can be said that along most desert coasts many of the shrubs are woody and stunted being about 12 in. (300 mm) high with round cushion-like out-lines. With bleached stems and few leaves

Desert Coasts, Fig. 7 Coastal deserts of the Somali Republic. (Adapted from Meigs 1966)



many look like skeletons of plants. A few annuals begin sprouting when warmer days occur. Although the number of families is large in most cases the number of species is small.

Again, speaking generally, the first dunes behind the back-shore differ in vegetation from the inland dunes. The difference is due mainly to high salinity and a high water table which often is only 0.5–4.0 ft (0.15–1.2 m) below the surface.

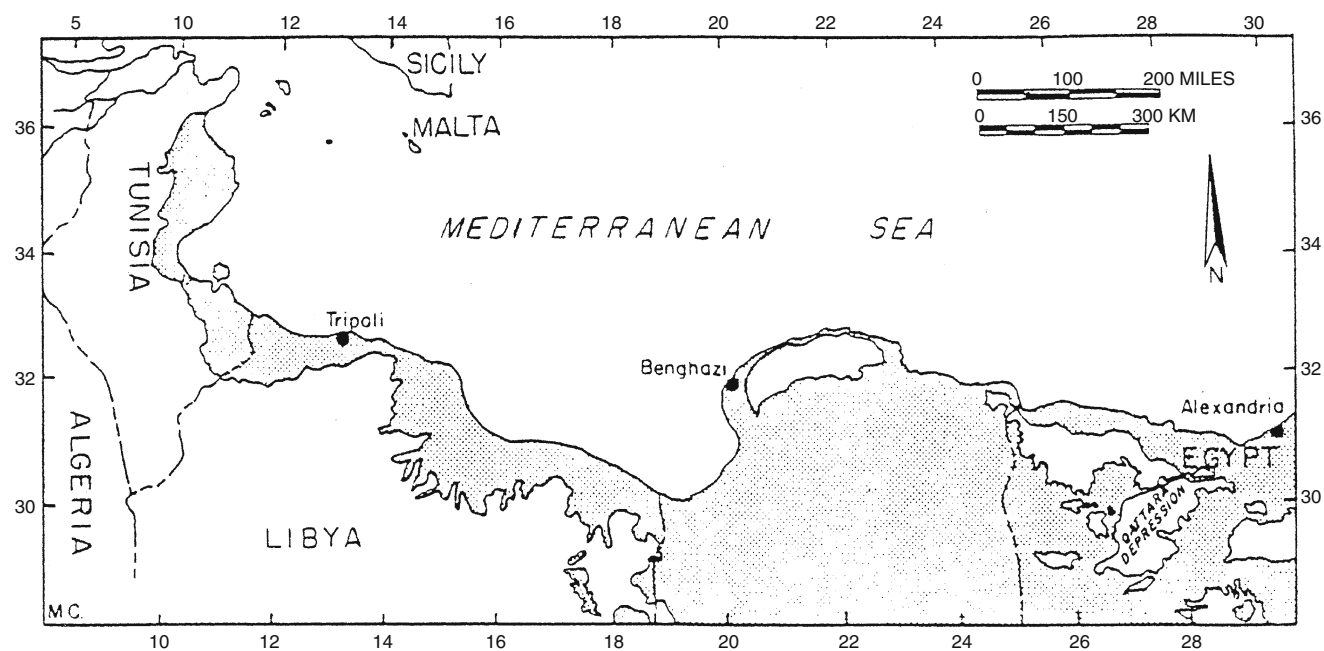
Natural and human factors disturb desert coastal vegetation and lead to temporary changes in plant cover. Disturbed areas result from such natural causes as varying moisture, wind and soil conditions, and destruction by sudden floods. Human activities along the coast which disturb plant successions include the clearing or destruction of the original

vegetation for purposes of cultivation, fuel, fodder, and building material. Animal grazing is the most destructive factor.

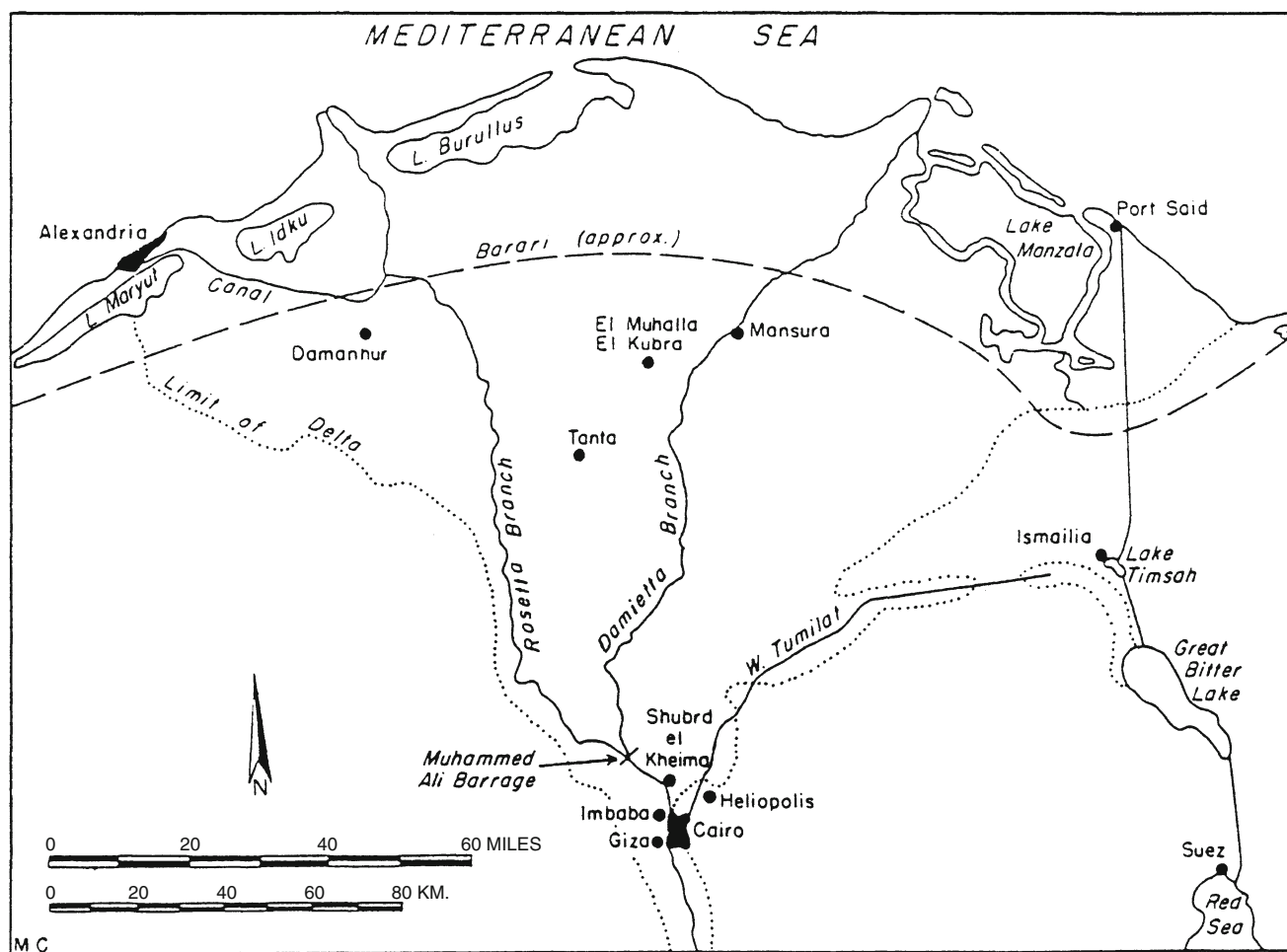
Several desert coasts have unusual vegetation types. For example, the Namib coast has vegetation that lives off of fog; both the Baja desert coast and the Galapagos Islands have unusual vegetation probably because, for a long period of time, they were so isolated, new species of plants developed.

Animals and Plants

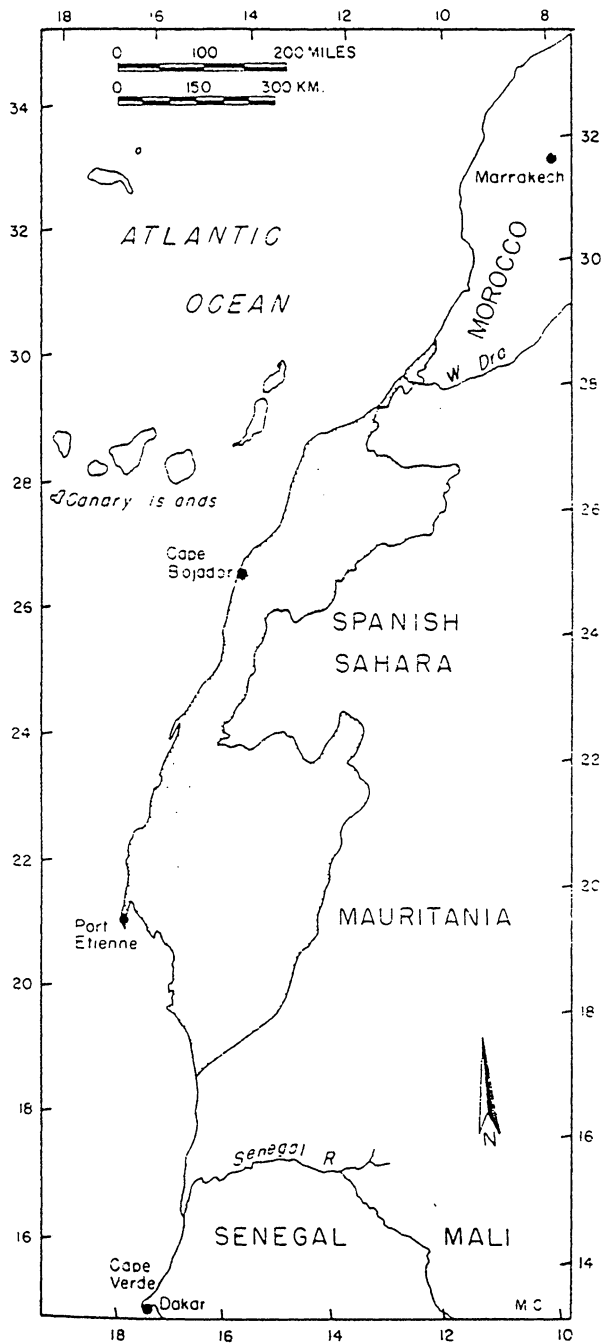
It is not easy for plants and animals to survive in desert coastal regions where extremes of drought and flood, of



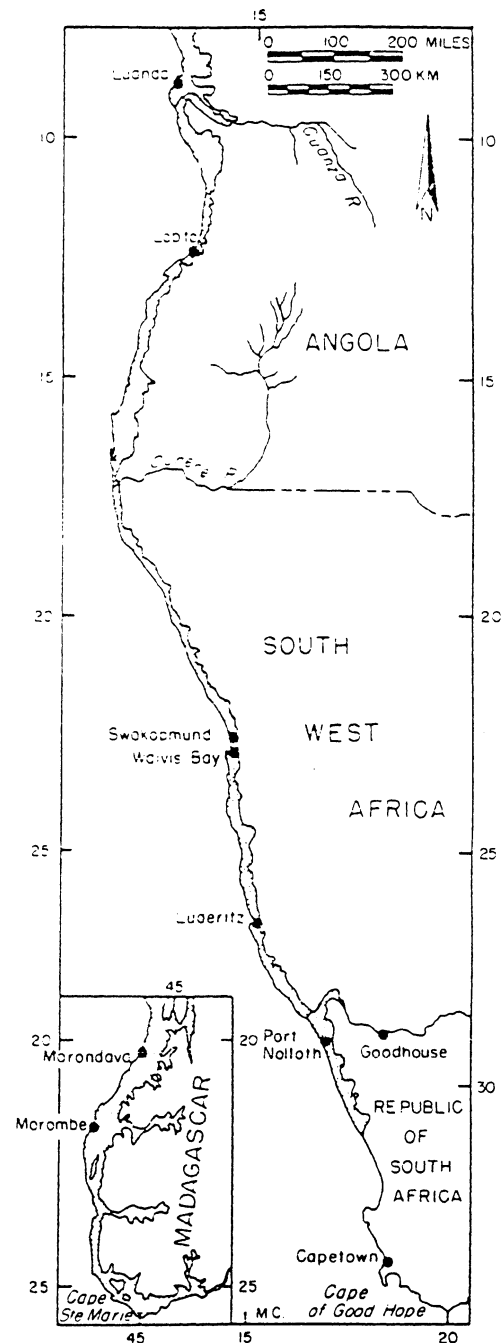
Desert Coasts, Fig. 8 Coastal deserts along the Mediterranean Sea. (Adapted from Meigs 1966)



Desert Coasts, Fig. 9 Coastal deserts of the Nile Delta. (Adapted from Meigs 1966)



Desert Coasts, Fig. 10 Atlantic-Saharan coastal desert. (Adapted from Meigs 1966)

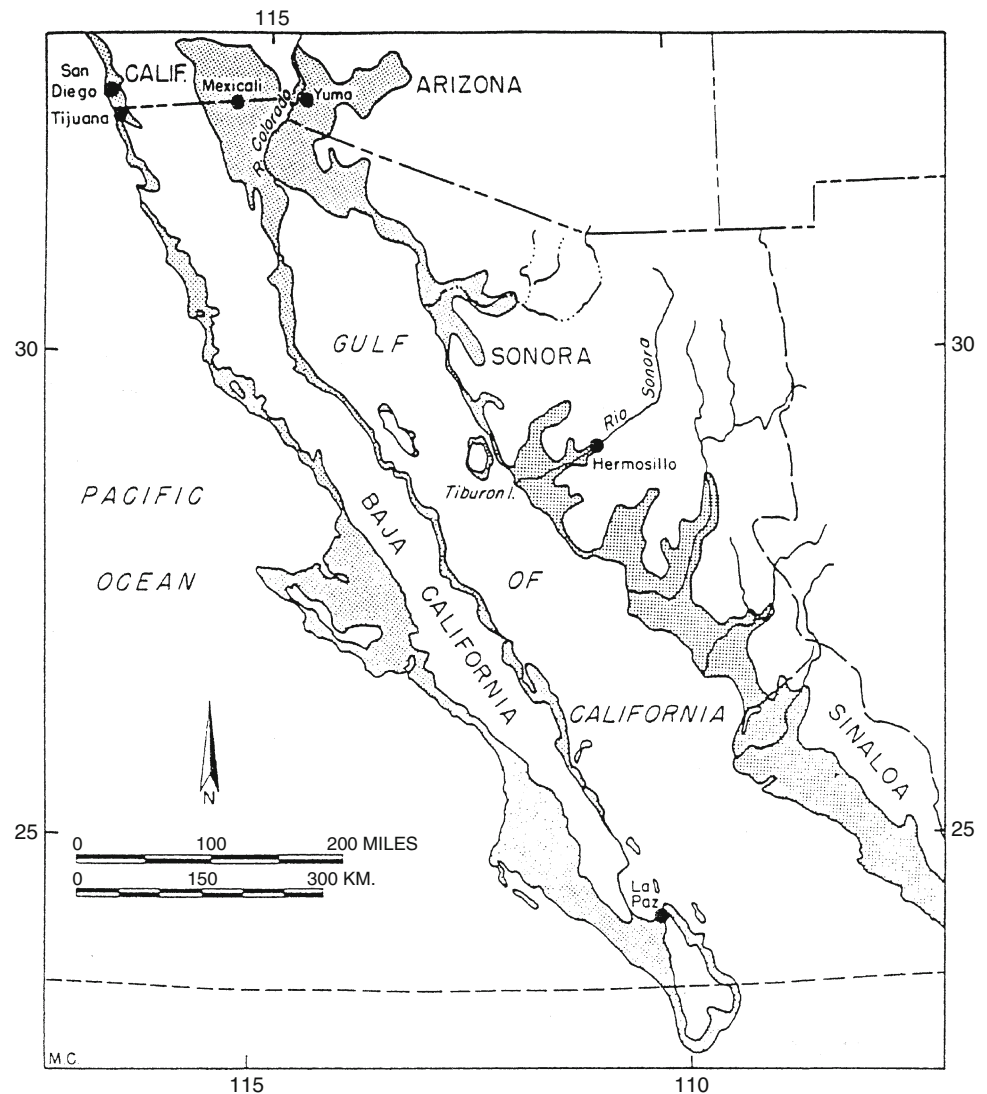


Desert Coasts, Fig. 11 Coastal deserts of southwestern Africa. (Adapted from Meigs 1966)

erratic rainfall, or no rainfall at all, only fog and dew, can occur. Therefore, it is not surprising that animals and plants have adaptations that are correspondingly extreme. Desert animals usually run faster than their relatives in temperate regions. Venomous forms are more poisonous than their allies beyond the desert's fringe. Spiny plants are more prickly, and distasteful ones more unpleasant, than anywhere else.

Most desert and semidesert coastal regions support some degree of vegetation. The smallest amount of plant life exists in areas of rocky *hammada* and cracking clay; but plants can grow even in such soil whenever it is traversed by wadis. And sandy desert usually supports a flora of some kind, except on the dunes, which tend to stay quite bare. Many plants, such as annual grasses and other small herbs, evade the more extreme desert conditions by completing their life cycle during the

Desert Coasts, Fig. 12 Coastal deserts of western Mexico and Baja California. (Adapted from Meigs 1966)



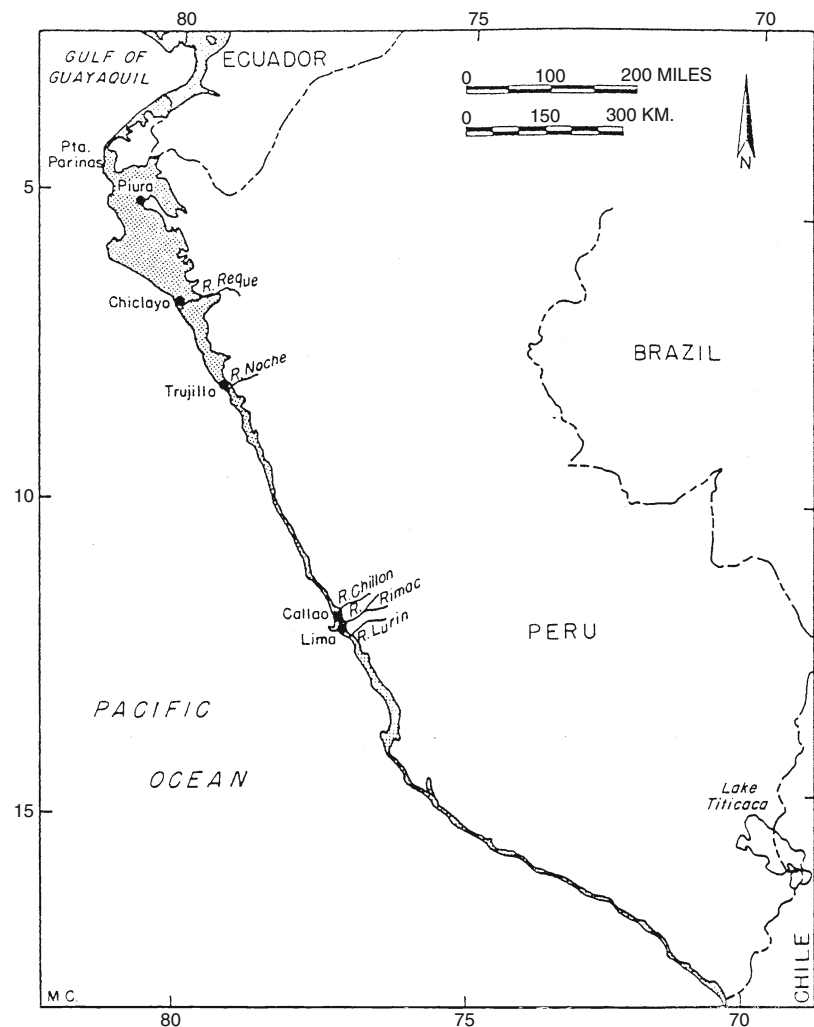
short rainy season and passing the remainder of the year lying dormant as fruits or seeds in the soil. Dry seeds can often survive high temperatures without losing their ability to germinate later under more favorable conditions (Cloudsley-Thompson 1975, p. 26).

Most plants of arid regions have evolved mechanisms not so much of drought tolerance as of resistance to drought. They fight the battle against water loss in a number of ways, often combining different protective devices within a single plant. One means of retaining moisture is to have leaf surfaces coated with fatty substances and resins; these making it harder for water vapor to escape. Other structural adaptations to drought include the presence of dense, hairy coverings (which again keep moisture from escaping) or of pores sunk deep into the leaf tissue so that the dry wind cannot blow directly into them.

Smaller and fewer leaves also help resistance to water loss. And – perhaps most important of all devices – a number of desert plants have developed internal tissues for storing water over long periods of time and for giving increased mechanical support to the bloated, water-filled organism after a heavy rainfall. Such plants (among them the cacti) are known as *succulents* (Cloudsley-Thompson 1975, p. 29).

Some plants can survive on moisture from the air; and at night, with the drop in temperature, dew condenses on their leaves and stems and runs down into the soil around the roots. The welwitschia of the Namib Desert in southwestern Africa is such a plant. It bears two curling leaves, split longitudinally into strips, on which dew condenses, providing water for the shallow roots. The ability of vegetation such as the welwitschia to absorb moisture from the night air, even when dew does not

Desert Coasts, Fig. 13 Peruvian coastal deserts. (Adapted from Meigs 1966)



form, can make vitally needed water available for snails, insects, and other small animals of the desert.

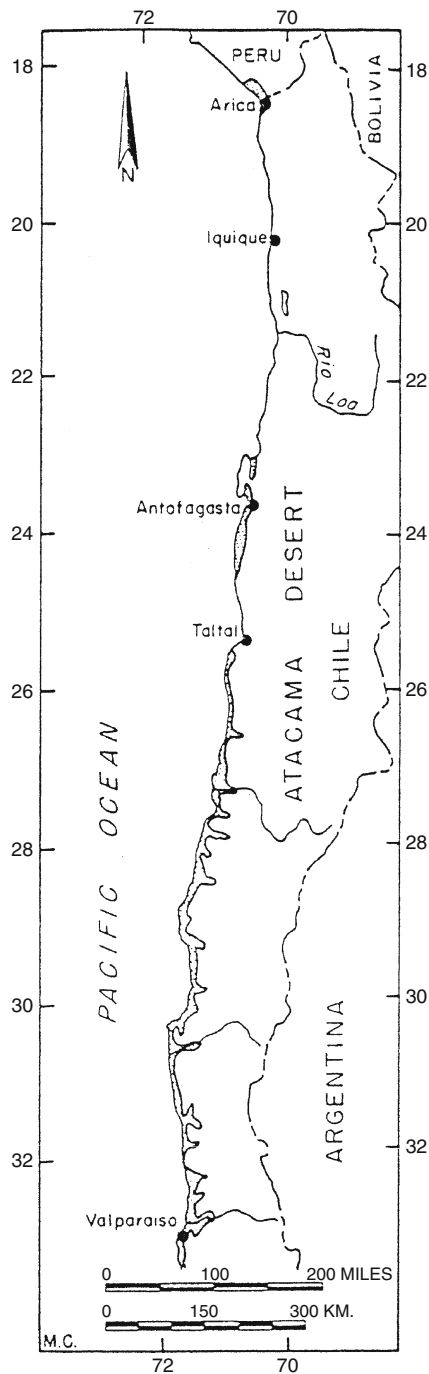
Desert vegetation is not entirely ephemeral. There are many plants that habitually grow where evaporation stress is high and water supply low, and that do not need to evade extreme coastal desert conditions. Such plants, all of which show characteristic adaptations to their environment are termed *xerophytes* (from the Greek words meaning “dry” and “plant”). Among them are trees, perennial grasses, and species that develop bulbs, corms, or tubers. Some xerophytes are characterized by the ability to survive long periods of drought and dehydration of the tissues without suffering injury. They are extremely resistant to wilting, and a few species can lose up to 25% of their water content before they begin to wilt.

Another characteristic feature of most desert coastal plants is that they possess painfully sharp thorns and spikes. The

probable function of these is to afford protection against browsing and grazing animals. Many contain poisonous or irritant latex. Other secrete resins or tannins in the bark or levels; the pods of some contain strong purgatives. All such devices tend to make the desert plants unpalatable to animals (Cloudsley-Thompson 1975, pp. 27–30).

Case Study of the Pakistan–Iran Climate Regime

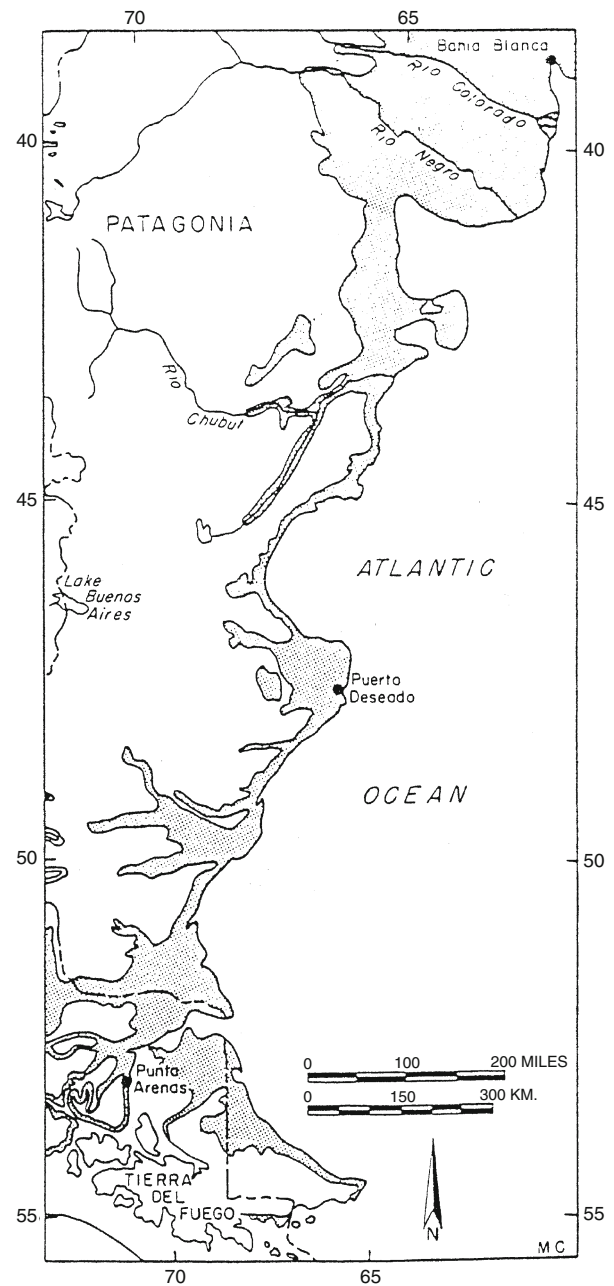
The Pakistan part of this desert coastal type is interesting because it is an area of transition between the Indian summer monsoon system to the east, and the winter cyclonic system of southwest Asia to the west. As a transition area, it receives scanty, unreliable rainfall, averaging less than 10 in. (250 mm) per year from several storm types. Six main weather patterns cross the region, the large subtropical



Desert Coasts, Fig. 14 The Atacama desert. (Adapted from Meigs 1966)

anticyclonic high-pressure cell which predominates most of the year; western depressions originating over the Mediterranean Sea; a modified monsoon pattern; eastern depressions originating over the Bay of Bengal and central India and local thunderstorms and dust storms.

The seasonal division of rainfall for southern Pakistan is shown in Fig. 3. Nearly all of Sind and the eastern part of Baluchistan, particularly the Las Bela area, have a pronounced summer monsoon maximum coming in June, July,

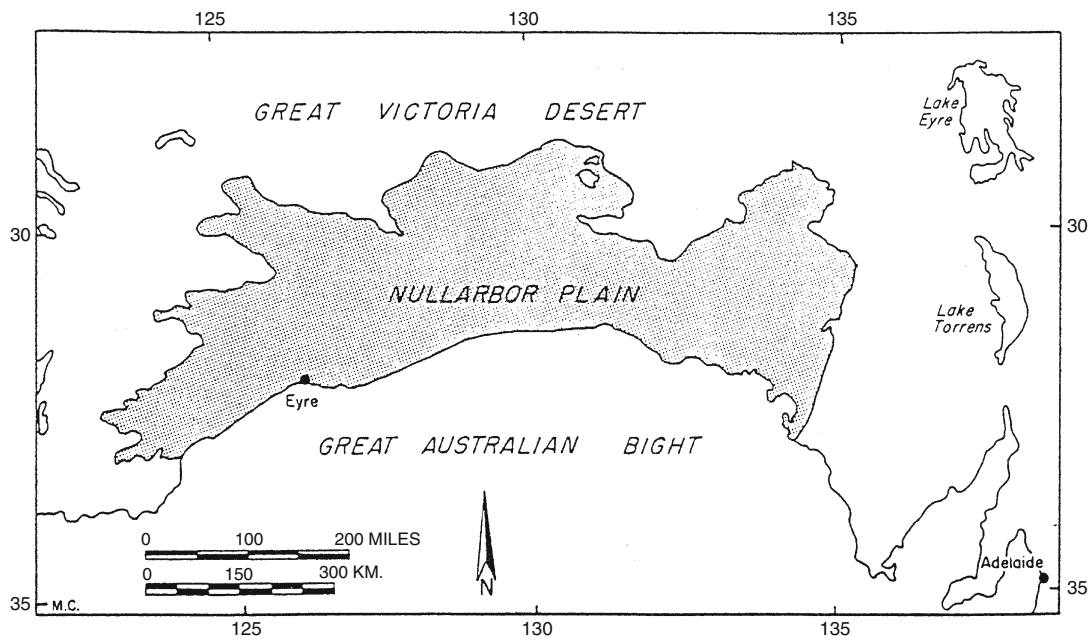
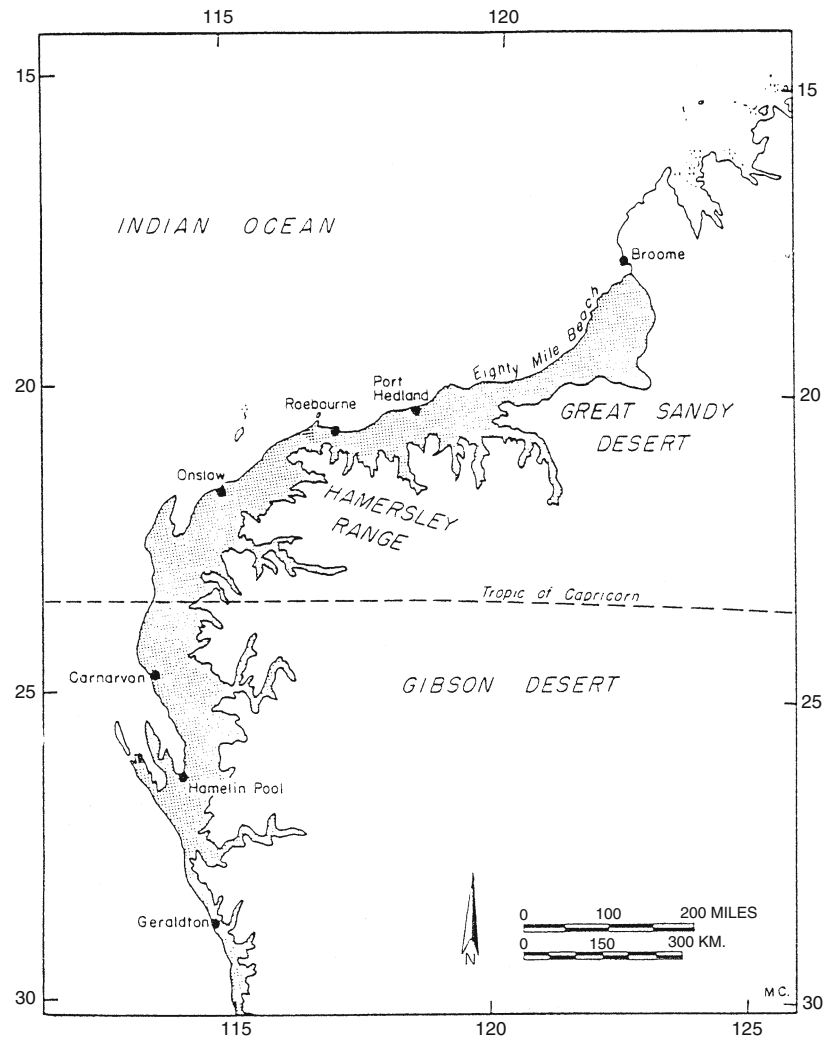


Desert Coasts, Fig. 15 Coastal deserts of Patagonia. (Adapted from Meigs 1966)

and August. This line conforms to the eastern slopes of the Kirthar and Sulaiman mountain ranges which act as a barrier to storms from the east. Iran and the western part of Baluchistan have a winter maximum coming from November to March with very little summer precipitation. Central Baluchistan is a region having both winter and summer rainfall with a slight winter maximum, while a thin band in eastern Baluchistan is an area having both winter and summer rainfall, but with a slight summer maximum.

Further reading on this subject may be found in the following bibliography.

Desert Coasts, Fig. 16 Coastal deserts of western Australia. (Adapted from Meigs 1966)



Desert Coasts, Fig. 17 Coastal deserts of southern Australia. (Adapted from Meigs 1966)

Cross-References

- Africa, Coastal Ecology
- Africa, Coastal Geomorphology
- Asia, Middle East, Coastal Ecology and Geomorphology
- Australia, Coastal Ecology
- Australia, Coastal Geomorphology
- Coastal Climate
- Indian Ocean Coasts, Coastal Ecology
- Indian Ocean Coasts, Coastal Geomorphology
- Meteorological Effects on Coasts
- Middle America, Coastal Ecology and Geomorphology
- North America, Coastal Ecology
- North America, Coastal Geomorphology

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Developed Coasts

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Definition

The term “developed coast” often implies a natural system altered by human action.

The type, frequency, and magnitude of alterations are influenced by social and economic value humans place on the resource. Today, developed coasts are modified and physically maintained to enhance navigation, recreation, and protection of settlements. Human modification of the coast has resulted in changes to the dimensions and locations of coastal landforms and the spatial and temporal scales of cycles of erosion and accretion. Globally, human presence on the coast dates back tens or even hundreds of thousands of years, with some of the earliest documentation on the Mediterranean coast, where deforestation and overgrazing increased the quantity of sediment delivered (Nordstrom 2000). Channelization, diversion, and impoundment of water resources had the opposite effect on the coastal sediment budget; levees and dams decreased sediment delivery from river sources, with the most conspicuous effects occurring in the twentieth century. Grazing initiated dune migration when practiced on the coast.

Direct human alterations to the coast increased in scale with increased occupancy and technological advances associated with the availability of steam power (Marsh 1885) and then the internal combustion engine that further enabled the creation and removal of landforms and redistribution of sediment. Pronounced changes to the landscape occurred about the mid-nineteenth century. Marinas and harbors were constructed, and inlets were regulated through sediment dredging and jetty construction.

Coastal tourism and resort development, beginning in earnest during the second half of the nineteenth century, resulted in further alteration of the coast. Seasonal house construction and infrastructure development (transport and utilities) along the coast resulted in elimination of dunes, while marshes were eliminated by dredging and filling to create navigation channels, marinas, and substrate for structures. Development near the shoreline increased vulnerability to wave attack and flooding during storms, requiring shore protection. Large-scale shore protection efforts, primarily groins and seawalls, were initiated in the late eighteenth

century in Europe and in the mid-nineteenth century in the United States (Nordstrom 2000). Use of these structures, referred to as hard stabilization or armoring, increased with increased construction of buildings and support infrastructures. Considerable attention is now placed on making protection structures more compatible with the geomorphic and ecologic environment where they are constructed (Nordstrom 2014). Beach nourishment (placement of sediment on the beach and nearshore profile from an external source) was first used in the early twentieth century and became popular in the second half of the century. Identified as soft stabilization, nourishment is now the preferred alternative to protect development from storm erosion and flooding in locations where suitable sediment sources are available (National Research Council 2014). Introduction of large volumes of sediment to developed shorelines that previously had a sediment deficit has created opportunities for the restoration of many geomorphic features lost to human development.

Characteristic Landforms

Many of the characteristic landforms found on developed coasts differ from their natural counterparts in size, location, and spatial and temporal scales of change but are identified using similar terminology. The attributes of landforms found on developed coasts can be the intended or unintended outcome of their modification for human use. Over time, assemblages of human-altered landforms may be accepted as natural and require further human modification to maintain their resource value (van der Meulen et al. 1989).

Beaches on developed coasts are generally narrow, because of reduced sediment supply, but nourishment can displace the shoreline seaward and create a beach that is initially wider than a natural beach. Nourished beaches, when designed for shore protection, are often higher in elevation than their natural counterparts, but beaches downdrift of shore protection structures such as groins will be lower. Most direct human modifications to beaches are intended to reduce shoreline mobility over the long term, although mobility may be higher than expected immediately after modifications such as inlet creation or closure or beach nourishment. The characteristics of sediment found on nourished beaches are dependent on the source, which may be derived from offshore, including inlets.

The location and size of dunes are controlled by direct human actions, such as bulldozing and placement of sand fences, and indirect actions through the passive influence of shore protection structures that trap wind-blown sediment (Jackson and Nordstrom 2011). Human modification by fencing, vegetation planting, or mechanical grading creates forms designed to optimize human values such as shore protection,

views of the sea, or ecological enhancement. Small incipient dunes are often lacking because beach wrack is eliminated by beach cleaning operations. Secondary dunes are often lacking because that zone is occupied by human structures. Sedimentary characteristics and internal structure of dunes that form around human structures by aeolian processes can be similar to natural dunes. Dunes created by bulldozing have poorly defined internal structure and contain sediment that may be too coarse or too fine to be transported by local winds. Dunes created by humans are often less mobile than those created by aeolian processes alone because fencing or buildings constructed close to the shoreline restrict migration.

Construction of shore protection structures, introduction of beach nourishment, and management decisions of communities can result in pronounced changes to beach and dune ecosystems (Dugan et al. 2008; Buleri and Chapman 2010). The presence of shore parallel structures such as seawalls and revetments can reduce backshore widths, alter abundance and diversity of species in the intertidal zone, and reduce availability of foraging areas (Dugan et al. 2008), but shore protection structures that incorporate ecological components can increase ecological value of hard stabilization (Buleri and Chapman 2010). Beach nourishment can also have negative effects on both flora and fauna, but adverse effects can be partially mitigated by matching natural and native sediments and timing nourishment during low beach use by birds and other organisms (Speybroeck et al. 2006).

Assessing Change

Landforms on developed coasts respond to energy inputs according to the same laws as natural landforms, but the spatial and temporal scale of change can differ because of human action (Nordstrom 2000). Post-storm recovery of landforms on developed coasts is impeded by houses and infrastructures, but recovery by human modifications can occur rapidly. Closure of inlets created during storms, bulldozing of sediments onto the back beach from overwash deposits on streets, and reconstruction of dunes using sand fences or earth-moving equipment rapidly restore the coast to the pre-storm landscape to achieve social and economic resource values. Human alterations to the coast may be more dramatic in the future in response to the increased hazards caused by sea level rise (Nichols and Cazenave 2010).

Shoreline changes and economically driven human interventions are becoming mutually linked, requiring knowledge of societal action across a range of scales, quantification of anthropogenic effects through time, and communication between scientists and managers (Lazarus et al. 2016). While geomorphology has a strong influence on shoreline change, human actions can be a greater influence on change

along developed coasts (Hapke et al. 2013). Human action can be considered either external to or a part of the natural system (Phillips 1991). As an external influence, human action can be considered an overlay or aberration to a natural coastal system. Human action may now have to be considered an integral process, requiring a better understanding of the functioning of the developed coast and ways to manage it to achieve both natural and human resource values (Nordstrom 2000).

Summary

The morphology of developed coasts is shaped by both natural and human processes. Human alterations of these systems using shore protection structures and beach nourishment have eliminated or changed the dimensions of beaches and dunes and their ecological function. Better understanding of the coupling of human and natural processes on developed coasts can lead to increasing natural and resource values and future management.

Cross-References

- [Beach Nourishment](#)
- [Changing Sea Levels](#)
- [Dune Ridges](#)
- [Sea-Level and Climate Change](#)
- [Tidal Inlets](#)

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Dikes

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Introduction

Dikes, especially sea dikes, are coastal constructions build to avoid flooding. The risk of flooding is detrimental to the safety of people and economic, cultural, and ecological values. This aspect has been of great importance since people first thought about defending their dwellings against flood hazards. In the distant past, dwelling mounts were built to protect families or small communities from the sea. They are known from several low-lying coastal areas in the world, for example, from the North coast of Germany and the adjacent Dutch coast, where they have been occupied ever since 2500 BP, as well as from the chenier coast of Suriname, where they date back to 1800 BP.

Population increase urged to a more active method of flood prevention. The people started to construct dikes to keep the water out of whole regions, thus protecting lives and properties against the sea. In the Netherlands, dikes have been built as a community activity from about AD 1100. Ever since, coastal defense has become more and more an engineering activity. Presently, the design criteria for sea dike construction are determined by three important groups of aspects: hydraulic, subsurface, and construction. The first two groups mainly determine the required height and strength of the dike. The last group of aspects is particularly important for the dike's duration of life. Details about the various aspects are given by U.S. Army Corps of Engineers (1984).

Hydraulic Aspects

Hydraulic factors influencing the marine water level at small temporal scales are the tidal range, wave height, wave run-up, and the setup of the water level due to wind conditions. Because dikes have to be calculated for extreme water

levels, especially a coincidence of extreme conditions of these factors are at stake, for example, high-water spring tide and wave characteristics and setup of the water level during storms. For the establishment of extreme water levels, high-water exceedence frequency curves are used. This method is based on the finding of a systematic relation between the height of a flood-tide level and the number of times this specific level occurs in a century. Extrapolation of the high-water exceedence frequency line enables a probability calculation for the chance that an extreme high-water level, not observed before, may occur. Moreover, the effects of long-term changes in sea level and storm frequency must be taken into consideration.

The final height of a sea dike is determined by the degree of risk of flooding, within the applied economic constraints, the community is willing to accept. In the Netherlands, for instance, the sea defense system is at “Delta height.” However, for the distinguished coastal sections of the Dutch coast this “Delta height” has different values. For the uninterrupted West-coast, which protects the most densely populated and economic heart of the country, the minimum height of the dunes and dikes is determined by a water level exceedence chance of 1 in 10,000 years. For most of the barrier islands in the north of the country the “Delta height” corresponds with a high-water exceedence chance of 1 in 2,000 years.

Subsurface Aspects

Subsurface aspects like subsidence and tectonic movements codetermine the behavior of the sea level. The Netherlands, for instance, are suffering from a relative rise in sea level, which is the combination of the eustatic sea-level rise and the subsidence of the land (Jelgersma 1961). Subsidence in this case is mainly due to the isostatic rebound after the last glaciation and the dewatering of the peaty subsoil in the coastal area for agricultural reasons, resulting in compaction. Especially in the past, this drainage has contributed to the general subsidence. At the same time, the central part of the Dutch coast is influenced by a graben system.

In the decision-making of the final height of a sea dike, the compaction of the subsoil due to the weight of the overlying dike body has also to be considered, as well as some compaction in the dike body itself.

The composition of the subsoil is relevant with regard to groundwater flow. A dike is meant to be impermeable for water. However, groundwater can be pressed to flow through the subsoil underneath a dike, due to a difference in hydrostatic pressure on both sides of the dike. The flow velocity depends on the volume of the overhead and on the permeability of the subsoil. This seepage causes saltwater intrusion landward of the dike. In case the currents are sufficiently

strong to erode the underlying sediment, this process affects the stability of the dike and ultimately may result in the collapse of the dike.

Construction Aspects

The construction of a dike requires building material. In the past, the preferably clayey material was locally dug. Nowadays, however, dikes usually have a sand nucleus, covered by clay to make it impermeable. Suitable clay usually is scarce, whereas sand occurs in large quantities. The advantage of this method is that sand not necessarily needs to be transported on a truck, but can be supplied in suspension by pipeline.

Because of the difference in hydrographic pressure on both sides of a dike, a groundwater table is formed in the sandy dike body. Although the clay cover has to be impermeable for seawater, it must simultaneously be able to let an excess of water from inside the dike through.

Generally, dikes have a delta form. To reduce the effect of wave runup, the slope of the seaward bank of a dike has to be small. On the contrary, the width of a dike increases in that case. A balance has to be found between cost and safety.

In the flood disaster in the Netherlands in 1953, the effect of overtopping water on the landward side of the dikes appeared to be the most important cause of dike failure. In order to reduce the effects of overtopping water, for example, the penetration of water into the center of the dike body, various degrees of unevenness can be applied.

To avoid erosion of the banks and undermining of the dike body revetments are applied. Usually these revetments are provided with a filter layer, to prevent erosion of the underlying material. Permeable as well as impermeable revetments are applied. It is necessary that the material used is flexible and tight to follow the compaction of the underlying sediment.

Cross-References

- [Changing Sea Levels](#)
- [Cheniers](#)
- [Coastal Subsidence](#)
- [Geotextile Applications](#)
- [Hydrology of the Coastal Zone](#)
- [Meteorological Effects on Coasts](#)
- [Polders](#)
- [Sea-Level and Climate Change](#)
- [Shore Protection Structures](#)
- [Storm Surge](#)
- [Tides](#)
- [Waves](#)

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Dissipative Beaches

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Definition and Classification

Dissipative beaches are systems where most wave energy is expended through the process of breaking. Guza (1974) was apparently the first to use the term dissipative beach. He indicated that the wave-energy status of a nearshore system could be determined using the surf-scaling parameter, ε :

$$\varepsilon = \frac{\alpha \omega^2}{g \tan^2 \beta}$$

where α is the wave amplitude at breaking, ω is the wave radian frequency ($\omega = 2\pi/T$, where T is wave period), g is the gravity constant, and β is the beach slope in degrees. The proportion of incident wave energy that is dissipated by breaking increases as ε increases. When ε is less than about

2.5, most wave energy is reflected off the foreshore. For beaches where ε is larger than 20, most energy is dissipated by the turbulence associated with wave breaking. Guza (1974) designated these latter beaches as dissipative. Thus, the relative degree of dissipation or reflection of incident wave energy in a nearshore system has been used as a rationale for the classification of beaches. This is usually accomplished under the rubric of nearshore morphodynamics.

Morphodynamics

The concept of nearshore morphodynamics was developed to characterize systems where form and process are closely coupled through feedback mechanisms (Wright and Thom 1977; Short and Jackson 2013). On beaches, waves ($q.v.$) interact strongly with sediments and morphology, and the form of wave breaking is one manifestation of these interactions. For constant wave steepness, H/L (where H is wave height), the breaker type will change as the nearshore slope changes. On a very low-gradient slope, spilling breakers would be expected. As the gradient increases, there should be a progression through plunging and collapsing breakers. Finally, on very steep beaches, surging breakers should occur (Galvin 1968). For a constant nearshore slope, the same sequence of breaker types will occur as wave steepness decreases, i.e., steep waves are usually associated with dissipative beaches. Breaker type is closely associated with the expenditure of wave energy in the nearshore (e.g., dissipation or reflection) and the development of nearshore morphology. The morphology, in turn, controls breaker type. These relationships are the underlying bases for the concept of

Dissipative Beaches,

Fig. 1 The dissipative beach system at Rosstown, Co. Donegal, Ireland. Note the low-gradient beach slope and multiple lines (at least 10) of spilling breakers. Maximum breaker height is approximately 1.5 m, and period is about 7 s



nearshore morphodynamics (see summary by Wright and Short 1984). The recognition of characteristic sets of dynamic relationships provides the basis for using morphodynamic regimes (or states) as a means for classifying beach types. For example, spilling breakers tend to occur on dissipative beaches. This contrasts with reflective beaches (*q.v.*), where collapsing or surging breakers are common. Plunging waves tend to occur on the intermediate beach states of the morphodynamic model, where neither dissipation nor reflection dominates the response of nearshore morphology.

Characteristics of Dissipative Beaches

In cross section, morphodynamically dissipative beaches display the classic forms of “storm” or “winter” beach profiles (e.g., Sonu and Van Beek 1971). According to Wright and Short (1984), other distinguishing characteristics include low-gradient nearshore and beach slopes, about 0.01 and 0.03, respectively, and with relatively fine sediment sizes. Dissipative beaches tend to have a substantial sediment volume and at least one linear nearshore bar, although multiple bar-trough systems are common. In dissipative systems, the beach and nearshore zone will exhibit minimal alongshore variability – the system is approximately two-dimensional. Incident wave energy is maximum at the break point and decreases shoreward. The classic dissipative system displays several coincident sets of spilling breakers (Fig. 1), and there is minimal energy remaining at the landward extremity of uprush. Infragravity motion dominates the inner surf zone and frequently causes linear scarping of the foreshore (e.g., Short and Hesp 1982). On meso- and macrotidal beaches, the nearshore may be dissipative only at lower tidal stages and reflective at high tide (Short 1991; Masselink and Hegge 1995). Short and Hesp (1982) have also linked the dissipative beach state to the formation of large-scale transgressive dune sheets – at least in the Australian context. This linkage was a key development for the derivation of later models of beach-dune interaction (e.g., Sherman and Bauer 1993). Coastal ecologists have linked beach states with species richness and abundance (Defeo and McLachlan 2011), finding both to be enhanced in macrotidal dissipative systems.

Cross-References

- Bars
- Beach Features
- Beach Processes
- Reflective Beaches
- Sandy Coasts
- Surf Zone Processes
- Waves

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Dredging of Coastal Environments

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Abbreviations

CEDA	Central Dredging Association (The Netherlands)
D&PC	Dredging and Port Construction (UK)
IADC	International Association of Dredging Companies (The Netherlands)
PE	Port Engineering Management Magazine (UK)
WEDA	Western Dredging Association (USA)
WODA	World Organization of Dredging Associations (The Netherlands)

Definition of Dredging

Dredging is moving material submerged in water from one place to another in water or out of water with dredging equipment. Dredging is applied in projects of navigation, for maintenance of beaches, for reclamation, to secure construction materials, and for environmental dredging.

Per Bruun: deceased.

Dredging for Navigation

From the very beginning of civilization, people, equipment, materials, and commodities have been transported by water. Ongoing technological developments and the need to improve cost-effectiveness have resulted in larger, more efficient ships. This, in turn, has resulted in the need to enlarge or deepen many of the rivers and canals, our “aquatic highways,” in order to provide adequate access to ports and harbors. Nearly all the major ports in the world have at some time required new dredging works—known as capital dredging—to enlarge and deepen access channels, provide turning basins and achieve appropriate water depths along waterside facilities. Many of these channels have later required maintenance dredging, that is, the removal of sediments which have accumulated in the bottom of the dredged channel, to ensure that they continue to provide adequate dimensions for the large vessels engaged in domestic and international commerce.

Dredging for Construction, Reclamation, and Mining

Dredging is an important way of providing sands and gravels for construction and reclamation projects. In the last two decades, the demand, and the associated extraction rates, for such offshore aggregates have significantly increased. Dredged aggregates have a wide range of uses.

Dredging for the Environment

Dredging can be undertaken to benefit the environment in several ways.

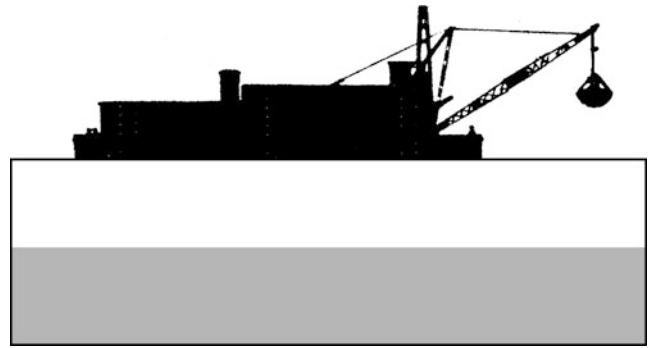
Dredged materials are frequently used to create or restore habitats. Recent decades have also seen the increasing use of dredged materials for beach replenishment.

Another environmental use of dredging has been in initiatives designed to remove contaminated sediments, thus improving water quality and restoring the health of aquatic ecosystems. This so-called “remediation” or “cleanup” dredging is used in waterways, lakes, ports, and harbors in highly industrialized or urbanized areas.

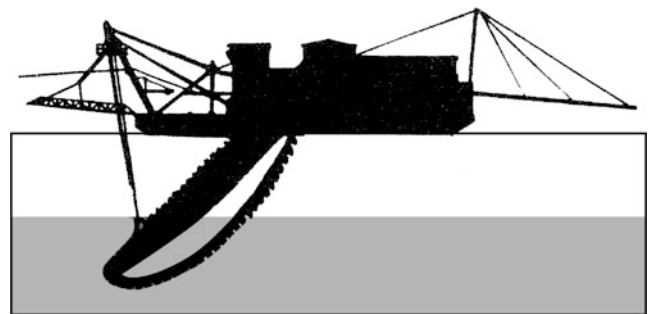
Types of Dredgers

Dredgers come in two different classes: mechanical and hydraulic.

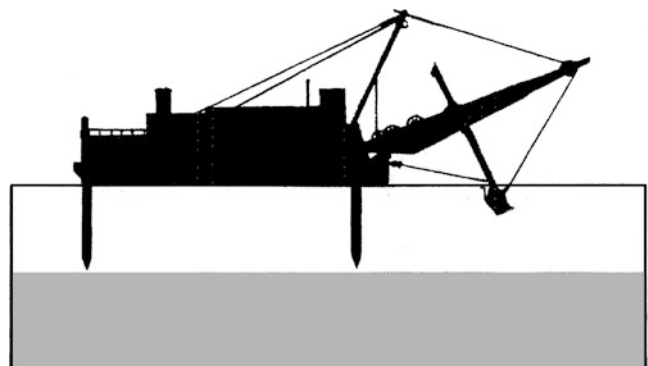
Mechanical types include the *clamshell* or *grapple dredges* (Fig. 1). Larger dredges of that type are no longer favored. The endless *chain bucket* (bucket-ladder) dredge (Fig. 2) was



Dredging of Coastal Environments, Fig. 1 The clamshell or grapple dredge. (Adapted from Bruun 1990b)



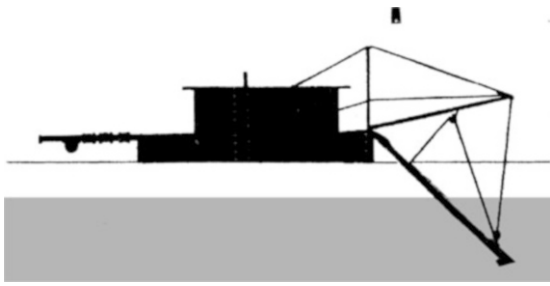
Dredging of Coastal Environments, Fig. 2 The bucket-ladder dredge. (Adapted from Bruun 1990b)



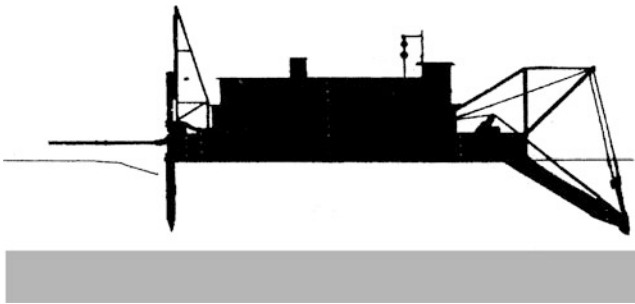
Dredging of Coastal Environments, Fig. 3 The dipper dredge. (Adapted from Bruun 1990b)

used widely earlier in Europe but not in the United States, although they were employed on a few projects like the Panama Canal. For environmental reasons they may still be placed back on the market.

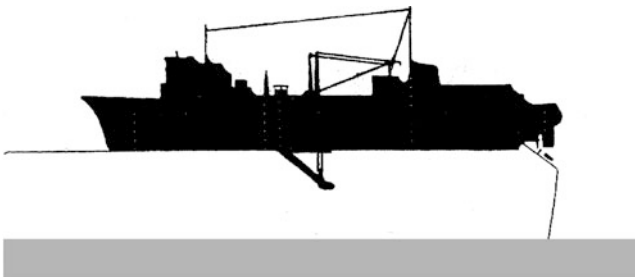
The *mechanical dipper dredge* with its heavy bucket (Fig. 3) moved by a very strong arm and boom is still used for dredging relatively loose (usually not solid) rock. The bucket may be provided with special cast iron teeth. Its volume may be several cubic meters.



Dredging of Coastal Environments, Fig. 4 The plain suction dredge. (Adapted from Bruun 1990b)



Dredging of Coastal Environments, Fig. 5 The cutterhead pipeline dredge. (Adapted from Bruun 1990b)



Dredging of Coastal Environments, Fig. 6 The self-propelled hopper dredge. (Adapted from Bruun 1990b)

The *hydraulic dredge* is the most important piece of dredging equipment. The *plain suction dredge* (Fig. 4) has no cutter but sucks material off the bottom and discharges it through a stern connected pipe leading to a spoil disposal area.

The *cutterhead pipeline dredge* (Fig. 5) has a rotating cutter on the end of the ladder and excavates the material from *in situ* condition and discharges it through the stern to pontoon and shore pipe. The dredge is controlled on stern-mounted spuds and is swung from one side of the channel to the other by means of a swing gear.

The *self-propelled hopper dredge* (Fig. 6) has a large hopper in which the dredged material is loaded for later dumping through doors in the bottom. This type of dredge is normally employed where the water is too deep for a pipeline

dredge or where spoil areas for such a dredge are not available within economic distances.

The hydraulic dredge has, without any doubt, become the most important piece of equipment in the entire harbor engineering field. Without the dredge, commercial navigation of waterways and rivers would be ended. Waterborne industry would collapse. Ocean shipping as it is known would be nonexistent.

Hydraulic dredges dig canals, ports, and harbors, do maintenance dredging in rivers, canals, and waterways, and excavate for construction of piers, wharves, docks, dams, and underwater foundations. They provide spoil for the reclamation of swamps and marshes; they construct dikes and levees, and dredge sand, gravel, and shell, as well as coal, gold, diamonds, and many other minerals for commercial purposes. The dredge's scope of operation is broad. Most of them have cutterheads.

Small hydraulic dredges, 20–40 cm diameter pipe, operate in water only a few meters deep. Large dredges, 65–90 cm diameter pipe, require more draft but can dig to greater depths. With the aid of booster pumps the distances solids can be pumped are unlimited.

Although the dredge's output is understandably greater in soft materials than in hard, it can excavate almost anything. It digs mud, silt, loam, clay, sand, hardpan, gravel, coral, and even rock. Boulders weighing 500 kg and more have been excavated and transported by dredges.

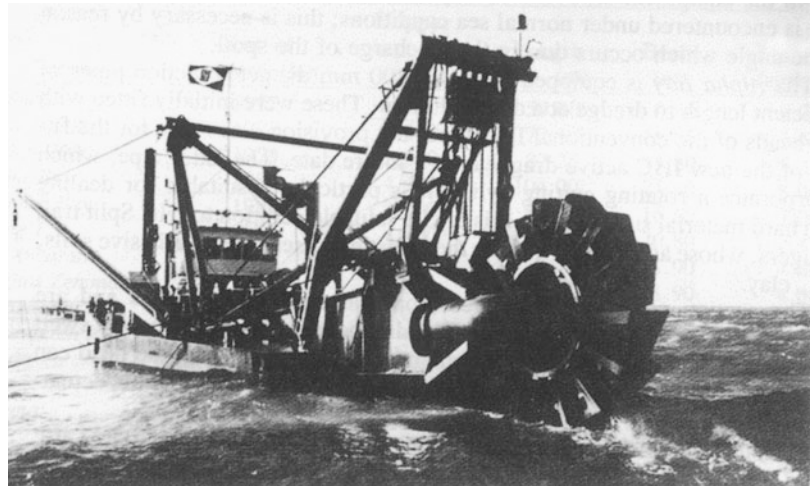
The self-propelled hopper dredge moves freely without any kind of mooring system. It can operate by pumping holes in the bottom side by side through a usually circular intake which is the reinforced end of the suction pipe, or it can trail along with its trailing suction head attached to the end of the suction pipe "vacuum cleaning" the bottom. This method is more practical and effective than the plain pipe sucking procedure where deep holes may become sediment traps for material which otherwise may have been flushed away. Various organizations like the IADC (International Association of Dredging Companies), the CEDA (Central Dredging Association), the WEDA (Western Dredging Association), and the WODA (World Organization of Dredging Association) organize conferences and publish magazines (See "Abbreviations").

Special Dredgers

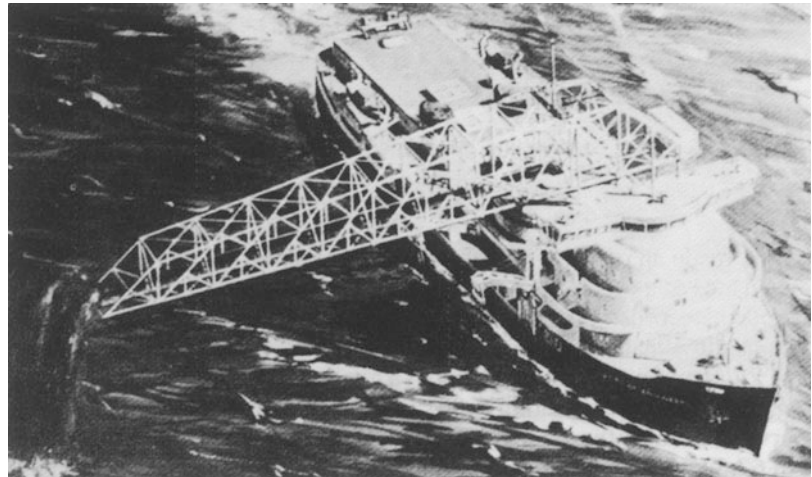
Wheel Dredgers

The IHC's (The Netherlands) Scorpio Dredger has a dredging wheel that consists of bottomless buckets to which sticky soil does not adhere (Fig. 7). They are arranged to form a tunnel from which dredged material cannot easily escape. This effectively prevents the ingress of debris. The result is higher solids concentration, no spillage or blockage, and reduced downtime. Several such dredgers are now in operation all over the world.

Dredging of Coastal Environments, Fig. 7 The Wheel dredger SCORPIO. (Adapted from Bruun 1990b)



Dredging of Coastal Environments, Fig. 8 The sidecaster and hopper dredge McFarland. (Adapted from Bruun 1990b)



The Water Injection Dredger

The Water Injection Dredge (WID) was developed in Europe in the mid-1980s and has been used regularly there since 1987 and recently tested in the United States by the US Army Corps of Engineers (USACE). The principle of WID is to fluidize the shoaled material to a condition that a gravity-driven density current is formed, which transports the sediment down a slope into deeper water where it is no longer an obstruction to navigation, and natural sediment transport processes can take over. This density current remains close to the bottom, minimizing turbidity. Fluidization is accomplished by injecting water directly into the sediment voids under relatively low pressure, through a series of nozzles evenly spaced on a horizontal pipe. A pump mounted on the dredge, barge, or other vessel forces water into the horizontal pipe or head. The pipe pivots from the dredge, allowing the head to be lowered with the aid of a winch. During operation, the head with the injection nozzles is as close to the bottom as possible, while the dredge proceeds across the cut area (Bruun 1990a, 1996).

Sidecasters

Sidecasters sidecast the dredged materials. They come in various sizes. Small sidecasters are excellent for emergency and shallow water operations, for example, in channels.

The McFarland (Fig. 8) also has a hopper. It was put in commission in 1967 and assigned to maintenance of Gulf Coast inlets (USACE). It is 91 m long, 22 m wide, with a loaded draft of about 7 m to permit efficient operation in relatively shallow water.

Other types of dredgers are mentioned in the following section on the latest development of dredging equipment.

The Latest Development in Dredging Equipment

The latest decades and years have produced dredging equipment on a scale never seen before including hopper dredgers of capacities up to 33,000 m³, with suction pipes of 1.4 m diameter reaching down to 60 m depth, with deep dredging

suction up to 131 m depth, and outputs exceeding to 10,000 m³/h. Large projects in the Far East (Hong Kong, Singapore, Thailand, China, Persian Gulf) and also in South America (Argentina, Uruguay) utilize these dredges. Large fills for industrial development including airports, deepening of navigation channels and basins all over the world are in progress or planned. This includes many projects in the United States (New York, Charleston, Savannah, Galveston, Los Angeles, Columbia River, etc.). Materials are used for fills or dumped offshore in designated areas.

The dredging companies involved are mainly Dutch or Belgian. They compete in producing the most effective equipment provided with the latest inventions in survey technique, controls, and maneuvering. They are applied in projects with quantities of tens of millions of cubic meters capital dredging. Smaller size equipment includes hopper dredgers of a few thousand cubic meters hopper capacities for relatively smaller projects, capital as well as maintenance.

Some larger cutterhead hydraulic pipeline dredgers are still being built, in particular for work on harder bottoms and in narrow or protected environments. Some of these are ocean-going with discharge pump capacities of many kilometers. An example is the Bos Kalis 89 cm "ORANJE."

Amphibious "crawl cats" and "crawl dogs" (Bruun 1996) and underwater pumps like the PUNAISE (Visser and Bruun 1997) have been built and tested in practice.

One of the most recent additions to the international fleet is the "LANGE WAPPER," a 13,700 m³ capacity trailing suction hopper dredger which entered service in May 1999. The 20,000 dwt new vessel boasts the extremely shallow draft of 9 m fully laden, setting it in competition with vessels in the 8,000–10,000 m³ range-yet with substantially higher production rates.

Built by IHC Holland, the Lange Wapper is 120 m long, with a beam of 26.8 m. The total installed power is 13,400 kW and dredging depth is in the range 30–50 m. For a draught of 8 m, the vessel can carry a deadweight of 15,500 t.

The Lange Wapper features state-of-the-art computer software to integrate all dredging, navigation, and surveying systems. It is thought to be the first software of its kind in the industry.

Specially designed by DEME, together with IHC Systems (the electronics subsidiary of the IHC group), the software provides a single program capable of handling all data relating to dredging, navigation, and surveying. Previously, software systems could only communicate with each other from standalone packages. Furthermore, up to 10 extra

systems were required onboard and extra personnel needed to operate them. The software onboard the Lange Wapper will be available via a number of networked computer terminals and will also be able to access additional data by satellite.

Enhanced controls and accuracies are the main advantages of this new system. Tight dredging parameters will be possible, which will enhance performance in environmentally sensitive conditions and offshore trenching or pipe covering operations.

All dredging projects, however, depend upon hydrographic surveys. As ports and channel systems throughout the world continue to involve expansions and extensions, the need for effective survey programs becomes essential. Throughout the past year there have been some major developments within this sector of the port construction industry.

Dredging for Nourishment of Beaches

Artificial nourishment of beaches is the most practical and economical measure of coastal protection. Suitable material must be available. Figure 9 shows split-hull dumping of dredged material. Figure 10 is a trailing suction hopper dredge pumping material on the beach and in the nearshore bottom, or through a buoy connected to shore by a (submerged) pipeline. In the United States, nourishment is often done by a hydraulic pipeline dredge connected directly from the borrow area to the beach by (submerged) pipeline.

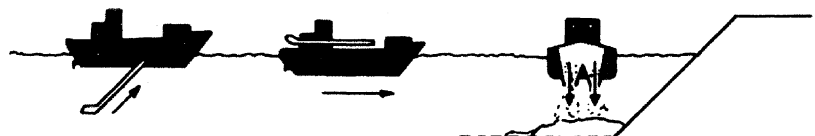
Profile Nourishment

The normal nourishment profile includes a wide horizontal berm with a front slope of about 1 in 10. Because beach nourishment is generally paid for by a public authority, public opinion is important in acquiring sufficient funds. Therefore, it is wise to design beach nourishment in such a way that the public sees that the beach is somewhat wider after the nourishment, but that there is no major adaption in the beach shape during the first storms in autumn. If the purpose of the nourishment is to make a wide recreational beach, this is very important. If the purpose is to prevent flooding, the best place is as high as possible on the beach. If the purpose is to combat chronic erosion, the best place is in the breaker zone.

Profile nourishment (Bruun 1990a, b) attempts to follow the natural profile in the nearshore as much as this is practically possible. The advantage of profile nourishment is that it decreases the introductory losses of fill producing a more stable condition. OVER THE BOW PUMPING (Fig. 10) is most suitable for such operations undertaken in countries like

Dredging of Coastal

Environments, Fig. 9 Direct dumping in the nearshore zone with hopper dredger. (From Bruun 1990b)





Dredging of Coastal Environments, Fig. 10 Trailing hopper dredger pumping through nozzle placed in bow of the vessel. (From Bruun 1990b)

Holland, Denmark, Germany and at its start in the United States. The Shallow Water Hopper dredgers, in short supply in the United States, are well suited for such operations. They may pump over the bow, if bottom is steep enough, or through a short pipeline (Bruun and Willekes 1992).

Dredging and the Environment

Introduction

The development of navigation requires still deeper and wider channels and basins, as well as fills for industrial harbor and other developments. This, in turn, carries with it the needs for major disposal areas for dredged materials.

Maintenance of depths requires removal of deposited materials which are sometimes or often polluted.

Beaches and shores are eroding and need protective nourishment by clean material which will then have to be relocated. Dredging may also be a useful tool for remedying past environmental interference. However, by its very nature, the act of dredging and relocating dredged material has an environmental impact. It is, therefore, of the utmost importance that we should be able to determine whether any planned dredging will have a positive or negative impact on our environment. Evaluation of environmental impact should examine both short- and long-term effects, as well as the sustainability of the altered environment.

While spoiling of material from dredging of tidal inlets on sandy shores has involved little difficulty because the material was clean and ample offshore dumping areas were available, dredging of harbors, estuaries, bays, lagoons, and waterways without tidal flushing, has gradually caused many problems with respect to the location of spoil areas.

In either case, a severe disturbance of the area's biology has resulted. This has necessitated a great number of regulations on the extent of such operations.

When using dredging techniques, we must be aware of the environmental effects of the changes we are trying to achieve, as well as the effects of the dredging activity itself which may include: alterations to coastal or river morphology; addition or reduction of wildlife habitat; changes to water currents and wave climates, which might affect navigation, coastal defense, and other coastal matters; reduction or improvement of water quality, affecting benthic fauna, fish spawning, and the like; improvement of employment conditions owing to industrial development; and removal of polluted materials and their relocation to safe, contained areas.

The marine environment is a complex combination of natural features and phenomena, supporting a diverse but largely concealed population. Because of this complexity it is extremely difficult to predict the effects of human-induced changes and short-term operations.

Past environmental ignorance has resulted in many rivers, ports, and harbors in the industrialized nations containing soils that have been contaminated by undesirable levels of metals and chemical compounds. When dredging in these soils, contaminants may be released into the water column and thence into the food chain. Thus, the environmental effects of dredging and relocation of the dredged material may be more severe and will require more detailed analysis. In certain cases, it is the very existence of the polluted soils that has led to dredging: by removing the polluted soils and relocating them to a more secure situation, the environment is improved. Long-term improvement does, of course, depend ultimately on preventing pollution at its source. The treatment and storage of polluted soils is a highly complex subject and requires detailed study.

Many rules and regulations have been issued regarding the handling of polluted materials. International conventions, for example, the London 1972 meeting (Bruun 1990a), have resulted in published advice. The PIANC (Permanent International Association of Navigation Congresses) has issued a number of guidelines including:

- Report of the International Commission. Study of Environmental Effects of Dredging and Disposal of Dredged Materials. Supplement to Bulletin No. 27 (1977)
- Report of the International Commission. Classification of Soils and Rocks to be Dredged. Supplement to Bulletin No. 47 (1984)
- Report. Disposal of Dredged Material at Sea. Supplement to Bulletin No. 52 (1986)
- Beneficial Uses of Dredged Material (1992)
- Dredged Material Management Guide (1997)
- Handling and Treatment of Contaminated Dredged Material, Vol. 1 (1997) and Vol. 2 (1998)
- Management of the Aquatic Disposal of Dredged Material (1998)
- Confined Disposal Facilities for Contaminated Dredged Material

The ADC (Association of Dredging Companies) and the CEDA have printed several reports, here particularly noted and referred to is the 1999 report on "Environmental Aspects

of Dredging.” In the United States, the USACE, Coastal Engineering Research Center in Vicksburg, MS, through its Dredging Research Division has published a great number of reports as mentioned later.

Management

Management alternatives for dredged material can be grouped into the following four main categories (listed in order of significance): beneficial use, open-water disposal, confined disposal, and treatment.

Beneficial Use

The definition of “beneficial use” according to the Dutch ADC and CEDA Guide #5, 1999, is “any use which does not regard the material as a waste.”

Dredged material is increasingly regarded as a resource rather than as a waste. The London Convention’s Dredged Material Assessment Framework (DMAF) recognizes this by requiring possible beneficial use of the material to be considered before a license for sea disposal may be granted.

Open-Water Disposal

Open-water disposal is the placement of dredged material at designated open-water sites in oceans, estuaries, rivers, and lakes such that the dredged material is not isolated from the adjacent waters during placement. Open-water disposal generally involves placement of clean or mildly contaminated material.

Open-water disposal of highly contaminated material can also be considered with appropriate control measures. This category includes unrestricted placement on flat or gently sloping waterbeds in the form of mounds or placement with lateral containment (e.g., depressions). For contaminated material, a cap of clean material can provide isolation from the benthic environment. If capping is applied over the mound formed by unrestricted placement, it is called level-bottom capping (LBC). If the capping is applied with lateral containment, it is called contained aquatic disposal (CAD).

Confined Disposal

Confined disposal is the placement of dredged material in an engineered containment structure (e.g., dikes, natural, or constructed pits) enclosing the disposal area and isolating the dredged material from surrounding waters or soils during and after placement.

Treatment

Treatment is defined as a way of processing contaminated dredged material with the aim of reducing the amount of contaminated material or reducing the contamination to meet regulatory standards and guidelines. Each project may

raise environmental concerns and not only when the material in question is contaminated. For instance, the “beneficial” use of clean, beach-compatible sand for beach nourishment may damage well-established sensitive marine habitats or species. Or the unrestricted open-water placement of clean material may have unacceptable direct physical impacts, such as smothering of bottom-dwelling organisms or local increases in suspended solids concentrations.

Assessments

The environmental assessment of management alternatives (with controls) should consider all potential impacts. Many countries require Environmental Impact Assessment (EIA) studies to be undertaken and the results must be documented in formal Environmental Impact Statements. (EIS.) Even if the law does not require such studies, it is advisable to include them in the project planning as EIS may prove to be very useful in the permitting procedure and in gaining public acceptance for projects.

For decision making, the following eight steps are identified in the Dutch Guide #5, by the IADC and WEDA (1999): establishing the need for dredging, characterization of dredged material, assessment of beneficial use options, preliminary screening of potential disposal alternatives, detailed assessment of disposal alternatives, selection for final design and implementation, permit application and processing, monitoring program design.

The World Bank (1990) distinguishes five main types of dredged sediments:

1. Material from maintenance dredging of ports, harbors, and navigation channels usually consist of fine-grained silt and clays and is often rich in organic matter and contains contaminants.
2. Material from maintenance dredging of sand bars at the entrances to harbors or channels usually consist of fine- to coarse-grained sand and is much less contaminated than (1).
3. Material from capital dredging within a port may vary considerably along the vertical profile. The upper layer is usually fine-grained, organic-rich, and contaminated, and the underlying layer is coarsegrained and uncontaminated. In some circumstances, even the deeper layers might contain elevated levels of contaminants from historical discharges.
4. Material from capital dredging of channels or outer harbor areas is likely to be coarse-grained and uncontaminated.
5. Material derived from remediation dredging is usually fine-grained, organic-rich and, by definition, highly contaminated. Its relocation requires special control measures.

Physical Properties

The basic physical characteristics are form and composition, grain size, specific gravity, plasticity, water content, shear

strength, water-retention characteristics, permeability, settling behavior, consolidation behavior, compaction, and organic content.

Chemical Properties

They include pH-values, calcium carbonate equivalent, cation exchange capacity, salinity, redox potential (electrified activity), dissolved oxygen, biochemical oxygen, organic carbon, nutrients, carbon/nitrogen ratio, potassium, and contaminants of any kind.

Biological Properties

Biological characterization of sediments may involve testing for the presence of microbial constituents and testing for toxicology.

Microbial constituents of concern include pathogens, viruses, yeasts, and parasites. Sediments should be tested for these constituents whenever dredging sites are close to sewage discharges and placement sites are close to sensitive areas such as drinking water intakes, recreation beaches, or shellfish beds.

Acute toxicity bioassays test the effects of short-term exposure. Toxicity is expressed as median lethal concentration (LC_{50}), that is, concentration which theoretically would kill 50% of the test organisms within a specified time span. The duration varies from hours to days.

Chronic toxicity bioassays assess sublethal effects which can result from prolonged exposure to relatively low concentrations. Such effects include physiological, pathological, immunological, teratological, mutagenic, or carcinogenic effects. Chronic toxicity bioassays may take several weeks. Toxicity bioassays may be used to evaluate: potential impacts of dissolved and/or suspended sediments on water-column organisms (elutriate bioassays); and potential impacts of deposited sediments on benthic organisms (benthic bioassays).

Metals

Metals may exist in the aquatic environment in four basic, but interactive forms. These dissolved forms are: free metal ions; complexed molecules; particulate forms absorbed to the surface of solid particles; and precipitates (mostly sulfides).

In free mode (the most toxic), metals are transported with the water and easily taken up by organisms. The particulate associated forms are relatively inactive. In sediments under natural conditions (anoxic, reduced, near neutral pH) only a small fraction of heavy metals is dissolved; the major portion is bound by sulfides or by structurally complex, large organic compounds. Release of these may be induced by increased salinity, reduced pH, and increased redox potential.

Sediment Properties Versus Dredging Placement Methods

In addition to the original properties at the dredging site, the behavior of sediments during and after placement will also

largely depend on the specific characteristics of the dredging and placement techniques.

Dredging techniques may be subdivided into two categories: mechanical and hydraulic dredging.

Mechanical dredgers (e.g., bucket or clamshell) use mechanical force to dislodge, excavate, and lift the sediments. The material is generally placed and transported to the site of discharge by barges. Barges may have bottom doors or split-hulls and the material can be released within seconds as an instantaneous discharge. Cohesive sediment dredged and placed this way remains intact, close to the *in situ* density through the whole dredging and placement process.

Hydraulic dredgers remove and lift the sediments as a slurry (mixture of sediment solids and large amounts of dredging site water). The slurry is transported to the discharge site via pipeline or in hoppers and released with large amounts of entrained water. Hoppers may release the material through bottom doors or through pump-out operations. Hydraulic dredging and transport break down the original structure of sediments.

Beneficial Use of Dredged Materials

The main application is for fills. Recent years have also introduced a number of practical structural usages.

Publications providing guidance on beneficial use include: Beneficial Uses of Dredged Material; Engineer Manual (USACE, 1986); Beneficial Uses of Dredged Material, A Practical Guide (PIANC, 1992); Handling and Treatment of Contaminated Dredged Material (PIANC, 1997); Guidelines for the Beneficial Use of Dredged Material (HR Wallingford, 1996).

Engineering Usages

For practical use of dredged materials, one may distinguish between rock, gravel, sand, silt, mud, and clay, differing greatly in grain sizes and shape. Sand and silts may become "quick." Clay is cohesive.

Gravel and sand are dredged for construction purposes. Different kinds of heavy minerals are extracted from sands in sometimes major operations, including from the deep sea (Bruun 1990a, b).

Coastal Protection

Dredged material is used for artificial nourishments of beaches which today is the most common coastal protection measure. The material dredged offshore or in entrances and navigation channels is pumped ashore if suitable for widening and raising the beach elevation.

Materials may also be used to produce protective offshore berms which may be designed as feeder berms, hard berms, and soft berms.

Feeder Berms

Feeder berms produced by dumping of materials nearshore serve partly as source of materials for a possible transfer of materials to the beach.

Hard Berms

The construction of hard berms involves the placement of suitable material, in depths up to 13 m, to create a permanent feature on the seabed approximately parallel to the shoreline with gentle side slopes that will intercept the troughs of incoming storm waves and decrease erosion of the shoreline. If the berm is made of sand, it will be modified in profile and some of the material will be lost, implying that some maintenance will be necessary. This technique may still be cheaper and more convenient (i.e., causing less disturbance to recreation) than direct beach nourishment.

Soft Berms

Naturally occurring underwater mudbanks are known to absorb water wave energy and thereby attenuate waves that pass over them. Energy reductions of 30–90% are not uncommon even in the absence of wave breaking. Attempts have been made to replicate this action by creating underwater mud berms from dredged material.

Open-Water Disposal

Material dredged may be disposed of in deeper waters or in shallow designated waters. Bypassing at tidal entrances is a common usage. Offshore disposals may be capped to protect the material from being eroded away. Polluted materials may be placed inside dikes. Figure 11 shows the large “Slufter” used by EUROPORT for disposal of dredged materials which are polluted. Land disposals may also use dikes. An

impermeable membrane is then placed to avoid penetration of pollutants to the surrounding areas. Material may also be placed in dredged pits in the bottom provided with a proper protective cover.

The Dredging Research Program in the United States by the USACE

The USACE is involved in virtually every navigation dredging operation performed in the United States. The Corps' navigation mission entails maintenance and improvement of about 40,000 km of navigable channels serving about 400 ports, including 130 of the nation's 150 largest cities. Dredging is a significant method for achieving the Corps' navigation mission. The Corps dredges an average annual 230 million m³ of sedimentary material at an annual cost of about US \$400 million. The Corps also supports the US Navy's dredging program in both maintenance and new work.

The Corps will continually be challenged in pursuing optimal means of performing its dredging activities. Implementation of an applied research and development (R&D) program to meet demands of changing conditions and generation of significant technology, adopted by all dredging interests, are means of reducing the cost of dredging the nation's waterways and harbors.

The Shallow Water Hopper dredger has always been on their agenda. There are a couple of additions to the fleet financed by private industry. They are of medium-size with about 5,000 m³ hopper capacity.

Reports on the development of dredging and on the Environmental Effects of Dredging, Technical Notes, are published regularly, the latter category covering the subjects

Dredging of Coastal

Environments, Fig. 11 The “Slufter” offshore diked disposal are used by EUROPORT (Rotterdam) for polluted materials. (Adapted from Bruun 1990b)



of Aquatic Disposal, Upland Disposal, Wetland Estuary Disposal, Regulatory Management, Beneficial Uses, Equipment, and Miscellaneous as described in numerous reports. Copies may be acquired from the USACE, Waterways Experiment Station, Vicksburg, MS.

The Long-Term Effects of Dredging Operations (LEDO) program by the Dredging Research Division of the USACE, Vicksburg, MS, “focus on cost-effective, environmentally responsible techniques for dredging and dredged material disposal in aquatic, wetland, and upland environments. Current research emphasizes risk-based procedures for effects assessment, exposure assessment, and risk characterization. The Program objective is to provide the latest proven technologies for identifying, quantifying, and managing contaminated sediments in support of cost-effective, environmentally responsible navigation.”

The primary benefit is a more timely complete, and cost-effective execution of the Corps’ responsibilities under the Clean Water Act (CWA), Marine Protection, Research, and Sanctuaries Act (MPRSA), Resource Conservation and Recovery Act (RCRA), and Water Resources Development Act (WRDA). This R&D program provides dramatic cost avoidances annually for achieving restoration goals for contaminated dredged material disposal as mandated by law.

Recent major products include:

1. A chronic/sublethal and genotoxic assay that meets national regulatory requirements.
2. National guidance for confined disposal facility contaminant loss assessment procedures. The guidance document was established as Corps policy to meet regulatory mandates.
3. A new analytical screening method that measures bioavailable dioxin and similar compounds with a 90% reduction cost and time required by previous methods (co-sponsored with another research program).
4. The Environmental Residue Effects Database (ERED), the first WWW-accessible database of bioaccumulation effects to improve the accuracy and defensibility of environmental impact predictions while significantly reducing evaluation costs.

LEDO research benefits/value-added to district projects can be demonstrated on a recent example. Since the bioaccumulation interpretation database is Web-accessible, it is immediately and cost-effectively available to all district projects in need of contaminated material assessment.

Dredging and disposal problems in the New York and New Jersey Harbors are mentioned in reports by WEDA (Mohan et al. 1999; Wakeman 1999). The World Bank 1990, has also issued its guidelines.

Conclusion

1. A strong development of the Dredging Industry is taking place with still larger and more efficient equipment. This development is most predominant in the low countries in Europe which virtually have taken over the world market.
2. Major projects on improvements of navigation channels and harbors remain. Artificial nourishment of beaches is still increasing.
3. Disposal areas for dredged materials are often running short due to environmental concerns. Diked areas, onshore and offshore, are becoming practices.
4. Environmental dredging is an important factor which requires much attention, planning, and monitoring. Beneficial usages of dredged polluted materials are developing.

Cross-References

- [Beach Nourishment](#)
- [Capping of Contaminated Coastal Areas](#)
- [Engineering Applications of Coastal Geomorphology](#)
- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Mining of Coastal Materials](#)
- [Reclamation](#)
- [Shoreline Response to Littoral Drift Barriers](#)
- [Water Quality](#)

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Drift and Swash Alignments

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Definition

Drift alignments are found on beach-fringed coasts where the dominant waves arrive obliquely to the shore and (with accompanying currents) maintain a beach parallel to the direction of the resulting longshore drift. They are typically found on straight or saliented coasts where the obliquely-arriving waves move sediment alongshore. Swash alignments develop where beaches have been shaped by waves arriving parallel to the shore, usually in curved patterns resulting from wave refraction. They are typically found in embayments where longshore drifting is limited and beach outlines run parallel to the crests of incoming waves.

Drift and Swash Alignments

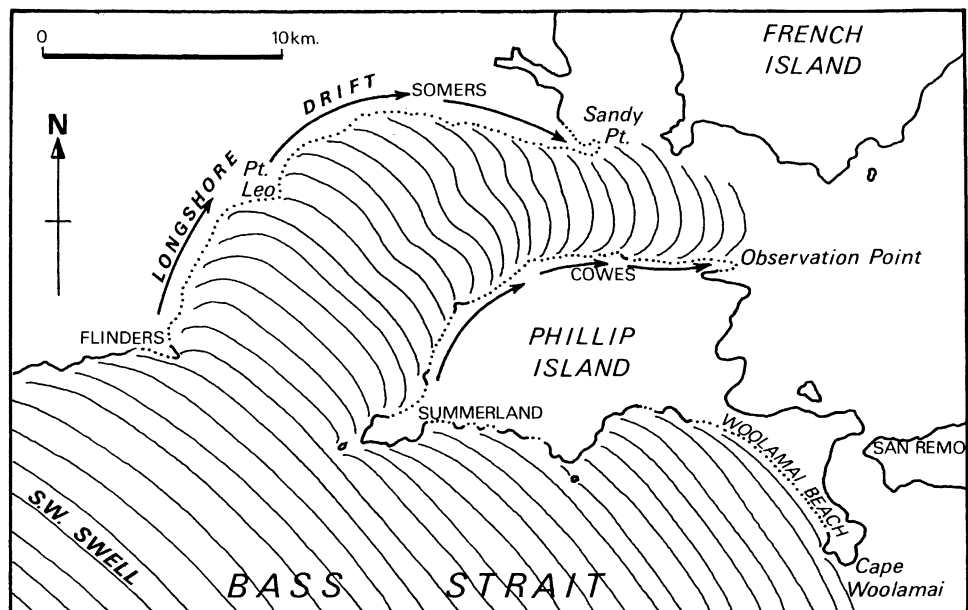
Drift and wash beach alignments are seen where an incoming south-westerly swell has produced beaches with drift alignments in Western Port Bay, Australia, and beaches with swash alignments on adjacent Phillip Island (Fig. 1). Drift and swash alignments can be developed experimentally on beaches in a wave tank (Davies 1980, Fig. 90): they are sometimes described as drift-dominated and swash-dominated beaches.

Drift Alignments

Drift alignments are best developed where waves arrive at an angle of 40–50° to the coastline. They are typically sinuous in detail, with intermittent lobes and cusps that migrate down-drift (with erosion on one side compensated by accretion on the other), and longshore spits and bars that diverge slightly alongshore. Tidal inlets also migrate alongshore, are deflected down-drift, and may be sealed off occasionally by drifting beach sediment. Variations in transverse profile occur as these features pass along the beach. A cyclic equilibrium may be attained where the beach alignment becomes adjusted in such a way that oblique waves impinging on the shore provide sufficient energy to transport the sediment arriving at one end of a drift-dominated beach through to the other, the configuration being maintained. Most beaches on drift alignments have an entirely updrift source of sediment (eroding cliffs or a fluvial input), but some incorporate sediment

Drift and Swash Alignments,

Fig. 1 South-westerly swell entering Western Port Bay, Victoria, Australia, produces drift alignments on beaches to Sandy Point and along the north coast of Phillip Island, and swash alignments in bays on the south coast of Phillip Island, notably at Woolamai Beach (Copyright-Geostudies)



derived from eroding cliffs or river mouths along the shore, as on the north coast of Hawke Bay, New Zealand (Bird 1996).

If the angle of wave incidence becomes greater or less than the optimum $40\text{--}50^\circ$ there is a slackening of longshore drifting and consequent deposition of sediment, but the drift alignment is soon restored when the oblique waves return to the optimum angle. A beach alignment can also be changed by variations in the rate of updrift sediment supply, and by erosion or accretion on updrift sectors of the coast.

Interruptions of longshore drifting result in changes along a drift-dominated beach. Projecting breakwaters designed to protect and maintain a harbor entrance show accretion on the updrift side and erosion downdrift: a well-known example is seen at South Lake Worth in Florida. The problem is often countered by introducing by-passing, whereby beach sediment is pumped through pipes from the accreting shore to the eroding shore downdrift, artificially maintaining the longshore drift.

Some drift-aligned beaches terminate in spits, such as the Langue de Barbarie on the Atlantic coast of Senegal in West Africa, a longshore spit that has grown southward in response to the dominant north-westerly swell to end in recurves at the mouth of the Senegal River. Cuspate forelands and recurved (angled) spits may include both drift-aligned and swash-aligned beaches, and as sediment eroded from the drift-aligned shore is carried round the point to be deposited on the swash-aligned shore they migrate downdrift. Dungeness, on the south-east coast of England, is a major shingle foreland that has migrated in this way. Such structures were termed traveling forelands by Escoffier (1954).

Alternatively, beaches on drift alignments may end in sectors of accretion alongside headlands, river mouths, lagoon outlets, or protruding breakwaters.

Swash Alignments

Beaches on swash alignments are smoother in outline than those on drift alignments, but are subject to alternations of profile erosion and losses of sediment offshore during storms and profile restoration by onshore sediment flow during calmer phases, a sequence known as “cut-and-fill.”

Swash alignments are well developed on beaches in embayments where intervening headlands exclude waves from all but one direction. The curving beach behind Disaster Bay, in southern New South Wales, a narrow embayment between high parallel sandstone cliffs, has a swash alignment with an outline shaped by incoming refracted ocean swell, but too limited in length for oblique waves to generate much longshore drifting. Numerous parallel dune ridges indicate that progradation has maintained this curved plan.

Swash alignments are also well developed on beaches on open coasts where the incoming waves arrive consistently

from one direction. In these situations, there is typically a simple relationship between beaches with curving swash alignments and the pattern of incoming refracted waves. In southern Australia there are long, gently curving beaches which receive south-westerly ocean swell from the Southern Ocean, refracted by nearshore shallowing: Encounter Bay is a good example. South-easterly swell entering Jervis Bay on the east coast of Australia is refracted into wave patterns that fit the curving swash-dominated outlines of bordering bay beaches that face in various directions, and a similar pattern has been documented in Storm Bay, Tasmania (Davies 1980). On the New South Wales coast a series of long asymmetrical (“zeta-curved” from their resemblance to the Greek letter ζ) beaches has developed on swash alignments, where incoming southeasterly waves are refracted round intervening headlands. Swash alignments occur on beaches on the Atlantic coasts of Europe, as on the coast of South Uist in the Hebrides, Rhossili Bay in Wales, and the Bay of Audierne in France.

Once established, gently curving beaches on swash alignments maintain their outlines through cycles of cut-and-fill when beach material is withdrawn seaward from them by storm waves, then swept back onshore in by a gentler swell. Any temporary divergence between incoming wave crests and the coastline initiates local longshore drifting which quickly restores the swash alignment.

Maintenance of harbor entrances on swash-dominated beaches by constructing breakwaters results in accretion on both sides, as sand is swept in from the sea floor, and a looped sand bar forms off the entrance, which may impede navigation. This has occurred at Lakes Entrance, Victoria, Australia, where a harbor was established in 1889 by cutting an entrance through a sand barrier. Navigation is here maintained by continual dredging of the sand bar (Bird and Lennon 1989).

Intermediate Cases

Few beaches are entirely swash- or drift-dominated. Chesil Beach in Dorset is a shingle beach which has a swash alignment shaped largely by south-westerly swell and storm waves, but longshore drifting occurs when westerly waves move sediment south-eastward and when south-easterly waves move it back westward. This alternation has led to lateral grading of this beach from granules at the western end to pebbles and cobbles toward the south-eastern end, near Portland Bill (Bird 1996).

On the other hand, beaches on drift alignments such as Sandy Hook in New Jersey or the beaches north of the Columbia River in Washington State are occasionally modified by waves coming in parallel to the shore, and sometimes show alternations of cut-and-fill with onshore-offshore (swash-backwash) sediment movement. The existence of

parallel beach and dune ridges formed by progradation behind beaches that appear to be drift-dominated, as on the North Beach Peninsula in Washington and Orford Ness in East Anglia, indicates occasional significant swash domination.

Changes in Beach Alignments

A lateral transition from a drift alignment to a swash alignment occurs on drift-dominated beaches that curve into sectors of accretion alongside headlands or breakwaters.

Where a beach outline is more sharply curved than the approaching waves it is drift-dominated, with a convergence of longshore drifting of beach sediment toward the center of the bay, where the beach progrades until it fits the outline of the arriving swell and so becomes swash-aligned. Convergent longshore drifting in Byobugaura Bay, near Tokyo, has prograded the central sector at Katakai in this way, with the arrival of sediment derived from the erosion of the northern and southern parts, and a similar evolution from drift alignments to a swash alignment has taken place at Giulianova on the east coast of Italy.

Where the beach outline is less sharply curved than the approaching waves, there is divergent longshore drifting from the center until the beach outline fits the wave pattern, assuming a swash alignment. This has occurred on the shores of the Andalusian Bight, in southern Spain, where erosion of the central sector has been balanced by progradation at Matalascanas to the south and Mazagon to the north.

The beach on the northern flank of Rheban Spit, in southeastern Tasmania, was built and prograded on a drift alignment as long as it received a longshore sediment supply from the north, but when this diminished the coastline was reshaped by updrift erosion and downdrift accretion until it developed a swash alignment adjusted to waves approaching from the north-east.

On some coasts, paired or opposed spits bordering embayments have grown on drift alignments, as at the southern entrance to Menai Strait and in Portmadoc Bay in North Wales. Convergent growth of such spits can lead to the development of a curving beach that becomes swash aligned. Beach ridge patterns on the Swina barrier in Poland indicate such an evolution on the coast of the Pomeranian Gulf.

A change from a swash alignment to a drift alignment may occur where incoming wave patterns are modified by an artificial structure such as a breakwater, or by shoal deposition or the growth of a coral reef. At Kunduchi in Tanzania, the sandy coastline had prograded on a swash alignment under the dominance of easterly swell, but the growth of reefs and shoals reduced the effectiveness of these ocean waves, and allowed locally generated and previously subdominant southeasterly waves to supervene, resulting in erosion until the beach assumed a drift alignment.

Summary

Beach-fringed coasts develop alignments related to the pattern of incident waves. Where wave crests arrive at an angle to the coastline they generate longshore drift, and the beach has a drift-dominated alignment, where wave crests arrive parallel to a coastline they anticipate and shape beach outlines, and the beach develops a swash-dominated alignment on which longshore drifting comes to an end.

Cross-References

- ▶ [Barrier](#)
- ▶ [Barrier Island Landforms](#)
- ▶ [Bay Beaches](#)
- ▶ [Cross-Shore Sediment Transport](#)
- ▶ [Cuspate Forelands](#)
- ▶ [Longshore Sediment Transport](#)
- ▶ [Spits](#)

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Driftwood

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One of the pleasures of beachcombing is viewing and collecting driftwood of innumerable shapes and sizes which have washed up on a beach. Driftwood is the collective term for all types of freely floating wood on a water body or lodged upon a beach. Driftwood can be found on most marine shores of the world, but is generally in highest concentrations along the shores of the middle and higher latitudes in the boreal forest and arctic regions of the world.

Trees along forested coastlines undermined by wave erosion are a source of driftwood and drift logs, but streams and rivers emptying into the sea are the primary contributors.

Limbs, root masses, and whole trees are eroded from the stream banks, or are products of logging practices or log storage and rafting. Trees and logs enter river channels to be transported down river during high flows or are lost within estuaries during log storage and log rafting. The drifting wood can become temporarily lodged in estuaries or flushed directly into the sea. One estimate is that over 30 million board meters of logs enters the sea annually from coastal logging operations from Alaska to California (Sedell and Duval 1985).

Driftwood Movements

The fate of driftwood discharged into the sea is determined by coastal winds, waves, and currents. Some driftwood can transit entire oceans ultimately washing up on some foreign shore. The majority of the wood afloat is more likely to be carried by nearshore waves and currents. Along the Pacific Northwest of the United States, nearshore waves and currents seasonally reverse. In the winter, southwesterly winds drive waves and nearshore currents to the north. Driftwood floating in these currents tend to end up on beaches far to the north from their river or estuary of origin. During the summer, northwesterly waves and currents carry driftwood southward along the Pacific coast. The volume of driftwood movement during the summer months is usually less as river flows are lower carrying fewer logs to the sea and waves are less energetic, dislodging less wood from beaches.

Large concentrations of driftwood are found along arctic beaches of Siberia, Canada, and Greenland. Rivers of northern Siberia and Canada's Mackenzie River carry driftwood into the Arctic Ocean during the summer. The wood is conveyed by the clockwise Pacific Gyral between the pole and Arctic Canada and the Transpolar Drift Stream from the New Siberian Islands toward the Svalbard and east coast of Greenland (Haggbloom 1982).

Driftwood on Beaches

Driftwood and floating logs are driven ashore by waves and rising tides. Smaller pieces of wood can be thrown high on a beach by waves. Large logs, 20–30 m long and 1–2 m in diameter are shoved up the beach face by the repetitive thrusting of oncoming waves. Strong storm-driven waves cram logs and driftwood into dense piles high on the beach face and backshore. Driftwood deposits differ in quantity along any given stretch of shore from a few scattered logs and smaller pieces of wood to huge matted piles up to 3 m high and tens of meters wide. The largest concentrations of driftwood are trapped in coastal embayments such as river mouths and pocket beaches. But large deposits can also be found on the backshore and berms of wide sand beaches and steep rocky high-energy beaches. The wood is driven high up

on the shore during storm conditions. There the wood may ultimately become buried by beach sediment. But for much of the wood, the beach is a brief stop off until it is again dislodged and refloated by the next series of storm-driven waves. The unrelenting pounding by waves and rolling in the surf chips away at the bark, branches, and root masses sculpting driftwood into smooth and artistic shapes.

Ecology of Driftwood

Driftwood is capable of remaining afloat for long periods of time. Large and small pieces attract growths of epifauna and flora as well as selected insects. It is well-known that both commercial and recreational fish are attracted to floating wood at sea. Tunas and dolphin-fish are routinely found under or near masses of driftwood (Gooding and Magnuson 1967). The reason is not clear, but it is thought that these fish are simply seeking shade or smaller fish attracted to the driftwood upon which they can prey (Maser et al. 1988).

Once lodged on beaches, accumulations of driftwood tend to trap wind-blown beach sand and coarser swash carried sediments. They create an organic fabric that helps beaches grow or accrete. This is probably best seen near river mouths where river-borne sediments and wood accumulate along the seashore forming sand spits mantled with driftwood and logs. If there is a sufficient volume of sediment delivered to the shore, driftwood helps to capture wind-blown sediment. The wood not only traps sediment but provides decaying organic matter for pioneering plant and dune grass species to take root. The buried wood also creates an internal structural fabric within the building beach or sand spit. It is not uncommon to see large drift logs buried many years ago protruding from the face of an eroding beach scarp. Carbon-14 dating of uncovered driftwood is useful in determining the ages of beach deposits and other environmental conditions. Thus accumulations of driftwood along beaches are thought to help stabilize and contribute to beach accretion processes. Conversely, accumulations of driftwood at the base of beach cliffs can become powerful forces of erosion by battering and abrading cliffs especially when driven by storm waves.

Numerous species of birds are attracted to accumulations of driftwood on beaches and within estuaries. Bald eagles, herons, and pelicans use the logs as perches to rest upon and capture nearby aquatic animals in the surf or mudflats of estuaries.

Driftwood Hazards

Driftwood clearly has some ecological benefits, but it also presents some liabilities especially for boaters and beachfront home owners. Floating driftwood, especially large logs, create a navigation hazard for ships and smaller pleasure boats.

Driftwood is found in highest concentrations near the shore, in bays, and estuaries generally in same location as ships and boats. Ships' hulls are usually strong-enough to withstand the impact of a floating logs, but damage can occur to rudders and propellers. Smaller pleasure boats traveling at high rates of speed are especially vulnerable to impact damage. Large pieces of driftwood can crack and even pierce hulls as well as inflict serious damage to rudders and propellers. In some waterways around the United States, driftwood is "tagged" with small flags for better visibility by boaters or hauled out of the water.

Shorefront property owners who build their homes closer to the shore than is wise are not only exposing themselves to wave erosion and flooding, but to the intrusion of their homes by wave-driven logs. Such incidents are relatively infrequent, but there are documented cases of large drift logs thrown into the living rooms of shorefront homes in Washington State's Puget Sound.

Beachcombers are attracted to the aesthetic beauty of beaches laden with driftwood. Unfortunately, the wood that is so attractive can also be hazardous. It is not uncommon to learn of a beachcomber who has been killed or injured by a drifting log thrown into the unknowing beach stroller. The State of Washington has posted signs along the shore warning the public of large driftwood.

Driftwood Management

Driftwood has long been harvested for firewood and even for minor building projects. For years the practice of harvesting driftwood from beaches has been accepted. However, some have questioned the advisability of continuing such practices, especially in light of the better understood ecological role driftwood plays in estuaries and on beaches. Recognizing the benefits of driftwood to natural ecological systems, the State of Oregon has adopted a policy prohibiting the removal of drift logs. This policy does not preclude the beachcomber from collecting pieces of driftwood, but it is intended to prohibit the large-scale removal of large logs for firewood or other uses. The State of Washington licenses individuals who wish to retrieve drift logs afloat or on beaches. Beyond licensing, Washington does not appear to have any driftwood removal policy applied to the general public.

Cross-References

- [Beach Erosion](#)
- [Beach Features](#)
- [Human Impact on Coasts](#)
- [Natural Hazards](#)
- [Marine Debris-Onshore, Offshore, and Seafloor Litter](#)

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Dune Ridges

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Dune ridges are known variously as “beach ridges” (Goldsmith 1985; Komar 1976), “parallel dunes” (Bird 1972) or “low foredunes” (Davies 1977). They are a product of a prograding coast with plentiful sediment supply, low tidal range, shallow nearshore slopes, and relatively low-energy refracted swell wave conditions (Fig. 1). Continuing progradation is associated with plentiful diathetic onshore sweep of beach sediment, alternating with episodes of cut and fill, leading to the formation of a series of beach ridges (Davies 1957). Each ridge is thought to represent the previous location of a beach berm, upon which over time a low foredune develops from “aeolian capping” (Bird 1972). The height and spacing of the dune ridges is also a function of the effectiveness of the dune colonizing vegetation in binding the sand.

Geomorphically dune ridges may be as low as 0.5 m in elevation above mean high water springs, but more typically 4–8 m high. In width they vary from 50 to 200 m, and individual crests may extend shore-parallel for many kilometers (Fig. 1). The ridge alignment parallel to the shore (hence their name “parallel dunes”) is suggestive of the swell wave environment breaker alignment. The ridges are normally composed of sand but may also contain gravel. The parallel dune fields are comprised typically of Holocene sands and may be some kilometers wide. For example, Komar (1998) shows an illustration of the dune ridge field at Nayarit, Mexico, which contain some 250 individual ridges which could have evolved in as short a time as 12–20 years, although Pethick (1984) suggests that time intervals between dune ridges more typically varies from 70 to 200 years.

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Dune Ridges, Fig. 1 Parallel dune ridges and swales on the coast near the Nassau River mouth, Queensland. (Photo: J. Mabbutt)

Separating the dune ridges are low swales characterized by impeded drainage and sometimes containing swampy vegetation. The variation in drainage characteristics leads to a tendency for variation in soil development on the well-drained ridge where oxidizing conditions prevail compared with the impeded drainage swale where reducing conditions and the podsolization process lead to dune podsol soils.

Dune ridges are widely associated with Holocene sandy barrier systems. They occur particularly along the Atlantic and Gulf coasts, Mexico, and southeast Australia, and New Zealand. Stratigraphically the dune ridges consist of aeolian low-angle cross-bedding in fine sands overlying low-angle dipping laminae of the beach face and berm, which overly marine deposits of the shoreface. These dune ridge systems evidently formed relatively rapidly (within ~1,000–2,000 years) after the sea reached its approximate present level from the postglacial transgression, presumably as a result of an abundance of sand being swept up with the postglacial transgression.

Cross-References

- [Beach Features](#)
- [Beach Ridges](#)
- [Cheniers](#)
- [Cross-Shore Sediment Transport](#)
- [Drift and Swash Alignments](#)
- [Eolianite](#)
- [Eolian Processes](#)
- [Sandy Coasts](#)

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Dynamic Equilibrium of Beaches

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Dynamic equilibrium of beaches describes the tendency for beach geometry to fluctuate about an equilibrium which also changes with time, but much more slowly. Beaches, as discussed here, refers to the visible beach including the shore and its underwater extension to a depth that is nearly

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static over the long-term. Beaches respond on a wide range of spatial and temporal scales to natural and anthropogenic forcing. Natural agents include waves, tides, currents, winds, and other elements whereas anthropogenic agents include interruption of sediment supply and induced subsidence through withdrawal of ground fluids including water and hydrocarbons. Changes of beach position can occur with or without corresponding changes in the beach profile volume. Over a long time period, coastline position can be considered as the superposition of a long-term trend about which substantial fluctuations occur. Some of these fluctuations are quasi-periodic and others appear random. The paragraphs below discuss the coastline trends followed by the fluctuations.

Long-Term Changes

Worldwide, the most dominant cause of long-term coastline change is sea-level rise. The global rate of sea-level rise is approximately 12 cm/century, although there is concern that global climate warming may increase this rate substantially in the future through melting of ice and thermal expansion of oceanic waters. The concept of an equilibrium beach profile is useful in considering the effect of sea-level rise on coastline position (Dean 1991). This has been formalized by Bruun (1962) as the so-called “Bruun Rule.” A sea-level rise will cause the profile to be out of equilibrium since the depth at a distance from the coastline is now greater than before and equilibrium can only be restored through a decrease in depth which, in the absence of external sediment addition, must be provided by volume eroded by coastline retreat. The Bruun Rule is a simple equation which depends on a number of factors including wave climate, and sediment characteristics and is expressed as the coastline retreat being a multiple of the sea-level rise, with the multiple usually ranging from 50 to 100. Thus in one century, the expected average worldwide coastline retreat would range from 6 to 12 m; however, the coastline trend at a particular location can deviate from this value by orders of magnitudes ranging from coastline retreat to coastline advancement. In addition, the Bruun Rule can be applied to *relative* sea-level rise resulting from natural or anthropogenically induced subsidence. An example is near the Mississippi River outlet where the relative sea-level rise rate is on the order of 10 times the worldwide rate noted earlier. Entire islands vanish in periods of decades.

Short-Term Changes

The most rapid short-term changes in coastline position occur during severe storms when the beach may retreat on the order of 50 m in several hours. The primary elements are the

elevated water levels (storm tides) and high waves which can cause significant modifications to the nearshore geometry including shore and dune retreat. As discussed for sea-level rise, the storm tide causes the profile to be in temporary disequilibrium and the waves provide the energy to accomplish the profile adjustment. Averaged over long sectors, the primary sediment transport during storms is seaward where a new deposit called an “offshore bar” may be formed or additional sand stored in an existing bar which may be displaced farther seaward during the storm. Elevated water levels and accompanying large wave heights can cause large coastline changes without corresponding changes in the sand volume in the profile. An example illustrating this is a wave tank in which the sand volume is constant; however, the sand can shift offshore resulting in coastline retreat. Coastline retreat associated with a storm and for which there is no volume change will generally be followed by coastline advancement and complete or near complete recovery of the pre-storm coastline position. The timescales of the recession and recovery (advancement) phases differ markedly with the recession occurring in hours to days and the recovery requiring months to years. The modes of profile recovery also differ. The profile changes during erosion as the sediment is transported seaward tend to occur with the profile “hinging” about a subaqueous point with the elevations decreasing landward of the hinge point and increasing seaward of the hinge point. The recovery phase is much more complex with the landward transport of sand occurring in individual sand “packets,” each of which, upon reaching the shore, advances the coastline seaward. During long periods of milder wave conditions following a major storm, a number of recovery packets may occur in order to complete the recovery phase. The interest in anticipating and designing for storm-induced beach and dune erosion has led to the development of numerical models (Kriebel and Dean 1985; Larson and Kraus 1989). These models are much more effective in representing the erosional phase than the more complex recovery phase. Seasonal coastline changes, described as the average annual coastline fluctuations are poorly understood and vary substantially for various coastlines. The seasonal coastline change range in Florida is approximately 10 m whereas along the south shore of Long Island, NY, the range is about 30 m. The range in southern California is relatively large and is evident by the presence of sandy beaches during summer and rocky beaches left as a lag deposit in the winter when the sand removed has been stored in the offshore bar. Studies have documented that these seasonal coastline changes are related to the wave characteristics with the seaward transport associated with higher waves of shorter period which generally occur in the winter and the landward transport occurring during the milder summer conditions. Some beaches experience somewhat organized beach fluctuations which vary substantially along the shoreline. These appear as “sand waves”

(Bruun 1954; Bakker 1968; Verhagen 1989; Thevenot and Kraus 1995) which propagate along the shore and may have amplitudes of tens of meters sometimes causing distress to property owners with homes located near the high-tide shoreline. The understanding of the causes and relative longevity of sand waves is poorly understood. Proximity to structures (natural or constructed) and/or inlets can increase the coastline fluctuations significantly relative to those along a long unobstructed coastline. The amplified fluctuations near structures is a result of changes in wave direction which causes accumulation of sand against and removal of sand from the beach adjacent to the structure. Similarly, high-tide shorelines in pocket beaches change their orientation in response to changes in wave direction causing large fluctuations near the ends of the beaches (Thompson 1987). The changes near inlets appears to be primarily due to the influence of and the interaction with the deposit of sand, termed the “ebb tidal shoal.” This feature can shelter the adjacent shores from waves causing accumulation, can shift in position thereby exposing the shore to increased wave energy and can release large amounts of sand to the shore in packets. Dredging of inlets can also cause large fluctuations (usually erosional) along the adjacent coastlines. Studies of the long-term coastline position data base in Florida has shown that the standard deviations around the coastline change trend line are approximately 10 m along long coastlines and are on the order of 50 m near natural inlets (Dean et al. 1998).

Summary

In summary, beaches are dynamic, generally exhibiting a trend of coastline change which can be either advancement or retreat, with fluctuations about this trend. Although the underlying processes governing beach change are well-understood qualitatively, our ability to predict quantitatively these changes is likely to remain poor for the coming decades.

Cross-References

- [Beach Erosion](#)
- [Beach Processes](#)
- [Beach Profile](#)
- [Changing Sea Levels](#)
- [Coastal Changes, Gradual](#)
- [Coastal Changes, Rapid](#)
- [Coastal Subsidence](#)
- [Depth of Closure on Sandy Coasts](#)
- [Storm Surge](#)

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Economics of Beaches

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Definition

Beaches are economic as well as natural resources. As natural resources, they add beauty to the coast and provide habitats for many creatures including birds and sea turtles. As economic resources, they provide services to people and properties that have an economic value. They also generate impacts on the economy and tax base.

Economic Services Provided by Beaches

An important service provided by the beach at the coast is reduced storm damage to upland properties (US Army Corps of Engineers 1996, Chap. 5). Beaches reduce storm damage by moving the waterline further from upland property. During storm events, water travels less far inland as a result of the beach, and so damage to upland property tends to be less. Of course, a beach does not eliminate storm damage to upland properties, and in severe storms it may provide little or no protection. For example, the coastline impacted by the center of a hurricane may receive little protection from its beaches, but the upland properties impacted by the outskirts of the hurricane may receive very considerable storm damage reduction from their adjacent beaches. The second major service provided by a beach at the coastline consists of the opportunity for recreation provided to beach users. Although there is no direct fee charged in most places for recreational access to the beach, such recreational services have economic value just like playing a round of golf or a visit to the movies.

Beaches also generate impacts away from the coastline. Chief among these are the impacts on the local economy of the expenditures made by tourists attracted to the beach as a recreational resource (*tourism and coastal development, q.v.*) (Schofield 1986, Chap. 14). There will also be benefits at the regional (e.g., state or province) level to the extent that tourists are attracted from outside the region and benefits at the national level from the spending by international visitors. The expenditures of beach tourists increase sales and production by impacted business resulting in increased business profits and increased earnings of employees. The well-known multiplier process, popularly called the “ripple effect,” will cause these economic benefits to be passed in a chain from frontline businesses to their suppliers and from employees of those businesses to businesses which provide goods and services to those employees. Of course, such positive economic impacts vary depending on the size of the economy, which is impacted. In small economies, many of these impacts are lost to imports. There may also be disbenefits from beaches. For example, the tourists attracted to the beaches may contribute to traffic congestion. The congestion in turn increases travel time in the local area, and the involuntary consumption of this time represents an economic loss to the travelers, which can be measured by their wage rates. Additionally, the congestion may increase the frequency of automobile accidents with associated property losses.

Government tax revenues will also be expanded as a result of beaches (Stronge 1998). The particular impacts, of course, will depend on the particular tax and fee structure in operation in the beach area. In many local communities in the United States, local communities obtain revenues by taxing property values. Beachfront property values tend to be higher than property values away from the beach partly because land use tends to be more intensive at the beach. That is, there are more hotels and high-rise apartment buildings at the beach. But, even beachfront single-family homes tend to have high values relative to homes elsewhere in the community. Other things being equal, land values on the beachfront

tend to be high relative to values elsewhere in the community because properties on the beach provide easier access to the recreational and amenity values of the beach. Structural values on properties tend to be positively correlated with land values, and so the high beachfront land values stimulate the construction of relatively expensive beachfront structures. It is not unusual to see small, inexpensive beachfront structures that are in good condition being demolished and replaced by larger, more expensive structures, because sharply increased demand for beachfront land has raised land values to levels where beachfront structures are out of equilibrium with underlying land values. The expenditures of tourists are usually subject to sales taxes, including accommodation taxes, general sales taxes, and taxes on gasoline. These revenue sources are favorably impacted by beach tourism. Finally, income taxes will tend to rise as a result of the increased earnings of workers in the tourist industry. Offsetting these positive impacts on governments may be the required increases in government expenditures resulting from beach tourism. Increased traffic congestion may result in increased expenditures for law enforcement personnel, for example. There will be increased expenditures to clean trash off the beaches and, perhaps, for lifeguards. The fiscal impact of a beach, therefore, is the net impact of the positive increases in government revenues less the required increases in the government expenditures.

Government Policies on the Beach

Governments regulate the activities of beachfront property owners because actions taken by one property owner may adversely affect nearby property owners or negatively impact the society at large (Fischer 1990). Many beaches are subject to *beach erosion* (*q.v.*), steadily diminishing in size on an annual basis. Beaches are often dynamic systems losing sand during some seasons and gaining sand during others. Along an unbroken stretch of beach, with no headlands jutting into the sea, sand tends to travel along the coastline, and it may travel in one direction during one season and reverse direction in another. The direction traveled tends to reflect the weather patterns and, in particular, the direction from which storms come. On an unbroken coastline with no headlands, storms move sand around on the beach, but net losses may be low. In some cases, storms may push sand upland away from the beach, and, in some cases, sand may be lost offshore. But in many cases, storms move sand away from the coastline, and then natural processes of wave action will return most of the sand to its original position after the storm is over.

Inlets, including gaps between coastal islands and river mouths, represent holes in the coastline, which interrupt the flow of sand along the beach. When the tide is coming in, sand

is pushed into the inlet, and when the tide is ebbing, sand is deposited offshore. These movements often lead to an offset in the coastline as beaches on one side of the inlet lose more sand than beaches on the other. The asymmetric losses reflect the fact that storms from one direction may be stronger than storms from another. When new inlets are constructed, or old inlets are improved, the equilibrium relationship between adjacent beaches is disturbed, and properties on one side of the inlet may be threatened by erosion. Such activities are usually undertaken to improve navigation, but they may also have desirable environmental consequences. In many cases, a regular program of dredging shoals that develop around the inlet can provide a source of sand for adjacent beaches that are adversely affected by the inlet. These involve intergovernmental cooperation, particularly about cost sharing, that were not resolved until comparatively recently in the United States. It may also be possible to bypass sand mechanically from an accreting side of the inlet to an eroding side. This relatively automatic beach maintenance program may be cost-effective, but sand bypassing often involves the construction of an industrial-type physical plant that may be viewed as unsightly by nearby property owners.

Beachfront properties away from inlets may also be threatened by erosion if they were originally constructed too close to the high waterline. Such structures tend to be relatively old, dating from a time when beaches were regarded as static and, perhaps, when storm events had widened a beach beyond its long-run equilibrium size. Older structures may also have been constructed at a time when regulatory authorities were unaware of the dynamic properties of the beach or when enforcement of regulations was inadequate. In any case, many beaches have properties that are in danger of damage from erosion. Under these circumstances, property owners will often construct structures such as seawalls or revetments to protect their properties from storm damage. These structures may interrupt the flow of sand to adjacent properties or accentuate the force of the waves experienced by neighbors. As a result, such structures tend to stimulate the construction of similar structures along entire segments of the coastline. The actions of the first property owners have negatively impacted their neighbors, and because the structures protrude into the beach area, the actions reduce the beach available to the public for recreational purposes and to birds and other creatures, which use the beach as a habitat. To avoid such negative externalities, governments regulate activities of beachfront property owners including requiring building setbacks and requiring erosion control structures, to pass through a permitting process. In the United States, all three levels of government require permits for beachfront construction. This occurs because beach systems cross local government boundaries and, in some cases, state boundaries and because there is a national interest in protecting wildlife and other environmental resources.

Beach Nourishment

In some cases, erosion control may be accomplished by placing sand on the beach to widen it, often called *beach nourishment* (*q.v.*) (National Research Council 1995). The conditions that would make beach nourishment successful will depend on the strength of the forces causing erosion. Additionally, it must be possible to obtain a source of sand that is cheap enough to make the project economically feasible. In other cases, it may be appropriate to relocate properties further away from the high waterline. This solution is most appropriate where there is available land to accommodate the relocated structure and where the structure is relatively small and easy to move. Such a structure might be a one- or two-story second home on a relatively large lot, for example.

When beach nourishment is an appropriate remedy for erosion (i.e., it would be successful), and when it is economically feasible, it has the added advantages of retaining the recreational and environmental values of the “natural” beach. Structural remedies, on the other hand, often reduce the available recreational beach and the available habitats for birds, turtles, and other creatures that use the beach. Additionally, off-beach benefits such as economic and fiscal impacts are also retained by beach nourishment as opposed to structural solutions. It is rarely economically feasible for one property owner, or even a small group of property owners, to undertake a beach nourishment project unless they own a relatively long stretch of coastline. This is because beach nourishment projects have large fixed costs largely unrelated to the volume of sand being placed on the beach. These include costs of sand searches, permitting costs, and mobilization and demobilization of equipment charges. For example, sand searches are expensive because not only must sand be found, but it must be compatible with the sand already on the beach. An archeological survey may be required to ensure that there are no archeological artifacts (shipwrecks) in the sand source. Various environmental studies may also be required as part of the permitting process. Additionally, equipment including a dredge may have to be brought from hundreds of miles away. Once at the site, there are setup charges and charges for dismantling the setup at the end of the project. On rare occasions, it may be possible to “piggyback” a small project on a large project that is occurring nearby, thereby greatly reducing the fixed costs incurred by the small project. But beach nourishment projects are highly complex, and it is difficult to group projects together because all projects may not be ready for construction at about the same time. Even when a relatively large group of property owners is interested in undertaking a beach nourishment project, they may not be able to agree on a formula for cost sharing. Beach nourishment is a public good in the sense that it is usually not possible to prevent consumers of the services provided by the beach from benefiting from the project unless they have paid

for them. This is because beach dynamics will redistribute new sand along the entire section of the beach, and property owners in the middle of the project will gain sand even if they have not paid. This is the well-known “free rider” problem associated with public goods. As a result, beach nourishment, when feasible, may not be selected by owners of properties threatened by erosion even when appropriate and, in principle, economically feasible. But because sand placement retains the recreational and environmental values of the beach, as well as the positive economic and fiscal impacts, society as a whole may wish to encourage beach nourishment when it is an appropriate remedy. Under such circumstances, governments may wish to encourage beach nourishment as a tool for erosion control (Stronge 1995). At a minimum, government may take a leadership role, financing initial studies and educating property owners about beach nourishment. In many cases, however, governments also subsidize construction and maintenance costs. Such subsidies, of course, will be part of a regulatory regime designed to discourage the use of structures to the extent legally possible.

Many beach projects combine nourishment and structural solutions. Projects adjacent to inlets include terminal groins designed to prevent end losses from nourishment projects. The groins also preserve the navigability of the inlet. Initial nourishment projects designed to restore a beach after a period of neglect may also contain breakwaters and other structural solutions to deal with subareas that are expected to experience relatively high rates of erosion during the life of the project. Additionally, as subsequent beach maintenance projects are undertaken, structural solutions may be included to provide protection to “hot spot” areas, that is, areas where local conditions lead to more rapid erosion than in the project as a whole.

Measurement Issues

Most US beach nourishment projects receive some federal subsidies through the Army Corps of Engineers budget. In order to qualify for such subsidies, beach projects must be economically justified, that is, they must yield benefits in excess of their costs. This requires an economic study to be undertaken in advance of project construction. The economic study projects the benefits and costs of the proposed project. A methodology has been developed over many years for projecting benefits.

A benefit projection methodology is, in effect, a model of how the project’s benefits are generated. Like all models, a beach projection methodology involves a simplification of reality. In particular, the model identifies a limited number of key benefits and does not attempt to include all the benefits of the project. Additionally, there are key parameters imposed exogenously on the model, such as the length of the planning

horizon and the discount rate used to convert future dollars to present worth or value. Given the potential for inaccuracy introduced by the modeling process, it is surprising that few follow-up studies have been undertaken that would evaluate the accuracy of the model in projecting benefits.

Because benefit projections are based on a model, there are a variety of issues that have generated some debate. First, most beach projects are justified on the basis of a very limited number of benefits, namely, property protection and recreational values. At the federal level, off-site benefits such as economic and fiscal impacts are not included, although these benefits may be of the most importance at the local level. The value of environmental benefits is not usually measured, although projects can be prohibited or forced into redesign if they have significant negative environmental impacts. Finally, there may be externalities from beach projects. A beach is a highly visible piece of infrastructure, and failure to maintain it may be used by potential migrants to a local community as an indication that the quality of invisible infrastructure (e.g., water, sewer) is poor. Restoration of the beach may lead to a general rise in property values over and beyond what would be predicted by projecting direct benefits.

Storm protection benefits are measured as the discounted value of expected dollar losses assuming the project is constructed, less the discounted expected dollar losses if the project is not constructed. The losses from storms are applied to structure values based on rules of thumb, such as assuming the proportionate dollar loss is twice the proportion of the structure that is damaged by a storm. Thus, for example, when 50% of a building is lost, all of its structure value is lost. Expected losses are calculated as the sum of the probabilities of storm events times the dollar losses resulting from those events. Storm protection benefits are discounted using an exogenously given real interest rate. The basis for the original interest rate is obscure, although it is adjusted regularly using a nominal interest rate on government securities. A lower interest rate makes it easier to justify projects, and a higher interest rate makes it more difficult. Opponents of Corps of Engineers' projects favor a high interest rate, while supporters of such projects favor a low rate. Some people favor basing the interest rate on the government bond yield, but others argue that this will favor public over private projects since government borrowing costs are lower than private borrowing costs. An alternative interest rate could be based on the home equity loan market, under the assumption that taxpayers had to borrow in order to pay their taxes. Beach benefits are discounted over a 50-year period, assuming that the project is maintained on a projected schedule. The 50-year planning horizon is somewhat arbitrary, although if the discount rate is high, future benefits and costs will be reduced to zero when discount factors are applied to values generated far into the future. Follow-up studies of the accuracy of projected maintenance intervals are needed.

Recreational benefits are measured based on willingness to pay by beach users. Two approaches have been taken. One is based on travel costs borne by the beach user, that is, out-of-pocket expenses and time expended by the beach user in traveling to the beach. This approach requires a variety of assumptions to be made about travelers. An alternative approach is to use surveys of beach users to collect information on willingness to pay. It is more common to conduct surveys as personal interviews on the beach rather than by telephone or mail in order to reduce nonresponse bias. There is a large literature in economics relating to contingent valuation surveys, including recommendations for wording questions (Portney 1994). One issue relates to whether the interviewee realizes that dollars expended on a beach visit are unavailable for alternative products. A second issue relates to allowing an interviewee to select from a list of willingness to pay values, as opposed to giving a yes or no to a single value. Interviewees may have a tendency to pick a middle value on a list. Collecting information by obtaining a response to a single value, the so-called referendum method, requires a larger sample size of interviewees. In the case of beach surveys, there are also unique problems. A certain number of people on the beach strongly oppose user charges for entry to the beach and report little or no willingness to pay values in an effort to discourage the imposition of such charges. There may also be a group who say that the value of the beach is infinite, since it is a beautiful natural resource. Finally, there are difficulties with estimating willingness to pay for beach improvements by interviewing users of the unimproved beach. The number of new users and the increase in average willingness to pay may have to be projected using very limited data.

Issues Relating to Funding Beach Improvements

Many beach projects take place on stretches of coastline where the upland properties are at least partly privately owned. Moreover, the average income level of beachfront property owners tends to be high since, as noted above, beachfront property values tend to be high. As a result, benefits from beach improvements may be skewed in the direction of relatively affluent beachfront property owners. In many cases, the public perception that benefits are skewed to beachfront property owners leads to the use of specialized funding sources as a means of financing government subsidies. Thus, for example, a special taxing district may be established to collect additional property taxes from beachfront property owners. In another case, a tax on accommodations may be used, particularly if such accommodations have large clienteles of beach users. In some cases, local funding is obtained by charging beachfront property owners for their storm protection benefits and by charging non-beachfront property owners for their recreation benefits.

In the United States, there are also issues about funding by level of government. Under current law, federally approved beach nourishments may receive subsidies in amounts up to 65% of the total cost for initial projects and up to 50% of the cost of maintenance projects. There are some peculiarities in the federal program for subsidizing beach nourishments. The federal government requires that projects seeking subsidies be economically justifiable based on storm protection benefits alone, and this is most easily achieved by projects with highly intense upland land uses, such as high-rise apartment buildings and hotels. On the other hand, the size of the federal government subsidy depends on the extent of public access and parking to the improved beach, that is, the size of the recreational benefits resulting from the beach improvement.

There have been efforts by the executive branch of the federal government to reduce the size of federal subsidies for beach nourishment projects. Most of these efforts have failed to obtain support in Congress. The Clinton administration attempted to limit the number of beaches that could qualify for federal subsidies by distinguishing between beaches of national interest and beaches of state and local interest in much the same way as highways are divided between national roads, state roads, and local roads. The same administration became concerned about the large subsidies that would be required to maintain existing projects, in an era of large budget deficits, and successfully reduced maintenance subsidies from a maximum of 65% of costs to 50%. Both the Clinton and Bush administrations believe that most of the benefits from beach projects accrue at the local level, and, therefore, they argue that most of the funding should be obtained from that level. They argue that, from a national viewpoint, economic and fiscal benefits represent a redistribution across regions rather than a substantial increase in the national totals. Recreational benefits also represent a redistribution, since beach users may use other beaches or alternative recreational facilities if a beach is not improved. The Bush administration has proposed lowering the maximum federal share to 35%. On the other hand, the federal government maintains control of local beaches as habitats for wildlife, and at least some erosion has been caused by navigation improvements to federally maintained inlets. The federal government is arbitrarily adjusting cost-sharing proportions without undertaking any study of the national benefits of beaches. Until such a study is undertaken, it is unlikely that an appropriate federal cost share will be obtained.

Beach Nourishment Outside the United States

In addition to the United States, beach nourishment projects have also been undertaken in other countries. Hanson et al. (2002) contain a comprehensive survey of beach nourishment project in Europe. The paper includes details on beach

nourishments in Denmark, France, Germany, Italy, the Netherlands, Spain, and the United Kingdom. In terms of the volume of sand placed by 1999, The Netherlands and Spain (with equal amounts placed) accounted for over 60% of the total placed by the seven countries. Germany was the only other country that accounted for more than 10% of the total. The Netherlands has a sandy coastline that is subject to erosion especially along its middle region, called the Holland Coast. A European Commission-financed EuroSION report dated 2007 can be found by searching for Holland Coast at <http://copranet.projects.eucc-d.de>. Most erosion is north of the mouth of the canal that links Amsterdam to the North Sea, and it has an important summer beach tourist industry. The plentiful supply of sand has ensured that beach nourishment has become the preferred method for combatting erosion since 1990. Most of Spain's beach nourishment projects are along its Mediterranean coast. Spain was the last of the seven countries to begin nourishing its beaches (1985), according to Hanson et al. The southeast coast has a relatively warm winter that has led to a substantial year-round beach tourist industry.

Beach nourishment in the European countries has a similar history to the US case. The use of this "soft" tool began in a small way in the 1950s and 1960s, but it was substantial by the 1980s. Many projects include a combination of nourishment and structures in Europe as in the United States. Hanson et al. note that the United States places more emphasis on storm damage prevention than do European countries. But it is inappropriate to compare a continental-sized country such as the United States (especially when Alaska with its long coastline is included) to a series of relatively small European countries. The emphasis on storm damage prevention in the United States reflects the historic role of the Federal Government in coastal protection and the primary role of the States in providing parks and recreation. Funding at the state level is driven by the desire to preserve the recreational value of the beaches and the associated tourist industries. US coastal communities are beginning to confront the impact of longer-term trends such as the rising sea level. They have much to learn from the Netherlands and other European countries in dealing with long-term threats to their coastal systems.

There is less beach nourishment activity in the rest of the world, although there are projects in Australia, and Hanson et al. include activity in South Africa in their Table 2 (Castelle et al. 2009). The Australian project in the Southern Gold Coast of Queensland includes a sand bypassing project.

Summary

Beaches provide services to people and properties that have an economic value. They also generate impacts on the economy and tax base. An important service is reduced storm damage to upland properties. They also provide an

opportunity for recreation to beach users. Beaches also generate impacts away from the coastline, especially the impacts on the local economy of the expenditures made by beach tourists. The well-known economic multiplier process will cause these expenditures and will cause these economic benefits to be passed in a chain from frontline businesses to their suppliers and from employees of those businesses to businesses which provide goods and services to those employees. Government revenues will also be increased, although there may also be increased government expenditures.

Governments regulate the activities of beachfront property owners because actions taken by one property owner may adversely affect nearby property owners or negatively impact the society at large. Many beaches have properties that are in danger of damage from erosion. Under these circumstances, property owners will often construct structures such as seawalls or revetments, and these may interrupt the flow of sand to adjacent properties or accentuate the force of the waves experienced by neighbors. As a result, such structures tend to stimulate the construction of similar structures along entire segments of the coastline. This may result in the loss of the sandy beach, reducing recreational opportunities and habitat for wildlife. Governments may provide leadership in developing soft solutions, especially beach nourishment. The large fixed costs of such projects make them economically infeasible for small groups of property owners, and their nature as public goods prevents free riders from taking advantage of the benefits while refusing to contribute to the costs.

A methodology has been developed in the United States for determining when beach projects are economically justified because the value of their benefits exceeds their costs. There are many debates about the individual parts of this methodology. These include which benefits should be included and how they should be measured. There also issues about how the costs should be apportioned among the different levels of government (federal, state, and local) and, in some cases, how costs should be shared among beachfront property owners themselves.

The history of beach nourishment outside the United States has been similar to the US experience. Toward the end of the last century, beach nourishment became the preferred policy option, although the projects also included structural solutions as well.

Cross-References

- [Beach Erosion](#)
- [Beach Nourishment](#)
- [Managed Retreat](#)
- [Setbacks](#)
- [Tourism and Coastal Development](#)

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El Niño–Southern Oscillation (ENSO)

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Local Fisheries (Enfield 1988, 1989)

The term El Niño (or “the child”) was originally used by Peruvian fishermen in the nineteenth century to refer to a Christmastime warming of coastal sea surface temperature (SST), often associated with an abrupt decrease in productivity of the local fisheries. The Southern Oscillation (SO) portion of El Niño–Southern Oscillation (ENSO) describes the global-scale surface pressure oscillation documented by workers around the turn of the century and first studied in detail by Sir Gilbert Walker (for historical reviews, see Rasmusson and Carpenter 1982; various chapters in Glantz et al. 1991; Diaz and Markgraf 1992, 2000; Diaz and Kiladis 1995; Allan et al. 1996; Diaz et al. 2001). It was not until the 1960s that Jacob Bjerknes linked the two processes and began to describe the complex interplay between the ocean and atmosphere, which comprises ENSO and can lead to dramatic perturbations of the global climate system. Here, we will be concerned with the implications of climatic change on ENSO, as well as considering the possible importance of ENSO itself for low-frequency climatic variability.

The root of what is now commonly referred to as ENSO lies in the tropical Pacific, although its influences eventually

spread far beyond that ocean basin. To begin to understand the ENSO cycle and its role in climate variability, the mean oceanic and atmospheric conditions over the eastern Indian and Pacific sectors are first considered, at least in general terms. We then examine the principal climatic anomalies worldwide that result from the development of the tropical anomalies. For further reading, see Streten and Zillman (1984) and Kiladis and Diaz (1989).

Mean Conditions

The South Pacific high-pressure system is the primary atmospheric circulation feature of the eastern South Pacific Ocean. This semipermanent surface high is associated with equatorward flow along the coast of South America, and strong southeasterly trade winds over the tropical eastern Pacific. These southeasterlies cross the equator and merge with the northeasterly trades of the north Pacific subtropical high at a latitude of about 80°N, giving rise to a zone of surface convergence and heavy rainfall known as the *inter-tropical convergence zone* (ITCZ). Further west, the southeasterly trades weaken and recurve into northeasterlies in the tropical southwest Pacific, merging with southeasterlies from surface high-pressure systems moving eastward from the region of Australia. This produces another northwest–southeast oriented rainfall maximum called the south Pacific convergence zone (SPCZ; Trenberth 1976).

The distribution of rainfall in the tropical Pacific is closely related to the SST pattern, which in turn can be understood to be driven by the surface wind stress and associated ocean currents (see Pickard and Emery 1982). One result of the Earth's rotation is to cause surface water to flow to the right of the wind in the Northern Hemisphere and to the left of the wind in the Southern Hemisphere. Thus, easterly trade winds along the equator result in a divergence of water away from the equator, which leads to the “upwelling” of relatively cooler water from depth to conserve mass. Similarly, equatorward wind stress along the western coast of South America leads to divergence of water away from the coast and strong upwelling along the steep subsurface continental margin. Regions of upwelling comprise productive fisheries areas, as the subsurface waters are often rich in nutrients. This effect, along with the northward transport of cold water in the Peru Current, gives rise to the rich fisheries and relatively cold SST for its latitude along the coast of Peru and the Galapagos Islands.

A direct result of the coastal and equatorial upwelling is to increase the stability of the atmospheric boundary layer by the cooling effect of the SST. This, along with the reduced evaporation over colder water, results in the strong suppression of rainfall in these regions. As a result, the coasts of southern Peru and northern Chile are the driest deserts on earth. Anomalously dry climates are also experienced along

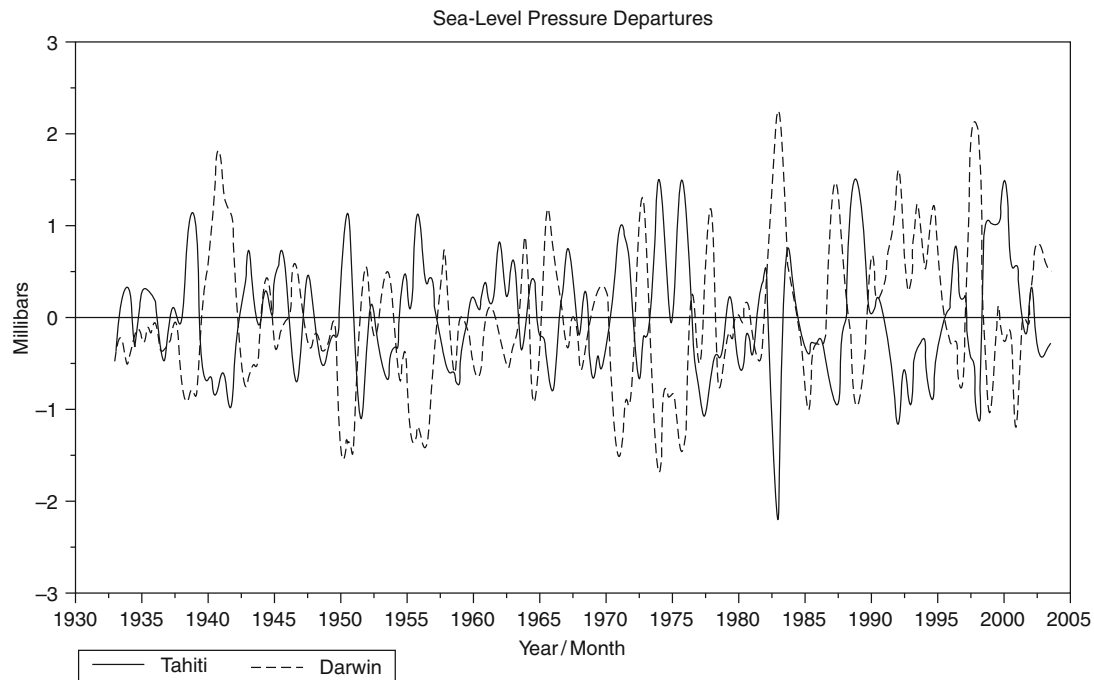
the so-called “equatorial dry zone” in the central and eastern Pacific as a result of the strong upwelling there. As one moves westward, equatorial SST increases gradually as the trades and upwelling weaken, eventually giving way to the “warm pool” west of the dateline. This region is the largest area of high SST in the world's oceans, averaging greater than 28 °C. The warm pool extends west-ward from the western Pacific through the seas surrounding Indonesia and into the eastern Indian Ocean, and is associated with high precipitation over these regions.

The ENSO Cycle

ENSO fluctuations are associated with marked deviations from the mean atmospheric and oceanic conditions described above. In the eastern Pacific Ocean, its most obvious manifestation is an increase in SST along the equator and coast of South America, with an associated increase in sea level and depth of the thermocline. This results from a decrease in upwelling due locally to a weakening of the South Pacific high and associated trade wind flow, as well as the eastward propagation of equatorially trapped large-scale waves in the ocean, which suppress the thermocline (Philander 1990).

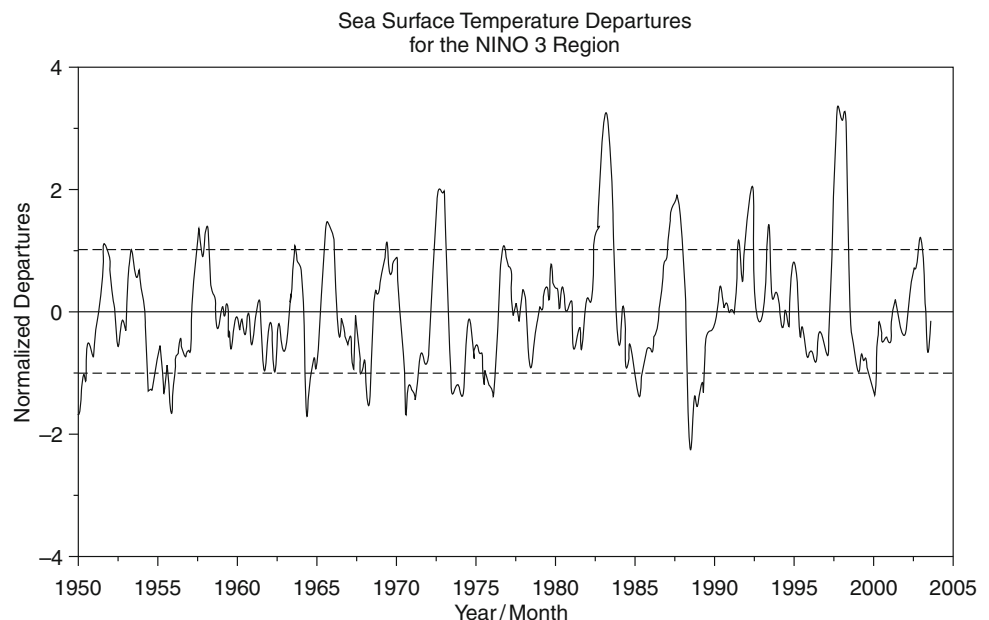
The weakening of the South Pacific high is the eastern component of Walker's Southern Oscillation and is associated with an increase in surface pressure over Australasia. Figure 1 illustrates how the surface pressure varies inversely between Tahiti (Papeete) and Darwin, Australia, on timescales of several months and greater. Various indices of the SO have been constructed utilizing normalized pressure anomalies between stations near the poles of the oscillation such as Darwin and Tahiti (see Trenberth 1984). However, the association between the SO itself and the eastern Pacific SST is not a simple one (Deser and Wallace 1987). The SO is most closely coupled to SST variability along the equator from about 160°W to the Galapagos, and only loosely related to SSTs near the dateline and in the traditional El Niño region of coastal Peru. Figure 2 gives the temporal evolution of normalized SST variations in the equatorial belt from 6°N to 6°S, 150°W to 90°W, which is known as the Niño-3 region (Rasmusson and Carpenter 1982). The correlation between the Tahiti minus Darwin sea-level pressure (SLP) difference and the Niño-3 SST is quite high (with about 64% of the variation in one series explained by variations in the other). The five most recent warm events can be seen on these plots, namely 1982–83, 1986–87, and 1991–92, 1997–98, a weak event in 2002–03, along with cold events in 1988–89 and 1998–2000.

Knowledge of the workings of the ENSO phenomenon can be obtained from an examination of its composite behavior (Rasmusson and Carpenter 1982), along with study of individual cases, which deviate markedly from this mean view (Fu et al. 1986). One interesting aspect of ENSO is the



El Niño–Southern Oscillation (ENSO), Fig. 1 Sea-level pressure anomalies at Tahiti and Darwin, Australia. Monthly departures have been smoothed with a five-month running mean, applied twice

El Niño–Southern Oscillation (ENSO), Fig. 2 Standardized SST anomalies in the El Niño-3 region for the period 1951 through July 2003. Reference period is the full period of record; monthly data have been smoothed with a three-month running mean, applied twice



tendency for phase locking to the annual cycle. ENSO extremes often first develop during the northern spring, when the trade winds are weakest and SST is highest on average over the eastern equatorial Pacific. The fundamental instability involved concerns a relationship between SST anomalies and atmospheric convection. In the majority of events, a positive equatorial SST anomaly develops initially near the dateline. Since SST in this region is on average about

28 °C, any positive anomaly in this region will create conditions conducive to anomalously active atmospheric convection (Barnett et al. 1991).

The development of ENSO events tends to peak during the late summer and early fall of the Northern Hemisphere. Notable exceptions occurred in the warm events of 1982, 1986, and 1991, which developed late in the calendar year and peaked during the northern winter and spring.

The feedback leading to a reversal of ENSO conditions is often mediated by the seasonal development of the strong trade wind regime of northern fall, as well as the depletion of heat energy in the warm pool. Much of this energy is apparently transferred to the atmosphere through anomalously high rates of evaporation, and portions of it may also be transferred to higher latitudes by the oceanic meridional circulation. Once the heat content of the ocean is depleted, the stage is set for the transition to the opposite phase of ENSO, the so-called “La Niña” or cold event (e.g., van Loon and Shea 1985; Philander 1985, 1990). Cold events also involve an unstable interaction, as described above, for warm events, except that it occurs in reverse, as cold SST in the eastern Pacific help maintain a strong westward pressure gradient and trade winds, in turn favoring the continuance of the SST pattern itself.

ENSO Teleconnections

Much of Gilbert Walker’s early work was geared towards prediction of Indian rainfall, and the SO was of great interest, as it was known that the strength of the monsoon was inversely proportional to the surface pressure over southern Asia. In his investigations, Walker (1923) established that high pressure over the Australasian region was accompanied by drought over India and Australia and cool, wet winters over the southeastern United States. Surprisingly, little work on atmospheric teleconnections was undertaken until nearly a half a century later, when Bjerknes (1966) uncovered evidence that El Niño conditions were associated with a strengthening of the storm track over the North Pacific. Since then, a wealth of work has been done on the temperature and precipitation signals associated with ENSO. Much of the synopsis that follows has been taken from the work on large-scale ENSO signals of Rasmusson and Carpenter (1982), Ropelewski and Halpert (1986, 1987), Lau and Sheu (1988), Kiladis and van Loon (1988), Kiladis and Diaz (1989), Diaz and Kiladis (1995), and Diaz et al. (2001).

Although the term “El Niño” was first used to describe the annual warming of the waters along the Peruvian coast, in the 1960s it became synonymous with the concurrence of abnormally high SST and heavy rains in the usually hyperarid Peruvian coastal plains. Bjerknes (1969) noted that anomalously heavy rains during El Niño also occurred at Pacific island stations such as Canton and Christmas in the equatorial dry zone. As mentioned above, this signal is one manifestation of the equatorward shift of the convergence zones during equatorial warming of the SST associated with ENSO. As a result, the regions normally under the influence of these convergence zones, such as the Caroline and Marshall Islands, Fiji, and New Caledonia, experience drier than normal conditions.

The most pronounced signals occur during the year that an ENSO extreme first develops (here called “Year 0”) into

the following year (Year +1), following the convention of Rasmusson and Carpenter (1982). ENSO appears closely tied to the Asian and Australian monsoon circulations, which are notably weaker during warm events (see Webster and Yang 1992). While precipitation increases in the central and equatorial Pacific during warm events, it becomes markedly drier over regions bordering the western Pacific and eastern Indian Ocean, such as Indonesia and Australia (Allan 1988; Nicholls 1992), India (Rasmusson and Carpenter 1983), and southeast Asia. Failure of monsoon rains during warm events can have catastrophic impacts (Kiladis and Sinha 1991). Conversely, cold events are often associated with flooding events in India (Parthasarathy and Pant 1985). The Australasian signal is more evident during the northern fall of year 0, and persists over the monsoon region of northern Australia into the December through February rainy season. However, during strong ENSO events, the entire continent of Australia can be affected by severe drought conditions (Nicholls 1992). See also the reviews in the special issue of the *Journal of Geophysical Research* [Anderson et al. (eds.) 1998].

It should be emphasized that, while warm events can result in devastating precipitation deficits over monsoon regions, these areas generally still receive a large amount of precipitation in all but the most extreme cases. Over certain regions of the data-sparse Indian Ocean, there is some suggestion that warm events actually see above-normal precipitation in phase with that over the central Pacific. Sri Lanka, the Seychelles Islands, and coastal stations of equatorial Africa are certainly wetter than normal during their September–November rainy season (Kiladis and Diaz 1989), and satellite data from three warm events (1982, 1986, and 1991) support the occurrence of enhanced rainfall over the southern Indian Ocean during the northern winter of those events. If so, this means that the main focus of precipitation over Australasia is actually not so much shifted eastward, but distributed more evenly across the equatorial Pacific and Indian Oceans.

As warm events evolve toward their “mature” phase during the northern winter season (DJF + 1), drier than normal conditions continue in the region of the western Pacific ITCZ from the Philippines eastward (Kiladis and Diaz 1989), and east of Australia in the normal position of the SPCZ (Kiladis and van Loon 1988). Similarly, a large region of southeastern Africa, including parts of Zimbabwe, Mozambique, and South Africa, has a marked tendency for drought during DJF + 1 of warm events. Precipitation in this region is highly seasonal, so this signal can have an especially large impact since it occurs during the normal southern summer rainy season (Nicholson and Entekhabi 1986). Farther north in Africa, there are indications that warm events favor drought conditions from the Sahel eastward to the highlands of Ethiopia during the normal summer rainy season of JJA 0 (see Janowiak 1988; Lamb and Pepler 1991).

While coastal Ecuador and northern Peru often experience flooding during warm events, other regions of the Americas often also register large rainfall anomalies. A consistent dry signal is found from JJA 0 through DJF + 1 over northern South America (Rogers 1988). Over northeast Brazil, the periodic occurrence of severe drought in the agriculturally rich Nordeste region in connection with El Niño events has resulted in severe economic hardship and occasional famines in this region (Hastenrath and Heller 1977; Chu 1991). Widespread and severe famine was reported during the great El Niño of 1877–78, while other severe El Niño episodes, including that of 1982–83, have led to great suffering. Although data are sparse, the northern Amazon basin also appears to be affected. In contrast, much of southern South America is wet during JJA 0, and this signal persists into SON 0 in Uruguay, and central Chile and Argentina (see Kiladis and Diaz 1989; Diaz and Kiladis 1995; Diaz et al. 2001). A remnant of this signal is still present in DJF + 1, when the southeastern United States and northern Mexico also shows above normal precipitation. In the heavy rainfall areas of the upper Paraná and Paraguay River Basins, MAM + 1 precipitation tends to be above normal, and this contrasts with the relatively dry summer wet season over Central America.

Although the best correlations between ENSO and temperature are observed in the tropics and subtropics, the Americas can experience large mid latitude temperature anomalies during ENSO events. Strong and relatively mild westerly flow from the Pacific into the North America during warm events is responsible for a large region of positive temperature departures from southern California northward along the west coast to Alaska, then inland across western and central Canada. This signal over northwest North America is one of the most reliable in the extratropics from event to event (Kiladis and Diaz 1989; Diaz and Kiladis 1992). Cold events are equally reliable in being associated with anomalously cold winter seasons in this region. In those years, a tendency toward a weak jet stream over the central Pacific leads to atmospheric “blocking” patterns in the Gulf of Alaska, which in turn are associated with anomalous northerly flow over northwestern North America. The increased warm-event storminess over the southeast United States discussed above is also accompanied by cooler temperatures in that region. Similar enhanced zonal flow over sub-tropical South America leads to above-normal temperatures along the west coast; as in North America, this signal is most extensive during its winter.

ENSO and Climate Variations

It is important to recognize that the ENSO teleconnections represent statistical averages over the last half-century, and the strength of these waxes and wanes on decadal timescales

for several reasons. For example, because tropical Pacific SSTs exert only a modest control on seasonal climate variations, random fluctuations of the teleconnections due to sampling alone can occur on multi-decadal timescales (Allan 2000; Diaz et al. 2001).

It has been noted elsewhere that tropical Pacific SST variations were weak during 1920–40 compared with 1980–2000 (Trenberth and Shea 1987; Allan et al. 1996; Allan 2000). Given that the strength of teleconnections increases almost linearly with the amplitude of tropical Pacific SST anomalies (e.g., Kumar and Hoerling 1998), a reasonable assumption is that stronger teleconnections will occur during decades when tropical Pacific SSTs exhibit larger interannual variance. An additional factor requires no change in ENSO temporal behavior per se, but involves slow, multi-decadal, changes in the climatological mean SSTs themselves. These can alter the atmospheric sensitivity to interannual variations in tropical Pacific SSTs because the teleconnections respond to the total, rather than to the anomalous, SSTs.

One plausible factor in this strengthening of ENSO teleconnections is a multi-decadal change in the life cycle of tropical Pacific SST anomalies during warm events (Trenberth 1984; Mantua et al. 1997). The strength of teleconnection patterns emanating from changes in tropical convection associated with ENSO variability differed considerably before and after 1976 (Diaz et al. 2001). The origin for this change is not known, and may merely reflect the stochastic behavior of the equatorial coupled system itself. Nonetheless, the relevancy for the teleconnection process is large because the warm-event life cycle since 1976 has become closely phased with the climatological annual cycle, and in particular the prolonged anomalous warming in spring of year +1 now coincides with the peak annual cycle SST warming.

In many areas where a significant ENSO modulation of precipitation is present, the strength of that teleconnection has varied over the past century. It is unclear to what degree the temporal variations of ENSO teleconnections in different regions are related. To the extent that there is a dynamical root to the changes observed in the past century, for example, due to changes in the ENSO SST forcing, then it is reasonable that changes in one region would be accompanied by changes in another.

Decadal scale patterns in tropical and north Pacific SSTs have been shown to correlate to decadal scale changes in the extra-tropical circulation, and associated temperature and precipitation patterns, particularly over North America. These changes have been statistically linked to different phases of a decadal-scale pattern of SST anomaly in the North Pacific known as the Pacific Decadal Oscillation (PDO) (Mantua et al. 1997); however, it is not clear what is cause and what is effect. One interpretation of recent modeling work on the connection between low-frequency changes

in tropical SST, ENSO, and decadal scale changes in the general atmospheric circulation suggest a complex interplay between the canonical ENSO system, the slow changes in SST in the Indo-Pacific over the last century, and long-term changes in the atmospheric circulation itself.

Cross-References

- [Climate Patterns in the Coastal Zone](#)
- [Coastal Climate](#)
- [Coastal Temperature Trends](#)
- [Coastal Upwelling and Downwelling](#)
- [Coastal Wind Effects](#)
- [Desert Coasts](#)
- [Meteorological Effects on Coasts](#)
- [Natural Hazards](#)

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Endogenic and Exogenic Factors

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Synonyms

Endogenetic; Exogenetic

Definition

Endogenic (or endogenetic) factors are agents supplying energy for actions that are located within the earth. Endogenic factors have origins located well below the earth's surface. The term is applied, for example, to volcanic origins of landforms, but it is also applied to the original chemical precipitates. Exogenic (or exogenetic) factors are agents supplying energy for actions that are located at, or near, the earth's surface or extraterrestrial. The term is commonly applied to various processes such as weathering, denudation, mass wasting, etc. In coastal science, these factors may be illustrated in two significant applications. One is the classification of coastlines and the other is the discussion of sea-level variations.

Factors in Coastline Classification

As early as 1885, Suess (as cited in Kennet 1982) described the fundamental difference between the coastlines of the Atlantic and those of the Pacific Ocean. This distinction is now the basis for classifying the world's coasts into three categories based on the endogenic processes of plate tectonics (Inman and Nordstrom 1971). Collision coasts are found at convergent plate boundaries such as along the west coasts of North and South America. They tend to be straight and

mountainous with narrow continental shelves. Marginal sea coasts, the second category, are protected from the open ocean by island arcs. An example would be the coast of Korea. Trailing-edge coasts are found on the interior areas of tectonic plates. When newly created, like the coast of the Red Sea, trailing-edge coasts appear similar to collision coasts; they are steep, lined by sea cliffs, and have a normal shelf. With age, however, they tend to have broad coastal plains, wide shelves, and low relief like the east coast of the United States.

The evolution of trailing-edge coasts corresponds to early concepts in geomorphology on the aging, or maturing, of continental margins by exogenic forces (Dietz 1952). Simply put, the continental crust is attacked by wave action and the continental margin shaped by the redistribution of marine sediment by coastal currents. Young, steep coasts are thereby replaced by the gently sloping surface of a thick wedge of coastal plain sediments. Turbidity currents and other elements of gravity-driven mass movements are then instrumental in the basic formation of the continental slope and rise. On a smaller scale, endogenic processes of faulting, volcanic, or other igneous structures provide other coastline types (Shepard 1963). These are typically erosional, rocky coastlines, referred to as erosional coasts in which the underlying geologic structure imposes the coastline pattern.

Coastal features formed by exogenic factors may be inherited or active. The surficial processes to which these coastal landforms owe their origin are no longer active. Coastlines formed by glacial action, such as fjords or drumlins, or drowned river valleys, are examples of inherited characteristics. Deltaic coasts, barrier islands, cusped forelands, tombolos, etc. are all features shaped by exogenic factors since they owe their origin to the active redistribution of sediment by waves and currents. Collectively such features are characteristics of accretional coasts, notwithstanding the fact that such coasts may have serious erosion problems.

Factors in Sea-Level Changes

Endogenic processes also control long-term eustatic sea-level changes. Deep-seated isostasy and the thermal structure of the oceanic lithosphere influence the relief of the ocean basin. Newly created seafloor stands in high relief being thermally expanded. With the passage of time, it cools, contracts, and subsides. As a result, changes in the rate of seafloor spreading and the total length of mid-ocean ridges changes the elevation of the oceanic platform, that is to say, the average depth of the ocean basin, and, consequently, the position of sea-level relative to the continents. A global drop in sea level of some 350 m at the end of the Cretaceous is attributed to the contraction of the mid-ocean ridge system, in part by the disappearance of spreading centers in the Pacific beneath North America.

Exogenic factors that control sea-level variations include seasonal cycles due to temperature variations, storm surges, or geostrophic adjustment of sea surface slopes by variations in the speed of coastal ocean currents. Anticipated sea-level changes, and other coastal impacts, due to global warming have exogenic origins. The current rate of eustatic sea-level rise is due predominantly to the combined effects of the liberation of water from the melting of glaciers and the thermal expansion of the surficial layer of the ocean.

Extraterrestrial Factors

The term “exogenic” is also applied to agents having their origins outside the earth such as solar radiation, comets, or meteorites. A collection of features that owe their origin to the impact of meteorites, such as coastal deposits of a tsunami resulting from the Cretaceous impact event, could be ascribed to this class of exogenic factors.

The Anthropocene

Although exogenic factors are usually driven by gravity or atmospheric forces, the extent of human impacts on the global environment make human activity an exogenic factor. Recognition of the global scale of human impacts on surficial processes worldwide is embodied as the “Anthropocene.”

Anthropogenetic factors of climate change, sea level rise, ocean acidification, changes in land-use, global freshwater demand, stratospheric ozone, and various biogeochemical cycles all impact the functioning of earth’s surficial systems.

Conclusion

The distinction between exogenic and endogenic factors provides the underlying direction in the search for causes of terrestrial phenomena and features. Of particular importance in the Anthropocene is the recognition of factors that human beings might manipulate, or have manipulated, to impact climate change or its consequences. While these are exogenic, for the most part, other factors are recognized as beyond the control of humankind but deserve monitoring for advanced warning of potential hazards.

Cross-References

- [Barrier Island Landforms](#)
- [Changing Sea Levels](#)
- [Faulted Coasts](#)
- [Holocene Coastal Geomorphology](#)
- [Volcanic Coasts](#)

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Energy and Sediment Budgets of the Global Coastal Zone

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The energy for the global coastal zone is primarily powered by solar irradiance and tides. The solar irradiance warms the earth’s surface unevenly, driving wind and current systems that redistribute heat and generate storms with rainfall and snow that erode landmasses. The erosion products of sediment and dissolved solids are carried to the sea by rivers and glaciers. Winds, waves, tides, and currents transport and redistribute sediment and sculpt coastal landforms.

Solar irradiance provides an energy flux of about 1.8×10^{14} kilowatt (kW) to the earth. About 2.5×10^9 kW of mechanical energy in windgenerated waves is incident on the world’s 440,000 km coastlines. Another 2.2×10^9 kW of tidal energy is expended in shallow seas. The total flux of mechanical energy of all kinds in shallow waters of the coastal zone is about 5.5×10^4 kW. Some of this energy is expended in coastal erosion and transport of the erosion products from the land, and the remainder drives frictional processes and is dissipated as heat. The erosion rate of land is about 6 cm/thousand years resulting in a flux to the sea of 5×10^9 ton/year dissolved solids and 20×10^9 ton/year particulate solids (1 ton = 1,000 kg).

Douglas L. Inman: deceased.

Introduction

The oceans respond to many different kinds of steady and impulsive forces applied at their boundaries with the atmosphere and seafloor. Although most energy in the nearshore waters comes from windgenerated waves and tidal currents, energy from a number of other sources such as internal waves and ocean currents may also be important locally.

One approach to understanding the relative importance of nearshore processes is to compare the sea's potential to erode the land with the land's potential to supply terrestrial erosion products. Such a comparison ultimately resolves itself into the balance between the budget of power in waves and currents and the budget of sediments available for transport. Of course, this balance varies widely from place to place, and even in the best-studied areas is but poorly understood. However, order of magnitude estimates can be attempted on a worldwide basis, and their consideration here gives some overall perspective for the relative importance of the driving forces that are operative in nearshore waters.

In a practical sense, the waters of the coastal zone can be considered as including the shallow seas and the waters covering the continental shelves of the world, an area of $29 \times 10^6 \text{ km}^2$, or about 5.5% of the surface area of the world and about 8% of the area of the world oceans (Table 1). The nearshore waters are bounded on the landward side by coastlines that total about 440,000 km in length, and to the seaward by the break in slope at the shelf edge (shelf break) marking the change from the relatively horizontal continental shelf to the steeper continental slope. The continental slopes are one of the striking geographic features of the earth and have a combined length of about 150,000 km.

Conventionally, the depth of the shelf break is taken as 200 m or 100 fathom, although in some localities the depth may be as great as 400 m. On a worldwide basis, the average

depth of the shelf break is about 130 m. The width of the shelf ranges from zero to over 1,300 km and averages about 74 km (Shepard 1963). Although having little geomorphic basis, international law defines *territorial seas* as usually extending seaward 12 nautical miles (22.2 km) beyond the coastline and an *exclusive economic zone* as usually extending 200 nautical miles (371 km) seaward.

Budget of Energy in Shallow Water

The current systems of the world's atmosphere and ocean are solar powered. The sun irradiates the earth with a total of about $1.8 \times 10^{14} \text{ kW}$, or an intensity of 1.37 kW/m^2 , the so-called solar constant. In terms of the rotating, spherical earth, this averages to about 342 W/m^2 of earth's surface. Of this amount about 30% is reflected and backscattered, 19% is absorbed by the atmosphere, and 51% is absorbed at the earth's surface (e.g., Gill 1982). Solar energy heats the earth's surface unevenly and generates winds. The oceans are relatively opaque to solar irradiance, so that the effect on the oceans is to stratify and stabilize the surface water. As a consequence, the ocean responds much more to wind, even though the intensity of thermal energy from the sun greatly exceeds the mechanical energy from the wind at the water's surface. Averaged over the earth's surface, the intensity of thermal energy absorbed is about 175 W/m^2 , while the sea surface stresses and pressures caused by a wind blowing 10 m/s expend about 1.3 W/m^2 . In contrast, the mean heat flow to the earth's surface from the interior is approximately 0.1 W/m^2 , and occurs mostly at the mid-ocean spreading centers (Pollack et al. 1993). Thus, it is the solar powered atmosphere and ocean currents together that maintain the earth's heat budget by accounting for a poleward transport of energy at the rate of about $8 \times 10^9 \text{ kW}$.

Studies of the energy budgets of the ocean and atmosphere (e.g., Webster 1994) show that the total energy content of the oceans ($\sim 160 \times 10^{25} \text{ joule (J)}$) is 1000 times larger than that of the atmosphere ($\sim 125 \times 10^{22} \text{ J}$). Because of greater mass and heat capacity, the oceans play a critical role in controlling and moderating weather and climate. However, in terms of the kinetic energy available for interacting with the oceans, the atmospheric kinetic energy ($\sim 6 \times 10^{20} \text{ J}$) is 200 times larger than that of the oceans.

The wind systems are the direct link between atmosphere and ocean and, through momentum exchange at the ocean's surface, expend kinetic energy at the rate of about 10^{11} kW (Malkus 1962; Newell et al. 1992). The prevailing wind systems cause all large bodies of water to have windward and leeward shores. The Pacific coasts of the Americas are in general windward coasts while the Atlantic seaboard is leeward coasts. The general wind flow is clockwise in the Northern Hemisphere around semipermanent, midlatitude areas of high pressure, for example, the north Pacific high

Energy and Sediment Budgets of the Global Coastal Zone, Table 1 Dimensions of major topographic features

Feature	Area (10^6 km^2)	Length of coastline (10^3 km)
Continents	138.8	210
Large islands (larger than $2,500 \text{ km}^2$)	9.3	136
Small islands ($25\text{--}2,500 \text{ km}^2$)	0.9	94
Total land	149.0 ^a	440
		Length of shelf break (10^3 km)
Oceans (depths $>200 \text{ m}$)	332.1	
Continental shelf ($0\text{--}200 \text{ m}$)	29.0	149
Total marine water	361.1	
Total area earth	510.0	

Source: Inman and Nordstrom (1971) for features with length scales greater than 10 km

^aUnited Nations (UNEP 1987) gives total land area of $145 \times 10^6 \text{ km}^2$

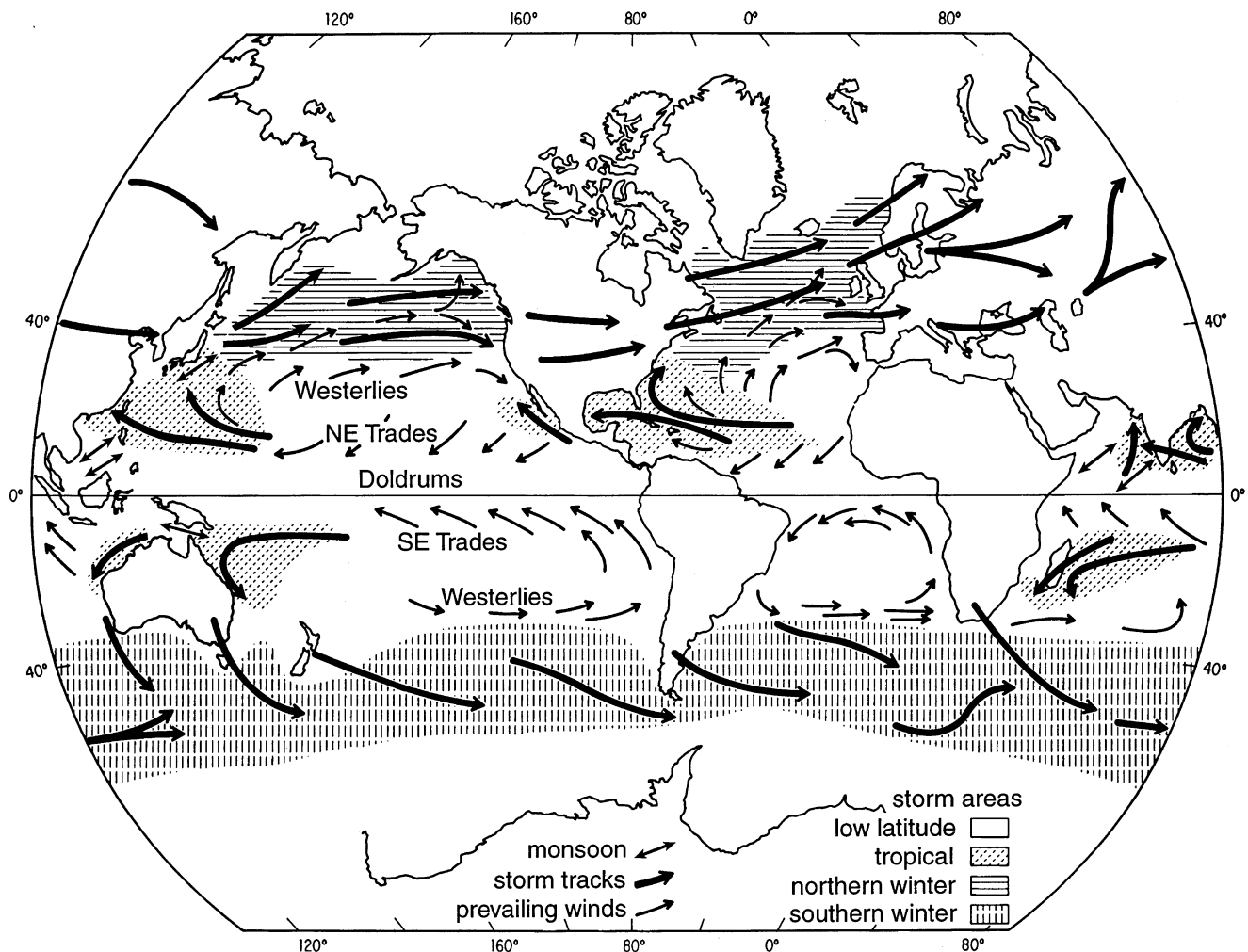
and the Bermuda high in the Atlantic Ocean. Circulation patterns in the southern oceans are essentially counterclockwise flowing mirror images of those in the Northern Hemisphere (Fig. 1).

Sea and Swell

The most intense wave action in all oceans results from cyclogenesis over the poleward-flowing, western boundary currents. These warm water jets carry equatorial heat to higher latitudes and produce strong temperature gradients that spin-up intense storms (Fig. 1). These boundary cyclonic events generate large, high-frequency *sea waves* along leeward shores and long waves that cross the ocean and appear as low-frequency *swell* on windward coasts. In the north Pacific, dry Siberian winds flowing over the warmer Kuroshio Current produce a series of cyclonic cold fronts that collectively have long fetches and generate the high waves for the Pacific coast of North America during the winter. These storms

produce swell when they are distant and sea waves as their tracks near the coast.

Long, low-frequency swell is typically absent from leeward coasts. Rather they are subject to periodic, intense sea waves from the poleward and easterly traveling storm fronts. The tracks for tropical (summer/fall) storms for the middle Atlantic seaboard of the United States essentially follow the Gulf Stream, while the extratropical (winter/spring) tracks take a more easterly course. In midlatitudes the tracks are channeled between the coast and the perimeter of the Bermuda high. This channel is sometimes known as the Atlantic seaboard cyclone track (e.g., Klein and Winston 1958). In the vicinity of Cape Hatteras, the highest waves from both extra- and tropical storms are from the northeast. The storm tracks are offshore and generally parallel the coast, so that the counterclockwise flowing winds in the northwest quadrant of the cyclone (winds blowing from the northeast) are the strongest and have the longest fetch (e.g., Dolan



Energy and Sediment Budgets of the Global Coastal Zone, Fig. 1 Prevailing winds and storm tracks for the world oceans

et al. 1990). Thus, prevailing waves along the windward coasts and the strongest cyclonic sea waves along the leeward coasts both have net components toward the equator. As a consequence, there is a worldwide tendency in midlatitudes (20° – 45°) for the net wave-induced littoral transport of sand to be toward the equator.

A three meter-high surface wave at sea transmits 100 kW for every meter of wave crest, or 100 MW/km. Since wave energy increases as the square of wave height, most coastal energy is from high waves during storms, particularly storms of long duration. For example, the northeast storm of March 1989 along the Outer Banks of North Carolina had average deepwater wave heights of 3 m (Dolan et al. 1990). These were not unusually high waves, but they persisted for 115 h and had a power expenditure of 11,500 kilowatt hours per meter (kWh/m) of coastline. The average energy flux of waves on the Outer Banks is about 2 kW/m, which sums to a yearly total of 17,500 kWh/m (Inman and Dolan 1989). Thus this one storm had a power expenditure (and sand transport potential) equivalent to 66% of that for the average year.

Data of instrumented buoys and light vessels in the North Atlantic and North Pacific all show that wave heights have increased significantly and continuously during the past 25 years (e.g., Graham and Diaz 2001). The North Atlantic data show that the average annual (root-mean square) wave height has increased about 2% per year between 1962 and 1986. Over the 24-year period (1962–86) the range in average annual *significant wave* ($\sqrt{2}$ times root-mean square) height at Seven Stones Light Vessel off Lands End, southwest England, ranged from 2.2 to 2.9 m with an annual increase of 3 cm/year. The “50-year return value” ranged from 12 to 18 m with an increase of 20 cm/year (Carter and Draper 1988). Data from six buoys between latitude 34° and 56° in the eastern North Pacific show similar increasing trends over the period 1975–99. The annual average significant wave height increased on average about 2 cm/year, ranging for the six buoys from 0.4 to 4.2 cm/year (Allen and Komar 2000).

A preliminary estimate of global wave energy can be attempted from the data obtained from instrumented buoys and weather stations in the North Pacific and North Atlantic. The measurements for the eastern North Pacific cover the 25-year time interval from 1975 to 1999 (Allen and Komar 2000). Over this period, the average annual (root-mean square) wave height increased progressively from about 1.6 to 1.9 m, generating a landward wave energy flux that ranged for these average waves from about 93×10^6 to 131×10^6 kW over the same period. Thus, the wave height increased about 20% while the corresponding energy flux increased about 40%. Similar wave conditions and wave height increases over about the same period were observed for the eastern North Atlantic Ocean (Carter and Draper 1988). If it is assumed that the energy partitioning between

periods of calm and storm results in an energy flux twice that estimated from the average annual wave height (i.e., the significant wave assumption), then the total annual energy flux over the 25-year period in the North Pacific becomes about 190 and 260×10^6 kW with an average for the period of about 230×10^6 kW for this 30° – 60° latitude portion of the windward coast of North America.

When the energy flux of 260×10^6 kW for the North Pacific Ocean is extrapolated to the world oceans, the total wave energy flux to the global coastline becomes 2.5×10^9 kW (Table 2, footnote a). This is in agreement with earlier estimates (e.g., Inman 1973) and gives an average energy flux of 5.7 kW/m to the 440,000 km of world coastline.

What is known about wave climate intensity suggests that windward coasts of oceans in the Northern Hemisphere are in phase with each other during El Niño and La Niña events and out of phase with lee coasts. In terms of decadal climate patterns, which are alternately El Niño or La Niña dominated, the prevailing wind stress and swell directions in the Northern and Southern Hemisphere appear to be out of phase (Trenberth and Hurrell 1994; Chao et al. 2000). For example, during the La Niña dominated climate of the third-quarter of the twentieth century, waves from the southern ocean appeared as “southern swell” along the California coast. Southern swell was absent from the California coast during the previous quarter of the century and again disappeared during the last quarter century (Inman and Jenkins 1997). This is because of a global realignment in predominant storm tracks with changes in the phase of Pacific Decadal Oscillation (PDO), where El Niño storm tracks tend to be more zonal than those of La Niña storms.

Energy and Sediment Budgets of the Global Coastal Zone, Table 2 Estimates of the flux of mechanical energy in the shallow waters of the world in units of 10^9 kW

Wind-generated waves breaking against the coastline ^a	2.5
Tidal currents due to surface tides on shelves and in shallow seas ^b	2.2
Currents due to internal waves and tides over shelves ^c	0.5
Large-scale ocean currents in shallow seas ^b	0.2
All other sources	0.1
Total	5.5

^aAssuming the average energy flux for the six instrumented NOAA buoys in the eastern North Pacific (Allen and Komar 2000), proportioned over 30° – 60° north latitude, is 230×10^6 kW (see text) and that for the 0° – 30° portion of the eastern North Pacific is one-half, and that for the leeward ocean coasts is one third that of the windward coasts, giving a total for the coasts bordering the North Pacific Ocean of 460×10^6 kW. The global flux is assumed to be equivalent to that from five and one-half oceans like the North Pacific Ocean (i.e., North and South Pacific Oceans, North and South Atlantic Oceans, Indian Ocean, and one-half “ocean” equivalence for all marginal seas)

^bInman and Brush (1973)

^cConservative estimate from Wunsch and Hendry (1972)

Consequently, the long waves of El Niño storms do not generally propagate out of the hemisphere in which they were generated. Conversely, the frontal circulation of La Niña storms is more meridional, generating waves that traverse latitude. Therefore, the Northern Hemisphere beaches are usually affected by Southern Hemisphere swells only when the Southern Hemisphere is in a La Niña dominated climate pattern.

Studies of windfields and pressure gradients indicate that the increase in wave heights in the eastern North Atlantic during the last quarter of the twentieth century resulted from intensification of the pressure gradients between the Iceland low and the Azores high which could be related to climate change associated with the North Atlantic Oscillation (NAO) (Kushnir et al. 1997). It is likely that some of the increase in wave intensity in the eastern North Pacific Ocean was associated with the PDO and increased sea surface temperatures during the last quarter of the century. However, it is unclear what portion of these increased wave heights in the North Atlantic and North Pacific oceans are also associated with global warming (Graham and Diaz 2001). If the increases are mainly due to decadal oscillations, they suggest that decadal climate lead to $\pm 30\%$ changes in wave climate (see entry on “► Climate Patterns in the Coastal Zone”).

Tsunamis

The most impressive waves in the sea are those generated by submarine earthquakes, landslides, and volcanic explosions (e.g., Van Dorn 1987). These waves, called *tsunamis*, are well known because of the loss of life and great damage to coastal structures associated with their passage. Of the many types of surface waves, tsunamis contain the highest instantaneous power surges. They are reported to have caused runup to heights as great as 50 m above sea level in modern times and to greater heights in ancient times. The explosion and collapse of the volcanic island of Thera (Santorini) about 1400 bc is thought to have caused a tsunami, about 11 m high at Crete (Yokoyama 1978), that may have ended Minoan maritime supremacy. The largest, well-documented tsunami was associated with the explosion and collapse of Krakatoa volcano in Sunda Strait between the Indonesian Islands of Java and Sumatra in August, 1883. The energy of the explosion is estimated to have been 2×10^{18} J (500 megatons of TNT) and it ejected 18 km^3 of material into the atmosphere (e.g., Simkin and Fiske 1983). The resulting tsunami was estimated to be nearly 40 m in height; it drowned 36,000 people and destroyed 5,000 ships. One ship, the *Berouw* was carried 2 km inland on Sumatra where it remains today. The tsunami was observed on tide gauges throughout the world, and dust from the explosion covered the entire world causing brilliant sunsets for many years.

Detectable tsunamis have occurred about once a year. However, truly large tsunamis with energy content as

high as 5×10^{15} J occur only about five times per century. Averaged over long periods, tsunamis would generate an energy flux of about 1×10^5 kW. Thus, because of their relative infrequency, these catastrophic waves contribute little to the worldwide budget of power.

Tides

The gravitational forces of moon and sun provide the driving forces for the ocean tides, with the moon's tide generating force being about twice that of the sun. The main lunar (semidiurnal, M_2) tide with a period of 12.42 h is the principal world tide; however, the interaction between lunar and solar forces provides a variety of tide-generating constituents. Locally, tidal amplitudes are determined in part by the interactions of the propagating oceanic tidal bulge with the shape of the ocean basin, favoring oscillations that are harmonics of the tidal forcing and the basin dimensions. The dominant tidal oscillations are usually referred to as *semidiurnal* (twice daily), *diurnal* (once daily), and *mixed* tides that are combinations of the two, such as those along the Pacific coast of the Americas.

The ocean surface tides, having wavelengths that are very long compared with the depth of the ocean, produce some motion even in the deepest water. However, this motion is usually only a few centimeters per second so that the energy dissipation by tides against the deep-sea bottom is slight. The principal tidal dissipation results from the flow of strong tidal currents in shallow areas, such as the Bering Sea and the Argentine shelf. One of the more impressive tidal coastlines occurs in Argentina. The tidal wave travels eastward through Drake Passage and then northward where it shoals over the broad Argentine shelf, attaining spring ranges of nearly 15 m at Rio Gallegos. From the southeastern tip of Tierra del Fuego to Bahia Blanca, a distance of 2,800 km, the tidal range is mostly in excess of 6 m.

“Tides are unique among natural physical processes in that one can predict their motions well into the future with acceptable accuracy without learning anything about their physical mechanism” (Cartwright 1977). Their prediction from tide gauge records has been possible since Darwin formulated the rules of harmonic tide prediction in 1883. But, prediction from harmonic constituents can be in error when changing wind fields and warm water associated with El Niño events raise and lower sea level by 30 cm and more.

The energy dissipated by various kinds of tidal currents against the shallow sea bottom is one of the major factors that causes the rotational deceleration of the earth. Much of our present knowledge of this dissipation results from investigations by geophysicists into the possible causes of the deceleration in the earth's rotation (Jeffreys 1920, 1970; Munk and MacDonald 1960). Tidal dissipation causes a lag in the tidal bulge of the Earth relative to the rotation of the earth-moon system. This bulge exerts a torque that results in a decrease in both the earth's rotation rate and in the lunar orbital velocity.

From telescopic observations of sun and moon over the previous two and one-half centuries, Spencer Jones (Munk and MacDonald 1960) determined that the moon's deceleration was $22.4 \text{ arcsec/century}^2$. This deceleration would be associated with an energy dissipation in the atmosphere, oceans, and solid earth of about $2.7 \times 10^9 \text{ kW}$. More recently, measurements of the change in the moon's semimajor axis using laser ranging techniques show that the frictional energy dissipation due to the M_2 tides is $3.1 \times 10^9 \text{ kW}$, giving a dissipation rate due to the solar and lunar tides of about $4.0 \times 10^9 \text{ kW}$ (e.g., Dickey et al. 1994). The allocation by process for these revised dissipations of tidal energy is not yet understood. The best estimates for the lunar tidal dissipation in shallow seas is that of Miller (1966). He used the method of energy flux into shallow seas, rather than the less accurate method of summing the energy loss due to frictional drag on unit area of the bottom. His value of $1.7 \times 10^9 \text{ kW}$ for lunar tidal energy dissipation, when corrected for the solar tidal flux, gives a total energy dissipation for shallow seas of $2.2 \times 10^9 \text{ kW}$ (Table 2).

Internal Waves

Internal waves are water oscillations that occur within the various density layers beneath the ocean surface. It appears that far more energy is transformed into internal (baroclinic) waves than was previously thought. The vertically integrated energy over the deep sea is typically about $4 \times 10^{-3} \text{ J/m}^2$ (e.g., Garrett and Munk 1979). Mechanisms generating internal waves include direct coupling between ocean surface waves and the density layers beneath the surface, scattering due to seafloor roughness (Cox and Sandstrom 1962), and generation by impingement of deep-sea tides on the slopes of the continental shelves and submarine canyons (Rattray et al. 1969).

Cox and Sandstrom (1962) found that internal waves were coupled to surface waves and estimate that the rate of conversion of tidal energy into internal wave energy in the Atlantic Ocean is roughly between 6 and 69% of the power of the surface tide. From this, Munk and MacDonald (1960) estimate that the conversion of energy from surface to internal modes on a global scale may be about $1.5 \times 10^9 \text{ kW}$. Most internal wave energy is contained in the areas of deep ocean, but some of the energy travels into shallow water and is dissipated against the sea bottom. Winant (1974) estimated that the dissipation due to runup of breaking internal waves and surges on the shelves of the world may range between 2×10^5 and $2 \times 10^7 \text{ kW}$.

Wunsch and Hendry (1972) measured currents from an array placed on the continental slope off Long Island, New York, and concluded that the onshore energy flux of internal waves was 6 W/m . Assuming this to be a typical value, they estimated that the worldwide onshore flux of energy would be about $0.7 \times 10^9 \text{ kW}$. A value of about $0.5 \times 10^9 \text{ kW}$ for the energy dissipation by internal waves over the world's shelves appears to be a reasonable estimate (Table 2).

Ocean Currents, Large and Small

The kinetic energy associated with large-scale ocean currents is very large compared with the energy in surface waves and tidal currents. Stommel (1958) estimates that the average kinetic energy of currents in the North Atlantic Ocean is about $2 \times 10^{17} \text{ J}$. Flows such as the equatorial and Antarctic Circumpolar currents transport water at rates of 50–200 Sv (Sverdrup = million cubic meter per second). Even the relatively small California current transports about 12 Sv, a discharge that is greater than that of all the rivers and streams in the world. In addition to the surface currents, the world ocean has a surprisingly large circulation of deep and bottom water estimated to be 50 Sv (e.g., Schmitz Schmitz 1995).

Intense western boundary currents, such as the Gulf Stream, develop instabilities with horizontal waves or meanders. These meanders form large loops that may close or pinch-off, forming circular eddies or rings. These *mesoscale eddies* are typically 100 km and more in diameter, and since they are in cycloidal balance, conservation of angular momentum causes them to persist for many months to a year or so, as they move thousands of kilometers away from the main current. Eddies that breakoff from the poleward side of the currents have warm-cores and cyclonic rotation, whereas those shed from the equatorial side have cold-cores and anticyclonic motion (e.g., Gill 1982). The details of the global distribution of the energy in mesoscale eddies is not well known except for satellite observations of western boundary currents. The effects of mesoscale eddies on the continental shelves is poorly understood, although it is known that shelf eddies are associated with wind changes and with coastal promontories (e.g., Davis 1985).

While the flow of major ocean currents is predominantly in deepwater, their boundaries overlap the continental shelves, and the currents dissipate some of their energy against the shallow sea bottoms. For example, the Guiana (North Brazil) Current, with a volume discharge of about 29 Sv, flows with velocities up to 1 m/s over the $0.5 \times 10^6 \text{ km}^2$ of shelf off the coastlines of the Guianas and northern Brazil (e.g., Fratantoni et al. 1995). The Falkland Current flows with velocities of 25–50 cm/s across the $1 \times 10^6 \text{ km}^2$ of shallow shelf off Patagonia. If the Guiana Current averages 50 cm/s and the Falkland Current averages 25 cm/s, they would dissipate energy at the rate of 0.13×10^9 and $0.03 \times 10^9 \text{ kW}$, respectively. Therefore, it seems likely that the worldwide dissipation of energy from the flow of ocean currents in shallow water may be about $0.2 \times 10^9 \text{ kW}$ (Table 2).

Other Sources

In addition to generating the ocean's waves, atmospheric winds, and pressures have important local effects nearshore. Pressure fronts generate shelf seiche, and the combination of pressure and wind stress on the water's surface produces local anomalies in sea level that induce circulation of water over the

shelf and in submarine canyons. Shelf seiche up to a meter in height or more are common along the Argentine coast (Inman et al. 1962), while the sea level anomalies associated with storm surges may exceed 4 m for severe storms (Pugh 1987). Also, winds blowing over the coastal zone dissipate energy on beach and sand dunes. On a worldwide basis, this dissipation may be about 1×10^7 kW, and because of the prevalence of sea breeze and other onshore winds, results in a net inland transport of beach sand.

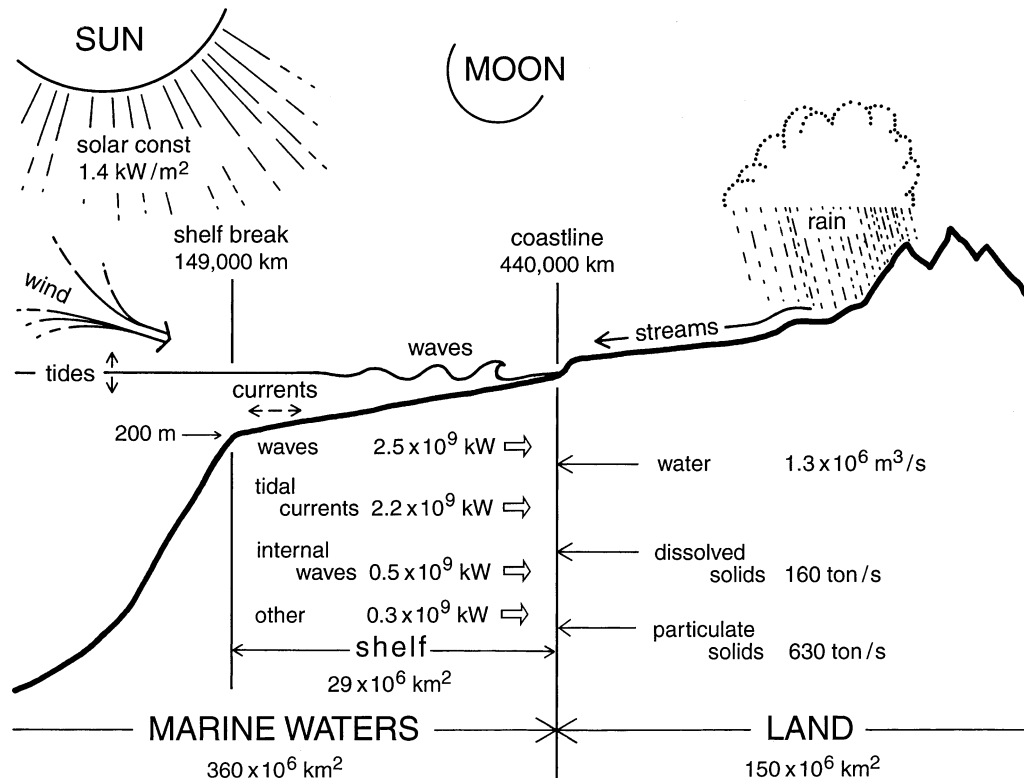
From the foregoing discussion, it appears that the total flux of mechanical energy in the shallow waters of the world is about 5.5×10^9 kW (Table 2). Dissipation and work by one process or another occurs over the entire shelf. The wind-generated surface waves expend their energy primarily near-shore, especially in the breaker zone, while tidal and other ocean currents and internal waves dissipate energy mostly over the outer portions of the shelf. Internal waves, edge waves, shelf seiche, and local winds may produce water motion over the continental shelf and submarine canyons.

Budget of Sediment

The coastline is the junction of the realms of terrestrial erosion and marine deposition. Thus, it is the unique singularity between air, land, and sea, and the coastline is the critical

datum for dynamical geology. In 1955, James Gilluly calculated that sediment is now being carried across this boundary at a rate great enough, in the absence of mountain building, to “erase all the topography above sea level in less than 10 million years . . .,” a very short time on the geological timescale. Thus, the supply of sediment associated with runoff from the continents is large, and it has increased during the past 10 million years due to climate variability, especially during the Pleistocene and more recently due to human intervention. The principal sources of beach and nearshore sediments are the rivers that bring large quantities of sand directly to the ocean, the sea cliffs of unconsolidated material that are eroded by waves, and material of biological origin such as shells, coral fragments, and skeletons of other small marine organisms (Fig. 2). Also, waves may transport onshore relict sands that were deposited along the inner shelf during lower stands of the sea (e.g., Inman and Dolan 1989).

Sediment yield increases with the relief of the drainage basin and with decrease in basin size, factors that enhance the sediment flux of small rivers along mountainous coasts (Inman and Jenkins 1999). Climate, rainfall, and type of geologic formation are also well-known factors in determining sediment yield. Maximum yield occurs for basins in temperate climates that receive an annual rainfall of about 25–50 cm, and decreases for precipitation rates either above or below this maximum (Langbein and Schumm 1958).



Energy and Sediment Budgets of the Global Coastal Zone, Fig. 2 Schematic diagram of budget of energy dissipation in nearshore waters, and the runoff of water and sediment from the land (after Inman 1973). (Data from Tables 1, 2, and 3)

In arid climates, the occasional flash floods transport large volumes of sand (Fig. 3).

Man's intervention in natural processes is accelerating erosion rates, perhaps by a factor of about two on a global scale (Milliman and Syvitski 1992; Vitousek et al. 1997). Significant anthropogenic effects began as early as 9,000 years ago with deforestation and the spread of agriculture from the "Fertile Crescent" (e.g., Heun et al. 1997). Man's intervention has accelerated in this century with increased deforestation, expansion of mechanized agriculture, proliferation of dams (Meade 1996), and since World War II with extensive urbanization of coastal lands particularly in central and southern California. Urbanization decreases erosion locally, but accelerates streambed erosion in response to the greater frequency and magnitude of peak streamflow due to runoff from impervious urban surfaces (e.g., Trimble 1997).

The effects of changing climate and sea-level rise have been studied extensively on a glacial/interglacial timescale (e.g., Hay 1994) and during the Holocene on a millennial scale (e.g., Bond et al. 1997). The effect of decadal scale climate changes on the sediment flux of the 20 largest streams entering the Pacific Ocean along the central and southern California coast was studied by Inman and Jenkins (1999). The annual streamflow ranged from zero during dry years to a maximum of $1.1 \times 10^9 \text{ m}^3/\text{year}$ for the Santa Clara River in 1969 with an associated suspended sediment flux of $46 \times 10^6 \text{ ton}$. Trend analyses showed that El Niño/Southern Oscillation (ENSO) induced climate changes recur on a multidecadal

timescale in general agreement with the Pacific/North American (PNA) climate pattern: a dry climate extending from 1944 to about 1968 and a wet climate extending from about 1969 to 1998. The dry period is characterized by consistently low annual river sediment flux. The wet period had a mean annual suspended sediment flux about five times greater, caused by strong El Niño events that produce floods with an average recurrence of about 5 years. The sediment flux of the rivers during the three major flood years averaged 27 times greater than the annual flux during the previous dry climate (see entry "► Climate Patterns in the Coastal Zone").

The effects of climate change are superimposed on erodibility associated with basin geology. The sediment yield of the faulted, overturned Cenozoic sediments of the transverse ranges of California is many times greater than that of the coast ranges and peninsular ranges. Thus the abrupt transition from dry to wet climate in 1969 brought a suspended sediment flux of 100 million tons to the ocean edge of the Santa Barbara Channel from the rivers of the Transverse Range, an amount greater than their total flux during the preceding 25-year dry period. These alternating dry to wet decadal scale changes in climate are natural cycles that have profound effects on fluvial morphology, engineering structures, and the supply of sediment.

Cliff erosion probably does not account for more than about 5% of the material on most beaches. Wave erosion of rocky coasts is usually slow, even where the rocks are relatively soft shales. On the other hand, erosion rates greater than 1 m/year are not uncommon in unconsolidated sea cliffs and

Energy and Sediment Budgets of the Global Coastal Zone,

Fig. 3 Streams carry erosion products from the land to the sea; El Moreno, Gulf of California, Mexico



dunes as found on trailing-edge coasts such as the east coast of the United States.

Following initial deposition at the mouths of streams entering the ocean, much of the sand-size portion of terrestrial sediments is carried along the coast by waves and wave-induced longshore currents. The sand carried by these longshore currents may be intercepted by submarine canyons that cut into the continental shelf and divert the sand into deeper water offshore. Since submarine canyons occur along most coasts, except those bordered by shallow seas, they probably account for most of the sediment lost into deepwater from the nearshore. Notable among the major sediment transporting submarine canyons of the world are Cap Breton Submarine Canyon on the Atlantic coast of France; Cayar and Trou Sans Fond, and Congo Canyons on the Atlantic coast of Africa; the Indus and the Ganges Submarine Canyons in the Indian Ocean; and the Monterey Submarine Canyon on the Pacific coast of North America. Observations by H.W. Menard (1960) suggest that most of the deep sediments on the abyssal plains along a 400 km section of the California coastline are derived from two submarine canyons: the Delgada Submarine Canyon in Northern California and Monterey Submarine Canyon in central California. These canyons cut the continental shelf almost to the shoreline, and consequently they trap and divert an estimated sediment volume of about $1 \times 10^6 \text{ m}^3/\text{year}$ (see entry on “► Littoral Cells”).

Traditionally, geologists have estimated long-term erosion and deposition rates from the amount of material deposited during geological time. Such estimates give erosion rates varying from about 1 to 4 cm per thousand years (cm/kyr) for drainage basins of moderate relief to as high as 21–100 cm/kyr for the Himalaya Mountains (e.g., Gilluly 1955). Assessment of the volume of sedimentary material on the continental United States and in its adjacent seafloors indicates that the average erosion rate of the United States during the past 600 million years was 3–6 cm/kyr (Gilluly et al. 1970).

An increasing number of measurements of river discharge have provided an independent estimate of contemporary erosion rates of the land. The importance of large rivers in transporting the denudation products of the continents to the sea has been known since Lyell (1873) described the flux of sediment into the Bay of Bengal from the Ganges and Brahmaputra Rivers. Since then the contributions from large rivers have been updated and summarized in many studies (e.g., Garrels and MacKenzie 1971; Inman and Brush 1973; Meade 1996) with estimates of the total flux of particulate solids to the ocean of about $16 \times 10^9 \text{ ton/year}$. The importance of small rivers to the global budget of sediment was first documented by Milliman and Syvitski (1992). They showed that small rivers (drainage basin $<10,000 \text{ km}^2$) cover only 20% of the land area but their large number results in their

Energy and Sediment Budgets of the Global Coastal Zone, Table 3 Estimate of the natural runoff of fresh water and solids from the continents into the oceans and coastal waters of the world

Discharge of water into the oceans from all rivers ^a	$1.3 \times 10^6 \text{ m}^3/\text{s}$
Flux of dissolved solids ^b	160 ton/s
Flux of particulate solids ^b	630 ton/s
Flux of particulate solids from sea cliff erosion ^c	$>2 \text{ ton/s}$
Flux of eolian particulate solids ^d	14 ton/s
Average erosion rate of land ^b	6 cm/kyr

^aAverage value ($1.1 \times 10^6 \text{ m}^3/\text{s}$) from Turekian (1971), Alekin and Brazhnikova (1961), and Holeman (1968) updated in proportion to increase in particulate solids from 530 to 630 ton/s

^bAssuming that continental rock of density 2.7 ton/m^3 erodes to form 20% dissolved load and 80% solid particulate load. Assuming flux of particulate solids is 630 ton/s (Milliman and Syvitski 1992)

^cEmery and Milliman 1978

^dWindblown dust, volcanic and forest fire ash (Prospero and Carlson 1981; Rea 1994; Prospero et al. 2002)

collectively contributing much more sediment than previously estimated, increasing the total flux of particulate solids to about $20 \times 10^9 \text{ ton/year}$ (630 ton/s) by streams and rivers (Table 3).

Balance of the Budgets of Power and Sediment

The global flux of erosion products from the land to the sea include 160 ton/s dissolved solids and 630 ton/s particulate solids (Table 3). The dissolved solids are incorporated into the $1.3 \times 10^6 \text{ m}^3/\text{s}$ flow of “fresh” water as added mass. Most of the particulate solids carried by stream-flow are fine silt and clay size material that move as suspended load and washload. The remaining solids are sand and coarser material that move mostly as bed and suspended load. Once the stream discharges into the sea, there are significant differences in the transport paths of the suspended and bedload sediment. Most of the fine suspended sediment moves with the river water and flows out over the sea as a spreading turbid plume, and the sediment, which is subject to flocculation and ingestion by organisms, is eventually deposited in deeper water. The coarser bedload material remains near the river mouth as a submerged sand delta and is later transported along the shore by waves and currents to nourish the downcoast beaches. The few reliable measurements of bedload indicate that it is ca. $\leq 10\%$ of the total load in large rivers but, in smaller, mountainous streams, may be considerably more (Richards 1982; Inman and Jenkins 1999).

It is of interest to compare the global fluxes of energy and sediments into the coastal waters of the world (Fig. 2). Since energy provides the capacity for doing work, the flux of energy associated with waves and currents is a measure of their potential to transport sediment. About one-half ($2.9 \times 10^9 \text{ kW}$) of the total flux of mechanical energy is associated with currents of various types, internal waves,

and turbulence, and this flux acts primarily upon the fine particulate solids that flow out over the ocean surface. The remaining energy (2.5×10^9 kW) is from ocean surface waves and works primarily on sand size solids that are deposited in shallow water. About 85% (535 ton/s) of the particulate solids are fine material, while about 15% (95 ton/s) are sand and coarser.

Waves move sand on, off, and along the shore. Once an equilibrium beach profile is established, the principal transport is along the coast. Theory and field measurements show that the littoral transport rate of sand along a beach is proportional to the energy flux of the incident waves, where the proportionality is a function of the direction of wave approach relative to the shoreline. If 10% of the world's estimated annual wave power at the coast (0.25×10^9 kW) is available for transporting sediment (i.e., the wave breaker angle is about 5°), the waves have the potential to transport the entire sand size load of solids in the world's streams (95 ton/s), a distance of 750 km along the coast each year. Thus, even under the inefficient coupling that exists in nature, it is apparent that the potential for sediment transport in nearshore waters is large.

Cross-References

- ▶ [Accretion and Erosion Waves on Beaches](#)
- ▶ [Beach Processes](#)
- ▶ [Climate Patterns in the Coastal Zone](#)
- ▶ [Coastal Currents](#)
- ▶ [Erosion Processes](#)
- ▶ [History of Coastal Geomorphology](#)
- ▶ [Human Impact on Coasts](#)
- ▶ [Littoral Cells](#)
- ▶ [Longshore Sediment Transport](#)
- ▶ [Meteorological Effects on Coasts](#)
- ▶ [Sediment Budget](#)
- ▶ [Tsunami Deposits](#)
- ▶ [Waves](#)
- ▶ [Weathering in the Coastal Zone](#)

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Engineering Applications of Coastal Geomorphology

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Coastal Geomorphology

Literally, the title “geomorphology” comprises earth, shape and science with global aspects in that order. The study of earth science, from the strictly biblical concept of evolution to modern applied geomorphology, has been a long process in itself. In a generic term, geomorphology is indisputably a part of geology, but in institutional affiliation it has been a course in geography in almost all countries around the world. Coastal engineers have also contributed to its development in more recent time. Traditional geomorphology is concerned with the local landforms on the earth's surface and their processes over time. Coastal geomorphology, as a branch within the traditional geomorphology, is a relatively new discipline. It explores the relationship between coastal landforms and their processes affected by factors associated with climatology, oceanography, fluid mechanics, sedimentation, and geophysics.

Before the World War II, geomorphology – including coastal geomorphology – was largely Davisian, based on the concept of cyclic development and erosion, and of Johnsonian, in which the details in short-term coastal changes were ignored. Both approaches considered only the changes in vast stretches of time and space, and were once regarded as a “rather academic and, literally useless approach to coastal geomorphology” (e.g., Pethick 1986, p. 235). Although these concepts had suffered a major change during the War, their influence can still be traced.

Beginning in the 1940s, detailed work and examples of applied coastal geomorphology have emerged. By the 1950s, the approach had become process-oriented. This advanced further, especially in the 1970s and 1980s. Since then, numerous reports have become available resulting from the work carried out by coastal engineers on various coastal problems around the world.

Between the 1950s and 1980s, another almost universal approach taken by coastal geomorphologists was to attempt

a classification of coastal landforms – to recognize and describe coastal features. During this period, some major classifications on coasts were: (1) *submerged, emerged, and neutral* coasts due to past sea-level variations, (2) four regional types based on plate tectonic theory, (3) *primary* and *secondary* coasts based on the influence due to terrestrial agency or marine processes, (4) *high, moderate, and low-energy* beaches, (5) *east coast* and *west coast* environment depending on the relative location of the landmass, and (6) *storm* versus *swell* wave environments based on the level of wave energy inputs. However, applications of coastal geomorphology to stabilize eroding beaches were scarce.

Despite the diversity of these technical classifications, it is obvious that many physical features, in terms of the material and the types of physiographic units, can readily be identified along the ocean's edge. These include cliffs, rocky coasts, gravel beaches, sandy beaches, estuaries, deltas, lagoons, coral reefs, salt marshes, muddy fields, and mangrove swamps. For the sandy beaches, notable forms are beach cusps, salients and tombolos, spits, barrier islands, river mouth bars, straight beaches, and especially, curved or headland-bay beaches. These distinct landforms are produced by variable interactions due to the prevailing oceanographical and meteorological factors, as well as the geology of the physiographic units. Davies (1980) has demonstrated a global viewpoint between the orientation of a sandy beach and the direction of the persistent swell waves, resulting from the global wind system. Unfortunately, many basic textbooks in coastal geomorphology and engineering have continued to handle the beach processes at a local level without mentioning its global ramifications. Coastline curvature has been cited in some texts, but no further studies were made on its usage.

Geographers have been attracted by the similarity in the shapes and profiles of various landforms in order to find an appropriate mathematical expression to fit these shapes. This may be seen in the classical examples found in hill slopes, glacial valleys, and the longitudinal profiles of rivers. Researches on coastal landforms have produced a parabolic equation for equilibrium beach profiles (Dean 1991) and a logarithmic spiral equation for the planform of headland-bay beaches (Yasso 1965). The latter was widely accepted by geographers and engineers since its inception, until Hsu and his co-workers critically assessed it in the late 1980s. Summary of the development in bay beaches and its applications for shore protection can be found in Silvester and Hsu (1993, 1997).

Headland-Bay Beaches

The coastline along the boundary between the land and the sea may be either straight over a large length or curved in plan separated by headlands, natural or man-made. The

curved beaches are aesthetically beautiful and much more stable, compared with their straight counterparts. They may be fully exposed to waves, partially sheltered or fully protected by natural headlands or engineering structures.

General Characters of Headland-Bay Beaches

Of the curved beaches, some may be curved gently, others more indented, or with secondary features along its length. This unique geomorphic feature is ubiquitous not only on oceanic margins, but also on coasts of enclosed seas, lakes, and river foreshores. The bay beaches formed between natural headlands on the coasts can be easily identified on nautical charts, maps, aerial photographs, remote-sensing images, and even in travel brochures. They also appear downcoast of groins, and as salient or tombolo in the lee of protruding headlands, detached (offshore) breakwaters, and breakwaters for harbors. These curved beaches are found in various shapes and sizes, in relation to the prevailing waves and natural outcrops or headlands. The curved planform is produced by the persistent swell waves diffracted from the tip of a headland, combined with wave refraction and a nearshore current circulation system in the lee of the headland.

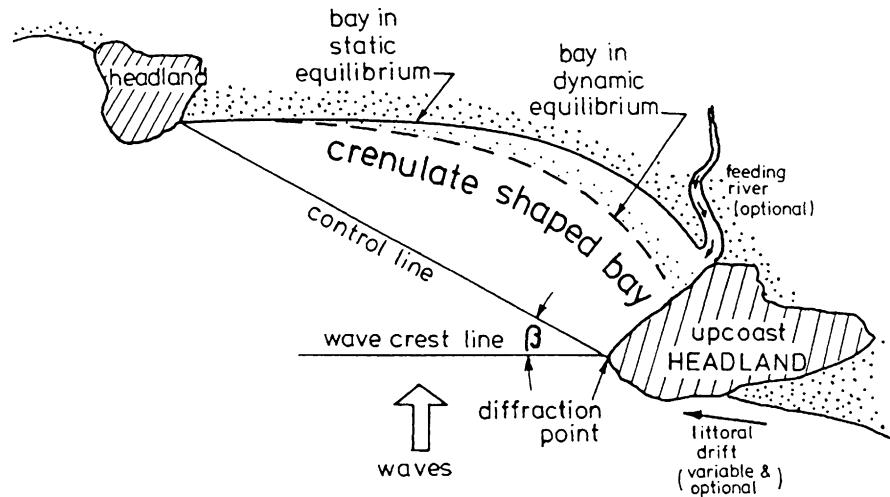
The curved or embayed beaches have attracted numerous names (Silvester and Hsu 1993), such as zeta curved bays, half-heart shaped bays, crenulate shaped beaches, logarithmic spiral beaches, curved or hooked beaches, headland-bay beaches, pocket beaches, and offset coasts. Despite their richness in existence and variety in names, stability of bay beaches was not fully tested by geographers and coastal engineers.

Since sandy beaches are constructed by persistent swell waves, their high-tide shoreline orientations are expected to align with the crest lines of these waves. Consequently, a series of asymmetrically curved bays can be found in many places. For example, they can be observed from Sydney to Fraser Island in eastern Australia, from Ilha de Santa Catarina to Rio de Janeiro in Brazil, and in other places. In these regions, the zeta shaped beaches have curved southern ends where they display a lower berm with finer sand particle and less steep beach faces than their straighter and more exposed northern ends. The consistency of coastline indentation and orientation can also be found in other continents around the world. It has taken some decades for this early recognition of bay shapes to become useful for stabilizing beaches in erosive conditions (Silvester 1979).

Silvester (1960) has questioned whether a bay beach is a stable physiographic feature. He believed that the orientation of the bay beaches is nature's method of balancing wave energy and load of sediment transport. In this manner, beaches have been maintained in position for tens, or even hundreds of years. The orientation of these bays also serves as an excellent indicator of the direction of net sediment movement along the coast. This macroscopic view of the coast

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Fig. 1 Headland bay beaches in dynamic and static equilibrium



imposed by nature is an integrated part of the coastal geomorphology, in addition to the microscopic view of wave kinematics and sediment movements in the marine environment.

Before the stable bay shapes can be applied for useful purpose, its planform has to be quantified and its stability assessed. In geomorphic term, the stability of a bay beach (Fig. 1) may be in *static equilibrium* or *dynamic equilibrium*, or even *unstable*. A bay beach is: (1) stable and in *static equilibrium*, if the littoral drift is negligible or supply from upcoast is nonexistent, or (2) stable but in *dynamic equilibrium*, when the littoral drift is still being supplied from upcoast and/or from sources within the bay, or (3) *unstable*, implying potential retreat due to imbalance of sediment input and output. While sand supplied to the system is balanced by that which can be removed by oblique waves, a bay beach can remain in place for decades in dynamic equilibrium. However, this is a rare case today. On the other hand, should supply decrease, due to various reasons to be discussed later in "Causes of man-made erosion," the high-tide shoreline will recede toward the static equilibrium shape, which is the final limit without the impediment of shoreline protection structures.

Three major mathematical expressions have been proposed to fit the planform of bay beaches since 1940s. They are: (1) the logarithmic spiral equation suggested by Krumbein (1944) and quantified by Yasso (1965), (2) the parabolic equation for bays in static equilibrium given by Hsu and Evans (1989) and later reassessed by Silvester and Hsu (1993), and (3) the hyperbolic-tangent equation suggested by Moreno and Kraus (1999). These empirically derived equations have different coordinates systems and origins, and controlling parameters related to bay orientations.

With a computer, all three methods can be applied to fit the planform of any bay beach (Moreno and Kraus 1999). But there are some important differences, especially, in the recognition of the point of wave diffraction, the location of

the coordinates origin, and beach stability requirement. The log-spiral and hyperbolic-tangent shapes have some drawbacks. First, their coordinate origins do not coincide with the tip of the headland (e.g., the point of wave diffraction). Second, these methods do not address the issue of beach stability or their equations were not produced from bays known in static equilibrium. Third, these two methods do not provide the resultant beach parameters, such as radii and angles, to ease potential applications. As such the log-spiral and hyperbolic-tangent forms cannot determine the effect of sheltering and future extension of any existing headland, or to predict the environmental effect of a new headland or structure on a bay beach.

Parabolic Equation for Bay Beaches in Static Equilibrium

Hsu and Evans (1989) developed a parabolic equation to fit the peripheries of 27 cases of bay beaches believed to be in static equilibrium in prototype and model conditions. This equation is more preferred than the log-spiral and hyperbolic-tangent equations for the reasons mentioned above. The equation is reproduced here for further discussion:

$$R/R_0 = C_0 + C_1(\beta/\theta) + C_2(\beta/\theta)^2 \quad (1)$$

There are two independent parameters in Eq. 1. The first is the *reference wave obliquity angle* β , or the angle between the incident wave crest (assumed linear) and the *control line*, joining the upcoast diffraction point to the near straight downcoast beach. The second basic parameter is the *control line* of length R_0 , which is also angled β to the tangent at this downcoast beach end. The radius R to any beach point around the bay periphery is angled at θ from the same wave crest line radiating out from the point of wave diffraction. The three C constants that vary with angle β were generated by regression analysis to fit the peripheries of the 27 data bays used. Examples of bay shape verification can be found in Hsu and

Evans (1989) and Silvester and Hsu (1993, 1997). Flexibility in choosing the downcoast control point is a merit of this empirical approach.

Despite having variable dimensions and indentations between the tip of the headland and the downcoast sandy boundary, the stability of a bay beach can now be determined using this parabolic bay shape equation. Values of the non-dimensional ratios of R/R_0 versus increments of 1° or 2° of β from 20° to 80° can be tabulated. This facilitates potential applications to locate the pairs of (R, θ) manually without the need of a computer, once β and R_0 are determined from maps or aerial photographs. When points of (R, θ) are drawn on the plan of the existing beach, the nearness of the existing bay shape to that predicted for static equilibrium can be compared and the beach stability assessed. If the predicted curve, for a hypothetical static equilibrium, is landward of the existing beach, the bay beach is said to be in dynamic equilibrium, or it may become unstable, for diminishing sediment supply.

Beach Erosion and Conventional Protection Methods

Continuous conflict has taken place between humans and nature over the narrow belt directly landward of the coastline of the world, where human pressure has been extensive. Within the flat ribbons of sand, between the sea and the hinterland, there are numerous problems of coastal flooding and beach erosion.

Causes of Human-Induced Erosion

The phenomenon of coastline changes are often complicated by oceanographic and meteorological effects on the inherent geological factors (form and material). With swell waves, nature helps move sediment landward to build up the berm and dune, while storm waves transport a large quantity of material to form a protective bar, causing beach disappearance. If waves approach a beach obliquely, a longshore current is induced, which promotes longshore sediment transport, or *littoral drift*. Compared with other landforms on the earth, which have evolved over a vast stretch of time and space, changes in high-tide shoreline positions and variations in beach profile may take place in a matter of hours. It is within this relatively rapid timescale in topographical change that has attracted the attention of many geographers and coastal engineers.

Coastline changes may be produced naturally or human-induced. Natural causes to beach erosion may include wave obliquity, storm attack, wind blown sand, existence of natural headlands, imbalance of sediment transport due to various topographical features, and sea-level rise. Although beach erosion has occurred naturally in many places, the problem is more serious due to the activities of people on the narrow

coastal strips locally or elsewhere. Overall, the major causes of human-induced erosion may due to: (1) disturbance to continuity of littoral drift, (2) wave sheltering by coastal structures, (3) offshore movement of sediment, (4) reduction in sediment supply from upcoast, (5) dredging or extraction of sand and gravel for construction, and (6) land subsidence (Uda 1997; Hsu et al. 2000). Among these, human intervention in and interception of sediment supply have been considered two of the most detrimental factors causing beach erosion in many countries around the world. However, some sources of erosion may be originated from a distant place away from the destination – the beach in erosion. For example, damming a river (for flood control, water supply, or power generation) reduces sediment supply to the beach. More recently, breakwater construction for fishing harbors and commercial ports have also caused beach erosion downcoast, with harbor planners and many consulting engineers responsible for the problems they have caused without considering the geomorphic effect on the beach.

Conventional Coastline Protection Methods

The family of seawalls (including bulkheads, revetments, sea dikes, and normal seawalls) were the very first structures ever constructed on a coast for protection against beach erosion and coastal flooding. They are built parallel to the beach at variable locations in relation to the local beach and surf. Their faces fronting the wave action may be vertical or inclined, with or without armor protection. They may even incorporate wave dissipation blocks near and in front of their toes, to reduce wave energy that causes scouring to the bottom at the toe. Beach flanking immediately downcoast of a long seawall is also common, for which some protection is still required.

Later groins were used to intercept littoral drift at places where erosion had occurred. They usually run normal to the beach and may reach out to the limit of the surf zone. Their length and height vary depending on the purpose they meant to serve. This type of structure has been widely used in many countries. Modern groins may be slightly curved, inclined to the local shore, or even in L-, T-, and Y-shaped or fishtailed. A bay beach may form downcoast of a protruding groin or between two units, where the tip of the upcoast unit serves as the point of diffraction, or upcoast control point for the bay beach so formed.

After groins, construction of detached breakwaters has become popular worldwide. They are either single or mostly in groups, with sandy salient or tombolo formed in their lee. In Japan alone, there were 7,371 units of such structures around the country by 1996, totaling 837 km (MOC 1997; Hsu et al. 2000). However, a new trend in Japan has been to replace them with submerged structures or artificial reefs, for aesthetic reasons.

More recently, beach nourishment has become the most “costeffective” alternative for coastal protection and creating recreational beaches, especially in the United States (Bird 1996). Often the beach fills were not protected by suitable structures, so replenishments were required at relatively frequent intervals. It has been envisaged that the cost of maintaining a wide beach would increase due to transporting the material from a distant source. In some cases, beach compartmentation was implemented.

Engineering Applications of Coastal Geomorphology

A trend has recently emerged in applying the principle of coastal geomorphology, especially the concept of stable bay shape, for beach protection. Examples showing a range of other engineering applications can also be found in the literature (Silvester and Hsu 1997).

Stability Verification for Existing Bay Beaches

With the parabolic bay shape equation (Hsu and Evans 1989), it is now possible to verify the stability of an existing sandy bay beach. The general procedure is as follows (refer to Eq. 1, with Figs. 1 and 2):

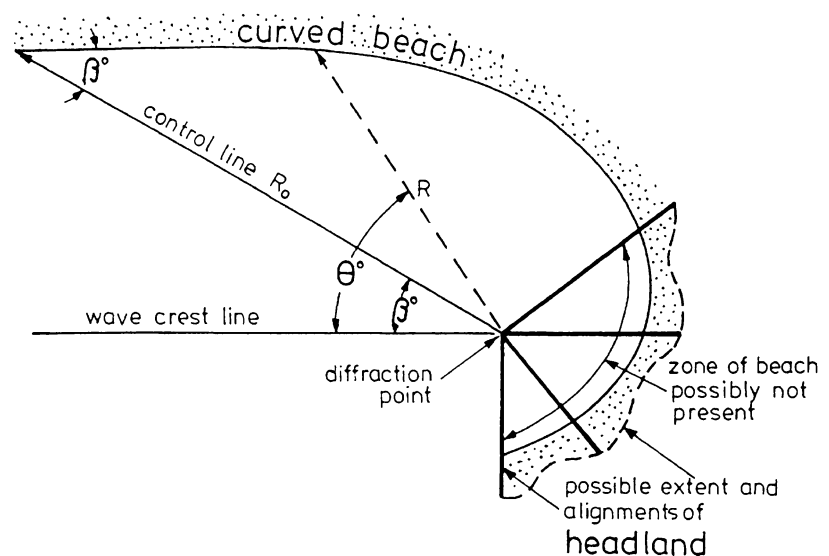
1. Draw a tangent from a point (referred to as the *downcoast control point*) on the straight downcoast section of a bay beach, and relocate this tangent to the point of wave diffraction (the *upcoast control point*), thus representing the *wave crest line* in the definition sketch (Fig. 2),
2. Denote the *control line*, by joining the upcoast and downcoast control points, assign the distance between them as R_0 ; also measure the reference obliquity angle β ,

3. Calculate the values of radii R for a range of corresponding angle θ (for $\theta \neq \beta$), using Eq. 1 or a prepared table of R/R_0 , so to give sufficient pairs of (R, θ) on the beach,
4. Sketch the predicted bay shape in static equilibrium by joining the points (R, θ) drawn on the existing beach planform,
5. Determine the stability of the existing bay beach by its nearness to the curve given by the prediction.

Should the existing bay shape conform well to the predicted bay shape in static equilibrium, the beach is said to be in *static equilibrium* under the prevailing swell condition. On the other hand, if the existing shape is seaward of the predicted curve, the beach is in dynamic condition, implying either in *dynamic equilibrium*, when sediment balance is maintained, or *unstable*, should sediment supply decrease. However, it must be remembered that the stable high-tide shoreline being predicted applies to swell condition or calm weather. When storm waves arrive, the beach berm will be eroded and material transported offshore to form a submerged bar, which may take some days or longer to return. In the case of static equilibrium, the same high-tide shoreline may result later, but all infrastructures should be kept landward of the receded storm limit.

The stage of bay beach stability may be switched between static and dynamic equilibrium. It can be either from a dynamic to another dynamic condition, or from a dynamic to static, or from static to a dynamic condition, depending on the initial condition of the bay beaches and variability in sediment balance and construction or extension of breakwater (Hsu and Silvester 1996). This concept of interchange ability for beach stability is extremely useful for practical applications. For example, it can be applied to: (1) the environmental impact study upon constructing or extending a jetty or

Engineering Applications of Coastal Geomorphology,
Fig. 2 Definition sketch for a parabolic bay shape in static equilibrium



breakwater within a bay beach in static equilibrium, and (2) the stabilization of a bay beach in eroding condition by locating the tip of a new headland to be installed at a proper location. In these situations, a stable bay beach in static equilibrium is created, thus minimizing the need for protective structures to be constructed downcoast later on.

Any new attempt to construct a jetty or breakwater for a marina or harbor at or within a bay beach already in static equilibrium will result in accretion in the lee at the expense of erosion downcoast, unless certain preventive measures are in place. This natural reshaping will take place within the entire embayment upon the construction of a new headland, despite some modest attempts with seawalls, groins, and detached breakwaters to prevent its fruition. On the other hand, should an engineering task be to construct a breakwater within a bay in dynamic equilibrium; the beach can conveniently be converted into static equilibrium by choosing a proper location for the tip of the new structure. This geomorphic approach of preventing beach erosion should benefit the community as a whole, in addition to the traditional engineering need for providing calm waves within a harbor.

Stabilizing Eroding Beaches

To combat widespread beach erosion, conventional methods have been used as mentioned in the previous section. Instead of constructing hard structures to defend the eroding beaches, it would be better to create a coastal environment in which the beach can protect itself by the formation of bays to meet the condition of diminishing littoral drift. This requires the reorientation of local high-tide shorelines to produce minimal or null littoral drift within the embayment. The concept is termed *headland control*, in which artificial headlands are used to help reorient the high-tide shoreline parallel to the crest lines of the prevailing swell waves. The concept can be applied to stabilize all shapes of eroding high-tide shorelines, be they straight, bayed, or convex in plan (Silvester and Hsu 1993, 1997). The artificial headlands may be either one of the conventional protective structures of groins and detached breakwaters with large spacing, or in combination. This can also be used to help retain beach fill staying longer on a renourished beach.

On a straight coast, a system of groins and/or detached breakwaters at close intervals has been used in many countries around the world. The concept of headland control can be applied to modify the conventional protection scheme already in place by removing some units thus widening the distance between them, or to install new headlands if protection did not exist. However, the most important difference, in the headlands used in the conventional defense and the present *headland control*, is in the distance between two adjacent units and their relative orientation to the coast. For a stable beach to form in the vicinity of these artificial headlands the space between two units must be sufficiently large, often to

several wave lengths, thus allowing a beach so formed to dissipate wave energy naturally under swell wave condition. The new beach should also have sufficient width and volume of material to satisfy the need of a design storm condition. Flexibility in varying the length, shape, distance offshore, and orientation of each unit would help optimize the performance of a stabilization scheme.

For a bay beach suffering erosion, due to reduction in supply upcoast and new breakwater construction for a harbor within an embayment, adding artificial headlands at appropriate locations can create several smaller bay beaches in static condition. The planform of these stable bays in static equilibrium can be sketched after simple calculations using Eq. 1, or aided by a prepared table of R/R_0 .

On an eroding convex coast, such as a barrier beach or spit, its narrow section may be overwashed or breached by storm waves accompanied by storm surge. Most barrier beaches have retreated in the past decades due to the supply from upcoast being drastically reduced. Attention should be given to a geomorphic solution rather than by hard structure alone. A series of stable bay beaches is preferred to a system of groins. The headland control concept allows the units to be installed with variable length, inclination to the coast, distance offshore, and spacing between units. The distance offshore and space between units should be gradually reduced in a cascade manner, thus allowing the local coastal processes to approach natural conditions. The reference wave obliquity angle/used in Eq. 1 for the successive bays formed between headlands should vary, being greater at the farther downcoast, where closer spacing is also desirable.

Salient Planform Behind Single Detached Breakwater

Waves diffract and refract in the lee of a detached breakwater and produce a nearshore current circulation system extending beyond the region of the shadow zone covered by wave diffraction. Consequently, sand accumulation in the form of salient or tombolo may occur behind the breakwater, depending mainly on the material available, the length of breakwater and its distance offshore from the original high-tide shoreline. Hsu and Silvester (1990) have demonstrated that the distance of the salient apex from the breakwater is preferred to that from the original high-tide shoreline, since the former relates more to the wave energy input and the effect of wave diffraction. Previous studies using the latter have resulted in unacceptable scatter when plotting the salient dimension versus the ratio of breakwater length and distance offshore.

They also illustrated that the curved beaches so formed can be related to the bay shape in static equilibrium, for which Eq. 1 can be applied. The analysis is limited to the case of normal wave incidence. The accretion assumes a parabolic form to a theoretical apex whose position can be determined using the dimensionless ratio of offshore distance to the

length of the breakwater, for which a table has been available (Silvester and Hsu 1997). This accretion is at the expense of the material beyond the extremities of the detached breakwater, and hence erosion at beaches downcoast. The final waterline at some distance either side of the breakwater will recede landward, implying beach erosion in this vicinity. However, to the knowledge of this author, this distance between the original high-tide shoreline and the receding one has not been fully quantified whether in laboratory experiments or in field conditions.

In the case of a salient formed in a multiple detached breakwater system, an accurate prediction method is not available. An attempt was made to analyze the planform using the same dimensionless ratios, as for single detached breakwater, and the gap width between units. However, the correlation based on the field data of the salients and tombolos collected in Japan did not yield reliable results for practical applications (Hsu and Silvester 1989).

Stabilizing Beaches Downcoast of Harbors

Many new harbors of variable capacities were constructed on coasts where no natural shelter is available, such as coastal indentation, offshore islands, headlands, capes, barrier spits, or reefs. This requires an initial breakwater to be built and possibly some extension over time as further commercial expansion demanding a large port, hence longer and larger structures. Even until recent time, harbor engineers and planners have considered that the construction of a harbor has priority over coastline evolution downcoast. They have often ignored the principle of geomorphology that would cause beach accretion in the lee of offshore breakwaters and erosion downcoast. Coastal scientists in Greece, Italy, Japan, and the United States have reported adverse effects on downcoast beaches (Silvester and Hsu 1997).

Similar to the sand accumulation behind single detached breakwaters, salient forms in the lee of a harbor breakwater, inevitably cause beach erosion downcoast and a possible silting of the refuge basin at the entrance. Compared with the case of single detached breakwater, a harbor breakwater does not have a clear middle point along its longitudinal axis. With the need for a breakwater to be extended over time due to commercial demand for a larger port, it will produce a larger salient in the lee, and hence erosion further downcoast (Hsu et al. 1993).

Hsu and Silvester (1996, p. 3986) have examined the situation of stabilizing beaches downcoast of harbor extensions. They believe that

In geomorphologic terms, the sandy shoreline of a bay downcoast of a harbor may be stable or in static equilibrium, or could be in dynamic condition, if sediment is still being supplied from upcoast or from downcoast to form a salient predicted by a static bay shape equation. Should commercial expansion demand a larger port, the general solution is to run a breakwater form

the headland or existing structure. This has the potential to create a new static equilibrium beach, often with accretion in the lee at the expense of beach erosion downcoast. It is strongly recommended that geomorphic approach be incorporated to stabilize downcoast beaches early in the planning stage of a harbor, or as remedial measures. By creating bay beaches in static equilibrium, the potential beach erosion downcoast of a harbor will be kept to a minimum or may be prevented completely. (Hsu and Silvester 1996)

Thus, instead of employing the conventional coastal protection method, in which seawalls, groins, and detached breakwaters were used without much success after the erosion downcoast has become serious, bay beaches of variable sizes should be considered to stabilize the eroding beaches. The geomorphic approach, involving the construction of stable bay beaches aided with beach nourishment, can be regarded as a permanent solution for this occasion. With the macroscopic view of the whole coastline rather than looking at a limited section, a better picture of the erosive problem will become clear. As such, erosion can be predicted long before it is likely to take place, and hence means can be devised to prevent it. Case studies have been reported for harbors in Australia, Japan, Taiwan, and South Africa (Hsu and Silvester 1996).

Designing Recreational Beaches

With the economic development around the world there is a community movement demanding a better coastal living environment and full utilization of the coastal zone. As a result, many artificial beaches have been constructed for recreation and tourism, as well as for the restoration of eroding beaches.

Silvester and Hsu (1997) have detailed the requirement for the creation of a stable bay shape for artificial and recreational beaches. One basic requirement is to construct suitable protective structures for preventing sediment loss from the artificial beaches and for providing a calm wave environment. Detached breakwaters with beach nourishment and/or groins have been used in all the cases so far reported in the literature.

Around the mid 1980s, two major Japanese ministries responsible for the preservation and protection of the coast against natural and man-made disasters launched two similar beach protection schemes incorporating hard structures and artificial beaches. These were the coastal community zone (CCZ) projects administered by the Ministry of Construction and the integrated shore protection system (ISPS) under the Ministry of Transport (Hsu et al. 2000).

Most of the beaches so designed are either partially or fully enclosed cells behind offshore breakwaters or between them, hence the term *pocket beaches*. Berenguer and Enriquez (1988, p. 1412) have observed that: "In the recent years, the increase of pocket beaches has been remarkable. In Mediterranean countries where tidal effects are insignificant, especially in Spain, Italy, France, and Israel, this system

of beach regeneration has been widely used because of its rapid results and its economical nature. Japan also has a lot of experience in this kind of work. In Spain, there are over 40 beaches of this kind. They are usually situated on stretches of coastline where there is a lot of tourism and in which there is also a fairly rapid erosional process . . .”.

Coastal Management and Planning

People have used public beaches for various purposes. But, many beaches around the world are still being subjected to erosion, due to improper coastal practices, human greediness and local political pressure for the convenient use on the narrow coastal strips. Coastal protection has also arisen from improper coastal zoning for residential and public development too close to the edge of the ocean, without knowing the stability of the beach involved. Many coastal communities have paid an expensive price through this experience of combating with the recurrent erosion problem. Another example is that landscape architects and consultants, who may not be familiar with the principle of coastal geomorphology, are engaged by an authority to beautify the coastal strips for tourism. Some of the additional facilities so created may fall prey to a storm, so the taxpayer's money is wasted.

It is essential in modern coastal management that beach managers in coastal communities be able to apply the principle of coastal geomorphology, and “to examine such questions as where the beach sediment came from, whether it is still coming, and whether the beach is stable, or shows long-term accretions or erosion, as distinct from short-term cut and fill or seasonal alternations” (Bird 1996).

For any development proposal on a bay beach, the stability of the existing beach should be investigated, prior to giving permission for residential development or structures to be built close to the water. The concept of the stable bay shape of Hsu and Evans (1989) is readily applicable to such cases, without the need of engaging in an expensive and time-consuming exercise, such as physical modeling and computer simulations. If the bay beach is in a dynamic condition, residential development and infrastructures should not be allowed seaward of the predicted static bay limit, plus a little buffer for a storm. Even if the bay is already in static equilibrium, a buffer distance is also required for a storm to act in the future. This would save the need for costly protective measures to be built later. A full investigation may then be carried out only after the preliminary desktop study using the principle of coastal geomorphology, if an elaborate exercise is still warranted.

Geomorphic Engineering: A Future Perspective

Coates (1980) has designated a specific term of *geomorphic engineering* to combine the fields of geology, engineering,

and environmental science together, which used to work independently on coasts. It is also important that biological processes in coastal geomorphology be considered. The traditional single task of coastal protection should gradually be replaced by the preservation of the coastal environment, by creating stable and safe beaches to coexist with nature. It is in such a manner that a careful balance among the conflicting requirements of protection, development, and conservation can be met, with planning and management using coastal geomorphology fulfilling the central role.

Cross-References

- Artificial Islands
- Beach Erosion
- Beach Features
- Beach Nourishment
- Beach Processes
- Coastal Changes, Gradual
- Coastal Changes, Rapid
- Dynamic Equilibrium of Beaches
- Headland-Bay Beach
- History of Coastal Geomorphology
- Human Impact on Coasts
- Sandy Coasts
- Shore Protection Structures

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Environmental Quality

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Synonyms

Environmental condition; Environmental health

Definition

Environmental quality is defined in this work as the condition of a particular marine environment (coastline, estuary, bay, harbor, nearshore and offshore waters, open ocean) measured in relation to each of its intended uses. It is usually assessed quantitatively and requires both indices of condition and change and established guidelines and objectives set by environmental health and resource agencies (Harding 1992).

Stressors, Monitoring and Assessment Programs, and Management

Estuarine and coastal marine environments are affected by a wide array of anthropogenic activities that have impacted environmental quality (Kennish et al. 2008; Kennish 2015).

This is so because more than half of the world's human population resides within 100 km of the coast, which creates multiple stressors in the coastal zone (Alongi 1998; Kennish et al. 2008; Kennish 2015). While environmental quality of estuarine and coastal marine environments is most frequently compromised by human activities, that of offshore waters and the open ocean may be adversely affected as well by waste dumping, atmospheric deposition of pollutants, climate change effects, and other human factors.

There are two major categories of anthropogenic stressors on estuarine and marine environments: (1) pollution and (2) habitat and resource degradation. Contaminant inputs from point and nonpoint pollution sources, as well as the physical alteration and destruction of habitat due to dredging and dredged-material disposal, coastal watershed development, marina and harbor construction, marine shipping and transportation, marine mining, mariculture, energy production (oil and gas recovery, electric power generation), shoreline installations, and other human factors, have degraded environmental quality. Overuse or misuse of coastal resources such as overfishing, excessive water withdrawal, subsidence-driven oil and gas extraction, and uncontrolled marine mining not only depletes natural resources but also can adversely affect habitats beyond the limits of sustainability. Overfishing reduces fishery production as well as genetic diversity and ecological resilience (Costanza et al. 1998). Marine sand/gravel mining impacts marine benthic habitats and communities and in some areas can deplete sediment needed for beach replenishment or can exacerbate shore erosion. Poor development planning and inadequate resource management have caused many acute and insidious environmental problems in the coastal zone around the world.

In the USA, enactment of the Clean Water Act in 1972, Clean Water Act Amendments in 1977 and 1987, and Oceans Act in 2000 greatly reduced point source pollution inputs to estuarine and marine environments. Most pollutants now enter these environments from nonpoint sources which are not easily regulated or controlled by land-based technological measures. Similarly, legislation has been adopted in Europe (European Water Framework Directive and European Marine Strategy) that forms the basis for determining the ecological integrity of surface waters of estuarine and coastal marine environments (Borja and Dauer 2008). As noted by Borja and Dauer (2008, p. 331), “An integral part of determining ecological integrity is the measurement of biological integrity, typically emphasizing analyses of plankton, benthos, macroalgae and fish.” They conveyed that effective ecological protection requires the following components: (1) assessing ecological integrity, (2) evaluating if significant ecological degradation has occurred, (3) identifying the spatial extent and location of ecological degradation, and (4) determining causes of unacceptable degradation in order to guide management actions.

Pollution degrades coastal water quality, adversely affects estuarine and marine organisms, causes health risks to humans, and results in diminished human use of coastal environments and resources. Examples include sewage and pathogen inputs, nutrient enrichment and eutrophication, and chemical contaminants (halogenated hydrocarbons, polycyclic aromatic hydrocarbons, metals, and volatile organic compounds). Chemical contaminants in particular can cause behavioral, immunological, neurological, reproductive, digestive, and genetic abnormalities, debilitating diseases, loss of physiological function, and death of estuarine and marine organisms. Pollutants enter estuarine and marine environments from direct pipeline discharges (point sources), nonpoint source runoff, river inflow, atmospheric deposition, and waste dumping. Habitats are impacted by pollutant inputs as well, leading to additional biotic impacts. Chronic exposure of organisms to pollution stress usually results in marked changes in population structure and function, including reduced abundance and biomass, increased mortality, and altered age/size distribution. Recruitment may be significantly reduced. At the community level, responses to insidious and cumulative pollution impacts frequently include dramatic shifts in species composition, decreased species diversity, loss of rare or sensitive species, and dominance by opportunistic species (Howells et al. 1990; Diaz and Rosenberg 1995; Diaz 2015). Imbalances in trophic systems typically occur.

Monitoring is a critical step in assessing environmental quality and determining the necessary management controls to remediate impacts and restore environmental conditions in impaired systems. Assessment of the status and trends in ecosystem health using an integrated approach will yield the information needed for effective long-term environmental management strategies. There are two primary reasons for conducting environmental monitoring: (1) to generate a database to address specific estuarine and marine problems and management needs in a local or regional context and (2) to define the status and trends in environmental quality in a broad spatial context or nationwide (Hameedi 1997). Several elements should be incorporated into estuarine and marine environmental quality monitoring programs to assess ecosystem health: (1) analysis of the main processes and structural characteristics of the ecosystem, (2) identification of the known or potential stressors, (3) development of hypotheses about how these stressor may affect the ecosystem, and (4) identification and application of the measures of environmental quality and ecosystem health to test the hypothesis (Harding 1992).

Meaningful biological measures of pollutant exposure, pollutant-induced adverse biological effects, and susceptibility of biota to environmental degradation are important in environmental quality monitoring programs. Habitat condition should also be targeted. Environmental quality monitoring programs typically include the following: (1) assessment of the temporal trends in the levels of environmental

degradation due to chemical contaminants and other anthropogenic sources; (2) determination of the magnitude and spatial extent of environmental degradation due to chemical contaminants and other anthropogenic sources; (3) incidence of pathological conditions, physiological dysfunction, and biological abnormalities in sentinel organisms resulting from environmental degradation; and (4) formulation of composite indices or sets of indicators that communicate information about the condition of the natural environment and are tractable toward public policy objectives (Hameedi 1997).

In the USA and other developed nations, a focus of estuarine and marine monitoring is the identification of specific sources of stress (e.g., municipal and industrial discharges, groundwater, oil spills, atmospheric deposition, etc.), biotic exposure to pollutants (in the water column or bottom sediments), and pollution effects (on populations, communities, and ecosystems). Biological measures have been used by major marine environmental quality monitoring programs to assess coastal environmental conditions on local, regional, and national levels. In addition, integrated measures of coastal ecosystem health have been applied that aggregate water quality and biological measurements into a quantifiable framework. Examples include the Index of Biotic Integrity, Benthic Index, and Sediment Quality Triad.

It is imperative for marine and coastal policy and management programs to formulate strategies that mitigate pollution inputs and habitat loss. Most effectively, this involves a concerted effort on the part of government agencies (federal, state, and municipal), academic institutions, business and industry, stakeholders, and the general public to address the myriad of problems in the coastal zone and the measures that must be implemented to restore, protect, and maintain the integrity and health of coastal environments. Over the past few decades, great effort has been expended to develop partnerships among these entities to address and resolve environmental quality problems.

An integrated global approach to maintain marine environmental quality has not been effectively formulated. A comprehensive global marine environmental management framework is needed to protect and preserve the integrity of marine systems worldwide. Instead, the global strategies for protecting marine environments derive from a patchwork of international and regional conventions, protocols, agreements, national laws, regulations, and guidelines. Greater international cooperation is needed to more effectively address marine environmental quality problems worldwide.

Conclusions

A consensus-building process has proven to be useful in establishing effective pollution abatement and control programs, revitalizing damaged habitats, managing exploited resources, and restoring environmental quality of estuarine

and marine ecosystems. An approach that has gained favor over the past few decades is to forge partnerships between government agencies, academic institutions, business and industry, and the general public to seek a unified goal for restoring, protecting, and maintaining the health of damaged estuarine and coastal marine ecosystems. Education and outreach initiatives have been implemented to increase public awareness and understanding of pollution and habitat problems, as well as other anthropogenic impacts, and to encourage public participation in the decision-making processes and stewardship of coastal resources. Future environmental management programs must focus on more integrated and well-coordinated initiatives to replace the historically narrow conservation and protection strategies, wasteful land-use practices, unsustainable coastal development (particularly in the face of climate change effects), resource-use conflicts, and escalating habitat degradation in the coastal zone.

Cross-References

- [Changing Sea Levels](#)
- [Coastline Changes](#)
- [Dams, Effect on Coasts](#)
- [Demography of Coastal Populations](#)
- [Dikes](#)
- [Dredging of Coastal Environments](#)
- [Estuaries](#)
- [Estuaries: Anthropogenic Impacts](#)
- [Human Impact on Coasts](#)
- [Marine Debris-Onshore, Offshore, and Seafloor Litter](#)
- [Mining of Coastal Materials](#)
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- [Sea-Level and Climate Change](#)
- [Shore Protection Structures](#)
- [Submerging Coasts](#)
- [Tourism and Coastal Development](#)
- [Water Quality](#)

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Eolian Processes

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Eolian processes are those processes relating to, or caused by the wind. The term eolian comes from the Greek *Aeolus*, the God of Wind. Eolian (aeolian – non-US equivalent) processes include the transportation and movement of sediments (particularly sand, clay, and silt), erosion of sediments, rocks and landforms, and the creation of dunes, sheets, and landforms. Eolian processes are most common on coasts and in arid environments, particularly hot and cold deserts.

Eolian sand transport is initiated when the wind velocity exceeds a critical threshold velocity. The threshold velocity required to initiate clay and silt transport is far greater than that of sand due to their platy nature and cohesive forces, but once motion is initiated, fine grained material (particularly silt) may stay in suspension for a considerable time period. Three main forms of grain transport take place, creep, saltation, and suspension. Creep refers to the movement of grains by rolling over the surface. Saltation (from the Latin *saltare*, to jump) is the bounding movement of grains in which grains rise almost vertically and then describe a long low trajectory downwind. Once saltation occurs, some energy is transferred to the surface by grain collisions and this aids creep and the ejection of further grains into the air stream. This process occurs at lower threshold velocities than that required for initiation of sediment transport and has been termed reptation by some authors (e.g., Lancaster 1995). Suspension occurs where grains are lifted sufficiently to overcome gravity for some period of time. The initiation of eolian sediment transport and eolian processes are affected by surface conditions (e.g., moisture, cements, and bonding agents, vegetation) surface roughness and topography.

Eolian transport causes deflation, which may merely lead to the slight lowering of a surface (e.g., beach, dune or sand plain), but may also be the dominant process leading to the development of deflation hollows (and slacks), basins, and plains. In coastal and desert environments, deflation leads to the formation of blowouts, parabolic dunes, and wind-eroded plains. Here, deflation erodes the surface down to the seasonally lowest water table, a calcrete or indurated horizon, a palaeosol, or rock and boulder strewn surface. Eolian transport also causes abrasion of landforms. Wind transported sand grains abrade surfaces forming ventifacts (a polished, faceted and/or abraded stone or pebble), yardangs (sharp crested linear ridge of wind-eroded rock), and zeugen (desert rock pillars and yardangs, which are considerably undercut; Whittow 1984).

In the coastal environment, eolian processes may lead to the development of various dune types. Where vegetation is important or present, coppice dunes, shadow dunes, fore-dunes, blowouts, and parabolic dunes may be formed. Where vegetation is less important and sediment supply is moderate to high, transgressive dunefields may occur and display a variety of dune types (e.g., sheets, barchans, transverse and oblique dunes, dome dunes, and star dunes) and associated landforms such as precipitation ridges and deflation plains and basins may be present (Hesp 1999). Deserts may display a wide variety of eolian dune forms including dome dunes, barchans, transverse dunes, longitudinal and linear dunes, shadow dunes, sheets, and star dunes. Each tends to be related to sediment volume, wind directional variability, and average annual velocity (Cooke et al. 1993). Vegetation processes are important in modifying eolian processes commonly leading to the complete colonization of once active dunefields in both coastal and desert environments.

Cross-References

- Beach Processes
- Desert Coasts
- Dune Ridges
- Eolianite
- Ripple Marks

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Eolianite

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Definition

Eolianite, also known as aeolianite, (a)eolian calcarenite, calcareous (a)eolianite, dune limestone, dune sandstone, kurkar (Middle East), or kunkar (India), is a generally consolidated coastal rock formation consisting of at least partially lithified windblown sand. Strictly eolianite could include any windblown sediment, such as silt (loess, brick earth), clay (parna), and volcanic ash, but conventionally it is restricted to dune sand, although finer sediment is sometimes present.

Evolution of Eolianite

Typically, eolianite is dune sand that has been partially cemented by secondary internal precipitation of carbonates from percolating groundwater. There is often evidence of several stages of such cementation. The proportions of carbonate vary but are typically at least 50% and often more than 90%, derived from shell debris, coralline material, bryozoans, and foraminifera, which lived on the seafloor. There are varying proportions of noncalcareous sediment, mainly quartz (quartzose sandstones with less than 50% carbonate are termed quartz-arenites).

Eolianite was originally described from Bermuda by Sayles (1931), although it had been noted by explorers on the Australian coast, where Vancouver in 1791 thought it was coralline while Charles Darwin in 1844 reported that it consisted of windblown sand and associated limestone. It is usually of Pleistocene age and may be overlain by unconsolidated Holocene dune topography. It often passes well below present sea level, for example, the eolianites of the Nepean Peninsula in Victoria, Australia, are Pleistocene dune formations, which extend more than 140 m below, and up to 60 m above, present sea level (Bird 1993). They contain fossil shells, plant remains, and occasional evidence of land animals.

Eolianite dune formations have been emplaced on the coast during Quaternary oscillations of land and sea level. During falling and low sea-level phases, sand derived from calcareous seafloor organisms was winnowed and carried landward by onshore winds to be deposited as transgressive dunes in the vicinity of the present coast. Similar sand was also carried shoreward during marine transgressions to form beaches and backshore dunes, as well as multiple beach and dune ridges parallel to the coastline. In southeast South

Australia, a series of eolianite ridges mark successively uplifted coastlines inland from, and parallel to, the present coastline between Cape Jaffa and Discovery Bay. They are separated by swampy swales that were formerly coastal lagoons like the Coorong, which extends behind the calcareous sand barrier on the coast of Encounter Bay.

Many eolianite coasts have a series of partly overlapping transgressive dune formations, indicating alternations of surface stability, when a capping of soil and vegetation developed, and instability, when the next transgressive dune moved in (Fig. 1). Each dune formation is capped by a soil horizon (paleosol) overlying a calcrete (limestone) layer formed by subsoil precipitation of carbonates. Traced laterally, these soil and calcrete layers vary in thickness and rise and fall as they mark out the contours of successively formed dune landscapes. Contemporary vegetation produced calcareous fossil root structures (rhizoconcretions) in the subsoil. The interbedded soils are up to 3 meters thick, typically terra rossa, incorporating eolian silt and clay accessions to the dune surface. Occasionally, the footprints of animals or birds are found on outcrops of sandstone, which were temporary land surfaces, buried when dune accretion resumed. Locally, there are soil pipes, cylindrical hollows formed where root penetration was followed by the local subsidence of soil into depressions deepened by solution and encased by precipitated calcrete. In places there are unconformities because the earlier dune landscape was eroded before the next dunes arrived. Gravel layers formed by the disintegration of calcrete outcrops during such erosion phases occur locally, sometimes cemented as breccias. Each phase of landscape stability came to an end with the arrival of the next transgressive dune formation, marked by dune sandstone, often with laminations (dune bedding) indicating the intermittent advance of a dune front and sometimes with calcareous fossil

stem structures (phytoconcretions) formed around buried vegetation. Subsequent erosion may expose root and stem concretions as “petrified forests,” as at Cape Bridgewater, Australia, and on Lord Howe Island, the dune calcarenite cliff behind Middle Beach shows columnar structures that were the trunks of palm trees engulfed by a rapid dune advance, the wood having been subsequently replaced by precipitated carbonates.

Commonly eolianites are capped by Holocene dunes, but where these are absent (or have been removed by erosion), exposed calcrete layers can develop karst topography (limestone mounds or ridges, pedestals, clay-floored vases and ditches, solution hollows). Some dune sandstone formations remain relatively unconsolidated and can be mobilized by wind action as active dunes if the overlying calcrete crust is removed.

Eolianite dune formations are sometimes sufficiently lithified to be cut back as cliffs up to 60 m high, the harder components (calcrete layers) dominating the profile while caves and coves are excavated in the less consolidated dune sandstones. These cliffs are fronted by subhorizontal low-tide shore platforms formed where shore rock outcrops are lowered by solution processes to a level that has become indurated by further carbonate precipitation (Hills 1971) (Fig. 1).

Eolianite dune topography is found on the southern and eastern coasts of the Mediterranean, in Morocco, on the Red Sea and southern Arabian coast, in western India, around the Caribbean (Kaye 1959), on the Brazilian coast, in South Africa, on oceanic islands such as Mauritius and Hawaii, and extensively along the western and southern coasts of Australia (Fairbridge and Teichert 1953; Semeniuk and Johnson 1985). On the eastern coast of Australia, beaches and dunes are quartzose, and eolianites have not formed: the

Eolianite, Fig. 1 Cliff and shore platform cut in eolianite at Jubilee Point on the coast of Victoria. The cliff section shows two Pleistocene dune calcarenite formations capped by calcretes and paleosols (arrowed), surmounted by unconsolidated Holocene dunes. The subhorizontal shore platform is exposed at low tide (Copyright-Geostudies)



dunes remain unconsolidated, either active and mobile or retained by a vegetation cover. On the eastern coast of the Mediterranean, there is a similar transition, the eolianite (kurkar) dune topography of the Israel coast passing southward into more irregular, mobile dunes on the coast of the Gaza Strip, where the sands have an increasing proportion of terrigenous (Nile-derived) quartzose sediment.

The submergence and dissection of eolianite dune topography formed during low sea-level phases are illustrated off the western and southern coasts of Australia. The rising sea during the Holocene marine transgression breached eolianite ridges and isolated Rottnest Island and Garden Island in Western Australia, submerging Cockburn Sound. Some submerged eolianite ridges have been planed off by wave action on the seafloor, but there are also remnants of submerged eolianite dune topography. Similar submerged eolianite dune ridges occur off the Bahamas, northeast Brazil, and Sri Lanka, and some of these have become reefs with coral and algal crusts.

Summary

Eolianite is a generally consolidated coastal rock formation consisting of at least partially lithified windblown sand. It occurs as dune formations that have been emplaced on the coast during Quaternary oscillations of land and sea level and contain calcareous sediment derived from biogenic deposits on the seafloor. Parallel or overlapping dune formations have been lithified by precipitation of carbonates and include calccrete layers and paleosols.

Some have been sufficiently lithified to form coastal cliffs and shore platforms and to be exposed as karst topography.

Cross-References

- [Beach Ridges](#)
- [Beachrock](#)
- [Cheniers](#)
- [Coastal Soils](#)
- [Dune Ridges](#)
- [Eolian Processes](#)

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Erosion of Coastal Systems

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Definitions

Coasts can be defined and better understood if they are analyzed as systems. Systems are composed of identifiable entities (elements or subsystems, such as cliff, dune, beach, nearshore), which change attributes (size, shape, elevation) and interact (changes in one will affect the other) in response to varying inputs of energy or matter. Generally, coasts are open systems (inputs can be gained or lost by the system) in a state of dynamic equilibrium, i.e., changes in morphology occur as the system adjusts to the varying inputs. The temporal and spatial scale in which the coastal system should be defined will depend on the issues or questions that are being analyzed (Woodroffe 2003). Analyzing coasts using a “systems theory” framework can help elucidating the complex nonlinear interactions that result in coastal change.

Coastal erosion is a morphological change due to loss of material from a coastal system or subsystem. It is generally expressed in terms of reduction in volume, lowering elevation, or landward displacement of a feature of interest, such as the cliff edge, dune toe, or high water line. Coastal changes can be quantified using net values or rates of change. Coastal systems are dynamic, and the rate and direction of change can vary through time; therefore, it is good practice to indicate the period of time in which changes were measured.

The Coastal System

In a coastal system (Fig. 1), local and regional settings, such as geology, sea-level changes, and sediment availability, are long-term control factors influencing the formation of landforms at varying spatial scales. The geology influences the configuration of coastlines (e.g., the presence of headlands and cliffs) and whether a local sediment source may be available (e.g., debris from chalk cliffs tend to dissolve rather

than form sand or silt particles). Geology can also control the rate of change and their variation through time and space. For example, cliff retreat is generally slower in sections exposing hard rocks or more resistant sedimentary rocks than in sections of poorly consolidated sedimentary strata. In coastlines formed by folded rocks, erosion rates can vary depending on whether the different angles of dip can increase or decrease the susceptibility to erosion.

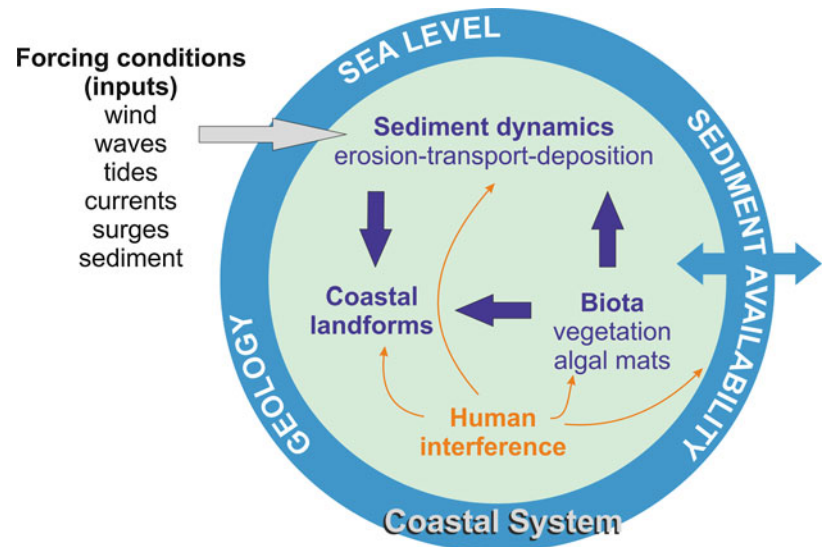
There are fundamental differences in the temporal variability of coastal changes between cliffed and low-lying sedimentary coasts. Cliffed coasts are erosional in nature, and cliff retreat is irreversible (i.e., there is no natural recovery from a cliff top retreat). Additionally, changes in cliffed coasts tend to be episodic, with large changes occurring in short periods of time interspersed with longer periods of very little change. In contrast, beaches show periods of accretion and

erosion as the morphology adjusts to decreasing and increasing wave energy, respectively. In low-lying sedimentary coastlines, both average and extreme forcing conditions are important drivers of coastal morphodynamics, and fluctuations in shoreline positions are more sensitive to sediment availability.

The iconic Durdle Door (Fig. 2), in the World Heritage Jurassic Coast (Dorset, southwest England), is an arch formed by wave erosion in an almost vertical limestone. This limestone is more resistant to erosion than the soft chalk cliffs and other sedimentary formations that are presently seen landwards. The tectonic movements that formed the European Alps caused intense folding of the rocks forming this coast and hence the vertical limestone. The spatial variation in the position of the shoreline (and rate of erosion) is influenced by the complex geology and fluctuations in water levels, which

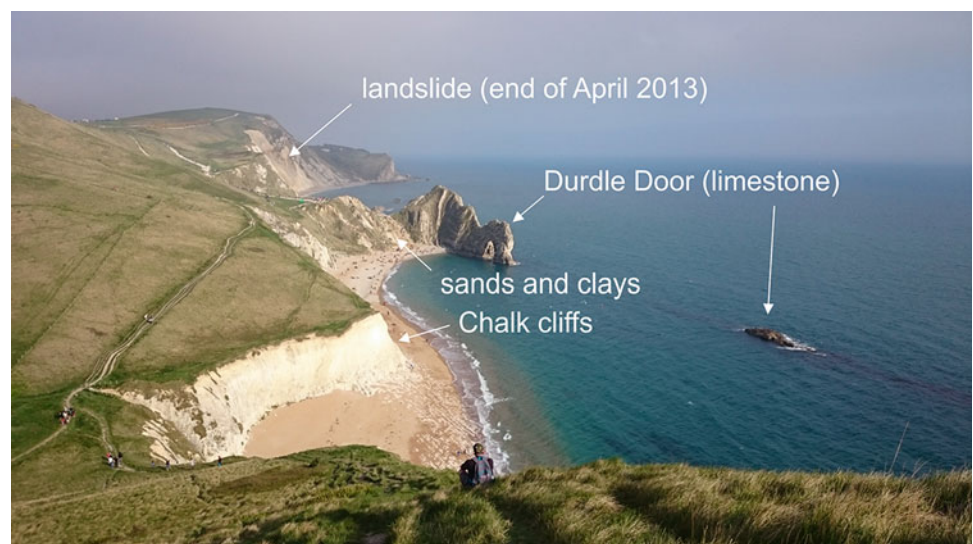
Erosion of Coastal Systems,

Fig. 1 Schematic representation of a coastal system showing its components (landforms, sediment dynamics, biota, and, increasingly more important, human interferences), long-term control settings (geology, sea level, and sediment availability), and forcing conditions



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Fig. 2 Section of the Jurassic Coast in Dorset (southwest England), where the position of the shoreline is influenced by the geology, with the chalk cliffs eroding faster than the limestone



control how often waves reach the cliff base. This eroding sedimentary coastline is prone to landslides, and cliff retreat is highly dependent on their frequency. Data from the British Geologic Survey landslide database show that the number of landslides increases after prolonged periods of heavy rainfall, particularly in southwest England (Pennington et al. 2015), where the Jurassic Coast is located. Although chalk does not supply beach-forming material, these cliffs have bands of flint nodules, which form the gravel found at the local beaches.

As a conditioning long-term factor, sediment availability is determined by a combination of geology, sea-level fluctuations (i.e., changes in physical accommodation), and exchanges in and out of the system, which in shorter-time scales vary as a function of forcing conditions. Most commonly, forcing conditions include a combination of wind, waves, tides, and currents, with their relative importance varying geographically. Coastal systems can be wave-dominated, tidal-dominated, river-dominated, or of mixed-energy depending on the type of process having the greater influence on sediment dynamics and landforms (Boyd et al. 1992).

Within a coastal system, erosion and deposition can occur concomitantly and vary spatially, in such a way that some subsystems may erode while others accrete. The system responds to changes in forcing conditions through morphodynamic adjustments (Wright and Thom 1977). The morphodynamic state of a coastal system is determined by the interactions taking place within the coastal system (how changes in one subsystem influence another) and the changes driven by external forcing conditions. The time that a system takes to reach equilibrium with the new conditions is called the relaxation time. Most studies are focused on explaining or quantifying erosion of particular subsystems (e.g., dunes or beaches) or within specific areas of a coastal system. Studies that are very detailed on a particular subsystem are called reductionist (Woodroffe 2003). Reductionist studies may identify and quantify beach erosion in a particular subsystem, while accretion or deposition in other subsystems may go undetected.

Landforms are shaped by a combination of (a) forcing conditions and the interactions they drive within the coastal system (i.e., the resulting hydrodynamics and sediment dynamics), (b) the presence of biota (i.e., the presence of vegetation or organisms that can change the susceptibility to erosion, transport, or deposition), and (c) human interference, which is increasingly pervasive affecting coastal systems' entities, attributes, and interactions both directly and indirectly (Fig. 1). In some situations and depending on the scale in which a coastal system is analyzed, it can be argued that human interference is also altering forcing conditions, therefore, playing a major role in determining the characteristics of the system. For example, tidal barrages can control tidal levels altering the tidal prism, current velocities, and duration of wet and dry conditions. Construction of dams

(sometimes many kilometers from the coast) can reduce the amount of sediment arriving at the coast, resulting in a sediment deficit that creates or exacerbates coastal erosion.

The Role of Feedback

Interactions between subsystems often relate to feedback mechanisms that are important to determine the dynamic state of a coastal system. These feedback mechanisms are evidences of the nonlinear and complex relationships characterizing coastal dynamics (Woodroffe 2003). Most commonly, interactions are self-regulating through negative feedback, i.e., the resulting morphological change will counteract its driving conditions inhibiting further changes. Negative feedbacks tend to maintain the coastal system in a state of equilibrium. A typical example of a negative feedback is the formation of nearshore bars as a result of beach erosion due to storm impact. The nearshore bar will cause waves to break, reducing the wave energy reaching the beach and, consequently, preventing further erosion. After the storm, as lower wave energy conditions return, the nearshore bar migrates landward and welds to the beach restoring sediment levels.

Positive feedbacks promote the continuation of change until a threshold triggering a different response is exceeded. An example of positive feedback is the formation of dune blowouts (Woodroffe 2003). Once the dune vegetation is removed in a certain area, it increases the exposure of sand to transport by wind, which will affect adjacent vegetation due to loss of substrate and damage by increased winds and/or smothering by mobile sand. Therefore, after the initial disturbance, blowout growth will continue due to vegetation degradation and increased exposure to aeolian transport. Positive and negative feedbacks have a key role in self-organization as a mechanism for changes in coastal morphology, as suggested for the formation of beach cusps. According to Coco et al. (2003), cusps' horns and bays develop through a positive feedback between the morphology and swash flow, which enhance the cusp relief up to a point when there is a decrease in differential erosion and negative feedback stabilizes the morphology.

Scales of Coastal Change

Erosion often occurs as part of natural cycles (working at a range of temporal scales) characterized by changing energy conditions that result in alternating periods of erosion and accretion. In most cases, erosion could be perceived as a temporary state, the erosional phase of a cycle, some of which lasting thousands of years, such as glacial and interglacial periods, while others may last few months, such as seasonal weather patterns determining the frequency and intensity of storms (Table 1). Considering the geologic (very long) time scale, dynamic processes with influence at the

coast tend to smooth out its configuration, as headlands are eroded (concentration of energy) and the sediment available is transported toward coastal embayments where it is deposited (dissipation of energy). It is due to tectonic uplifts, isostatic rebounds, and sea-level changes that coasts continue to be rejuvenated and have the diverse and remarkable landscapes that are so widely appreciated.

In this context, erosion is a natural process and of great importance for the dynamic equilibrium of coastal systems. It reflects the natural capacity of coastal systems to make morphological adjustments in response to changes in forcing conditions (i.e., energy and/or sediment inputs). However, in many locations, human activities have altered the natural environment, affecting the biota, coastal landforms, sediment dynamics, and their interactions (Fig. 1), in ways that have hampered the natural adaptive capacity of coastal systems. Most often, human interferences drive positive feedbacks that enhance change and push the system to exceed thresholds that can alter trends and response mechanisms. For example, revetments can reduce cliff retreat but create or aggravate sediment deficit, as cliff erosion will no longer supply material to the beach system. Additionally, wave reflection caused by the revetment can increase erosion resulting in lowering beach levels. In most cases, a reduction in sediment supply combined with an enhancement of sediment removal will increase the relaxation time or cause a shift in the system.

In the literature, scales of coastal changes have been described using a variety of terms that are not necessarily standardized or applied consistently (cf. Cowell and Thom 1994; Stive et al. 2002). Decadal changes may be considered as “short term” in a study of coastal evolution associated with sea-level fluctuations in the Quaternary and as “long term” in research assessing seasonal or interannual changes due to the impact of storms. Table 1 shows the terminology used to reflect the different temporal scales and examples of natural processes as suggested in Stive et al. (2002). Their original

table includes spatial scales, which are omitted here to avoid misinterpretation. Although it is possible to suggest the potential spatial scale in which the different natural processes may have an impact, the actual extent of such influence in any specific location will depend on the interactions with other processes acting at different scales and the characteristics of the coastal system.

Due to the widespread impact of human-induced processes on coastal changes, examples are included in Table 1. Contrasting with natural processes, which are commonly related to cycles, human-induced coastal change is most often episodic but causing long-lasting impacts on coastal systems. Human-induced impacts can alter the dynamics of coastal systems favoring erosion, transport, or deposition at levels that may take a considerable time for the system to recover or reach dynamic equilibrium under the new conditions. Therefore, in Table 1, the temporal scale associated with the human-induced processes should be interpreted as the likely duration of the impact or the time in which the system may reach dynamic equilibrium after disturbance (the relaxation time).

Causes of Coastal Erosion

There are several causes or factors acting at a range of temporal scales (Table 1) that can contribute to coastal erosion. Evidence of coastal erosion can be easily identified worldwide (Fig. 3) in the form of beach or dune scarp, cliff failure, fallen vegetation, and exhumation of relict deposits, just to mention a few examples. In very general terms, erosion will result from a reduction in sediment supply caused by (a) changes in forcing conditions (usually an increase in energy conditions and/or decrease in sediment input) or (b) natural or human-induced alterations within the coastal system that will result in shifts in the sediment dynamics (e.g.,

Erosion of Coastal Systems, Table 1 Temporal scales of coastal changes associated with natural (Adapted from Stive et al. 2002) and human-induced processes or factors (arrows indicate factors occurring across scales)

Scale	Temporal scale	Natural processes or factors	Human-induced processes or factors
Very long term	Centuries to millennia	Geological setting Long-term climate change ↑ Relative sea-level change Sediment availability	Anthropogenic climate change
Long term	Decades to centuries	↓ Regional climate variations Coastal inlet cycles ↑ Migration of sand waves Extreme events	Sediment retention by dams Stabilization of inlets ↑ Obstruction of littoral drift by hard engineering structures
Middle term	Years to decades	↓ Wave climate variations Surf zone bar cycles	↓ Degradation of coastal wetlands Land reclamation Removal of dunes
Short term	Hours to years	Seasonal weather patterns Oceanographic conditions	Managed realignment Beach nourishment



Erosion of Coastal Systems, Fig. 3 Some examples of coastal erosion evidence: (a) beach scarp in Oceanside Beach, (California, USA); (b) vegetation holding a backbeach scarp in Cayo Guillermo (Cuba); (c) gravel washover fans in Medmerry (south England) stripped sediment from beach exposing relict deposits; (d) erosion of soft cliffs in Arboletes (Antioquia, Colombia) where some families had to be relocated further inland; (e) dune scarp in Hermenegildo (southern

Brazil); (f) aeolian transport in an incipient blowout in the Sefton coast, one of the fastest eroding shores in the UK (Formby, northwest England); (g) exhumation of peat and lagoonal muds where the Conceição lighthouse collapse due to beach erosion in 1995 (southern Brazil); (h) a lighthouse close to the edge of a retreating cliff in Richards Bay (KwaZulu-Natal, South Africa); (i) rock platform along the Coogee to Bondi walk (Sydney, Australia)

loss of vegetation able to promote sediment stabilization). The most commonly cited causes of erosion include storms (mainly due to increased wave energy and water levels), relative sea-level rise, coastal engineering works, and other interferences related to human occupation. In most locations, erosion will result from a combination of these factors. The relative importance of each factor varies geographically and depends on the time scale of analysis. Tsunamis and other extreme or rare events not mentioned above may be important at specific locations.

Storms

Storms are events of high wave energy driven by strong winds, sometimes leading to storm surges and increased longshore sediment transport (when waves approach the coast at an angle). In storms, waves are higher (have more energy) and break across a wider surf zone, where they are able to put sediment into suspension. Once in suspension, particles are more easily transported by currents both cross-shore and alongshore. Higher waves result in the movement of larger particles and stronger longshore currents (the speed

will also depend on the angle of wave approach), which in turn can increase the volume of sediment being transported. Therefore, high-energy waves are the key factors enabling coastal erosion during storms. Considerable erosion can result from the impact of individual large storms, a cluster of storms, or a storm season (usually but not always the autumn or winter).

Not all coasts show the same susceptibility to storm impact. Rocky coasts will be less affected than sedimentary coasts, and these tend to be more resilient in areas where high-energy conditions are more common than in areas where stormy conditions are less frequent (Cooper et al. 2004). The magnitude of impact will depend on (a) the characteristics of the coast, including sediment type and availability, (b) whether the dynamic processes at hand are able to exceed the thresholds required for erosion to take place, (c) duration of the storm or stormy period, and (d) storm frequency (as this will determine whether there is time for recovery between storms). In low-lying coasts formed by unconsolidated sediment, this threshold is defined by the critical bed shear stress and particle size. Retreat of soft cliffs may occur when the cliff face is destabilized by removal of sediment at the cliff base (exceeding the angle of repose) or by overload due to water saturation (usually triggered by rainfall combined with wave action at the cliff base).

In coastal areas, strong winds can cause widespread and critical damage to infrastructure, sometimes causing fatalities associated with falling trees, broken electric cables, or flying debris. However, erosion due to aeolian transport may be restricted under certain conditions. Winds can be a direct cause of erosion of non-cohesive sediment (mainly sand) when surfaces are dry and vegetation is absent. Rainfall is an element of most storms, and higher water levels tend to “wet” a wider part of low-lying coasts, which precludes wind transport. However, winds play a key role in increasing wave heights and contributing to storm surges.

Storm surges are temporary high water levels exceeding the predicted astronomic tidal level due to a combination of wind setup (i.e., strong onshore winds force water toward the coast increasing water levels), wave setup, and lower atmospheric pressure. In general terms, the water level increases 1 cm for each one millibar drop in atmospheric pressure. In hurricanes, for example, the atmospheric pressure in the eye of the storm is considerably lower than outside the storm. Besides influencing water levels, this gradient determines the wind speeds. Atmospheric pressure as low as 909 millibars was recorded at the eye of hurricane Maria before the landfall in the Caribbean islands of Dominica as category 4 and Puerto Rico as category 5 (on 18 and 20 September 2017, respectively). Considering that the average atmospheric pressure at sea level is around 1013 millibars, a drop to 909 millibars would account for an increase of 104 cm in local water levels.

At landfall in Puerto Rico, the storm surge reached 180 cm, and sustained wind speeds were 155 mph.

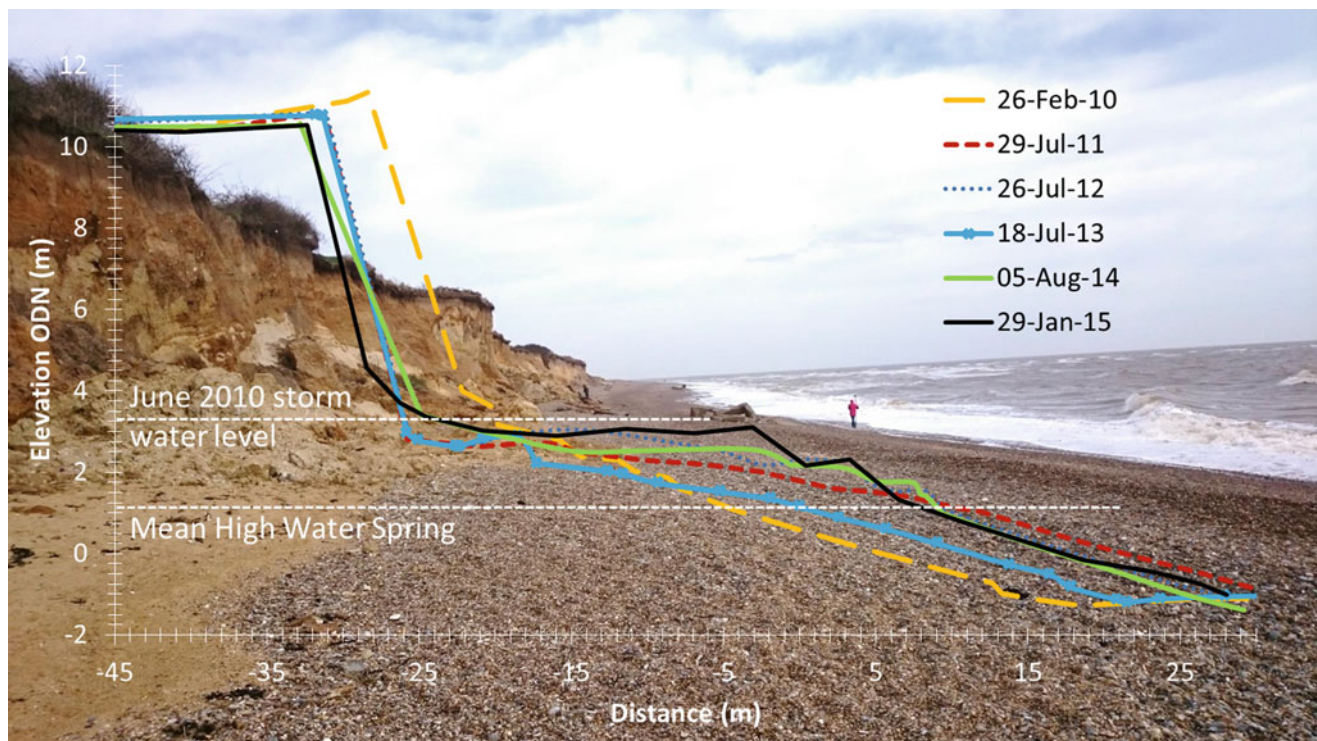
The main contribution of storm surges to coastal erosion is that the higher water levels allow waves (and associated currents) to reach areas that are not usually exposed to them. Therefore, waves will affect a larger area, including zones that may be less resilient to change and more prone to erosion. Dune scarping (Fig. 3e) can occur when high water levels enable wave erosion at the dune toe. In mesotidal or macrotidal locations, storm surges will have a greater impact if occurring around high tide; as at other times, it will only reach areas already within the tidal range. In urban areas, storm surges are a great flooding hazard due to overtopping of seawalls. Similarly, storm surges can facilitate beach erosion due to overwash, which occurs when water levels are high enough to transport sediment from the beach over the dunes or coastal ridges and deposited behind them as washover fans (Fig. 3c). Beach erosion through overwash contributes to coastal retreat and barrier rollover (typically triggered by rising sea levels).

Storm impacts on beaches are often described in terms of changes in beach profile, which can result from individual storms or from seasonal changes in wave energy. Profile changes are a result of cross-shore transport, in which high wave conditions move material from the beach to the near-shore. If water levels are high, waves can erode the dune toe or base of the cliff and supply material to the beach, as was observed in Thorpeness (Suffolk, east England) after the cliff base retreated around 5–6 m after a storm in June 2010 (Fig. 4). A similar change was observed after a prolonged stormy period during the 2013–2014 winter. It is important to note that the tidal range in Thorpeness is just over 2 m and the effect of storms on cliff retreat requires a storm surge to coincide with high tide.

Most often, storms are likely to erode material from the beach, lowering beach levels and supplying sediment to raise the seabed in the surf zone. The beach profile tends to become steeper than its pre-storm conditions and assume a concave shape. One or more nearshore bars can form, which, combined with the accreted seabed, will help dissipate wave energy and prevent further erosion. After the storm, constructive waves bring the sand back to the beach. When sediment deficit develops due to loss of material that was transported either cross-shore or longshore, the beach may take longer to recover after a storm or may not fully recover.

Climate Change and Relative Sea-Level Rise

Enhanced erosion and flooding of low-lying coasts are the most often cited impacts of climate change on coastal areas, which are expected to occur as a result of sea-level rise and enhanced storminess (e.g., FitzGerald et al. 2008). Due to increasing global average temperatures since the Last Glacial Maximum, sea levels have been rising 100–120 m in the last



Erosion of Coastal Systems, Fig. 4 Beach profiles measured just north of the Thorpeness village (Suffolk, east England), approximately at the location where they are overlain (photograph

date: 22 Apr 2017). Note that the first and last profiles were measured in the winter, all others in the summer. (Data provided by the Environment Agency)

20,000 years or so. Although there is much debate on the magnitudes and rates of change (e.g., Horton et al. 2014), there is little doubt in the scientific community that the global trend of sea-level rise exists and will continue throughout the twenty-first century. However, it is important to recognize that estimates of global averages might not be relevant at the local level, as the rate of sea-level change at any particular location is influenced by a range of local and regional factors (e.g., Woodroffe and Murray-Wallace 2012).

There is less certainty about the predictions of increasing frequency and intensity of storms due to inconclusive results from both models and analyses of historical data (e.g., Esteves et al. 2011). The uncertainty in the results of these studies is mainly related to constraints of data availability (e.g., times series of data are relatively short and restricted to few locations) and the high variability in meteorological and oceanographic conditions inherent to seasonal, interannual, and decadal cycles. Therefore, it is possible that the signal of enhanced storminess cannot yet be identified, but the likelihood of it happening should not be ignored. An increase in the frequency and intensity of storms could lead to greater mobilization of sediment, resulting in faster and/or more widespread coastal retreat. However, erosion can also release sediment to the coastal system, and this could either reduce erosion or promote accretion in some areas.

Sea-level rise driven by climate change occurs as a result of an increase in water volume in the world's oceans due to both thermal expansion of seawater and the melting of glaciers and ice sheets. At long and very-long temporal scales, sea-level fluctuations are a key factor determining the configuration of coastlines over hundreds of kilometers. For example, the ubiquitous presence of barrier islands along the east coast of the USA is a result from over 100 m rise in sea level in the last 20,000 years (since the Last Glacial Maximum). The melting of land-based ice during interglacial periods (i.e., periods of increasing temperatures) increases the volume of water in the oceans basins. Although this process has worldwide consequences, the effects will vary geographically.

For some time now and partly due to dissemination in the Intergovernmental Panel on Climate Change (IPCC) reports, there has been some emphasis on rates of sea-level rise based on global averages of data collected during the twentieth and twenty-first centuries. The widespread attention on estimates of global averages and associated future predictions contributes to a generalized misunderstanding about their relevance and applicability at the local level without consideration of spatial variations, including for planning purposes (Woodroffe and Murray-Wallace 2012). Although it may be fair to assume that rising sea levels will increase risk of coastal flooding and erosion, the actual coastal response at any site

will depend on a combination of factors with local or regional influence.

The magnitudes and rates of sea-level changes at any specific location usually refer to relative sea level, which is the net difference in sea level considering vertical changes in both the sea surface and land elevation (e.g., tectonics, isostasy, subsidence due to compaction). For example, the melting of ice sheets since the Last Glacial Maximum resulted in isostatic rebound (due to changes in the load of land-based ice) and hydro-isostatic changes (due to the load of the water released from the ice sheet), which still affect the relative sea level in Britain causing regional differences. Isostatic rebound is causing land uplift of more than 1 mm a^{-1} in southeast Scotland and land subsidence of over 1 mm a^{-1} in southwest England (Woodworth et al. 2009). In some locations, uplift due to tectonics or isostatic rebound may balance out or reduce rates of sea-level rise, but other factors can influence the relative sea level, such as the changes in gravitational pull of ice sheets and glacier on adjacent waters (Hsu and Velicogna 2017).

Rising sea levels usually lead to shoreline retreat due to marine transgression (i.e., landward displacement of the shoreline) and associated coastal flooding and erosion. If sediment dynamics could be ignored, the horizontal distance of shoreline retreat could be calculated by knowing the change in water-level elevation and the average slope of the coastline. For example, if the slope is about 1.9° , it has a gradient of 1:30 (i.e., one meter change in elevation every 30 m in horizontal distance). Therefore, 1 m rise in sea level would move the shoreline 30 m inland of its current position. However, this is an oversimplification as it disregards any changes in morphology and sediment movement associated with sea-level fluctuations. The actual shoreline displacement will depend on the rate of sea-level change, the availability of sediment, and the morphological changes that will take place.

Marine transgression occurs when the sea-level rise outpaces the accumulation of sediment, increasing accommodation space. Transgression is a key factor of barrier island (or coastal barrier) rollover, which occurs as overwash events become more frequent due to increasing water levels. Barrier rollover tends to steepen the beach face and create gentler slopes in the back-barrier environments as a result of material being eroded from the former and deposited in the latter. An evidence of coastal rollover during transgression is the exhumation of peat and lagoonal muds at the beach face (Fig. 3g). Peat and lagoonal muds form in relatively sheltered conditions (e.g., coastal embayments or back-barrier environments) where halophytic vegetation (e.g., saltmarshes) develops and fine sediments accumulate. Washover fans (Fig. 3c) are usually found in the upper beach or behind dunes, an indication of the material that was removed from the beach resulting in the exposure of the relict deposits.

The morphological change that will result from sea-level rise and the rate of change depends on processes acting at a range of scales (FitzGerald et al. 2008). Barrier islands may move inland tens of kilometers over thousands of years, but this process depends on successive overwash events to continue acting (each individually at much shorter scales) over the long time frames of barrier rollover. Barrier breaching and segmentation are also expected to occur due to sediment loss and impact of major storms, and barriers sometimes disappear due to subsidence.

In a transgression, coastal environments migrate inland to occupy their preferred position within the tidal range (Fig. 5a, b). As saltwater penetrates further inland, coastal wetlands, such as mangroves and saltmarshes, will start colonizing new areas, substituting grasslands and other vegetation not adapted to saline waters. If sediment is available, coastal wetlands are able to keep up with moderate rates of sea-level rise through vertical accretion. Where there is a sediment deficit, sea-level rise will outpace vertical accretion, and intertidal habitats may be lost as they become increasingly or permanently submerged. Some species occupy specific positions within the tidal range and may be lost due to increased duration and frequency of inundation. A similar effect will occur if further inland, topography is higher or gradient is steeper; in these cases, the horizontal distance between the high and low water line will reduce through time, resulting in smaller intertidal areas (Fig. 5c). The loss of intertidal habitats may result in erosion due to the direct release of sediment previously retained by vegetation and a reduction in the sheltering effect caused by the vegetation.

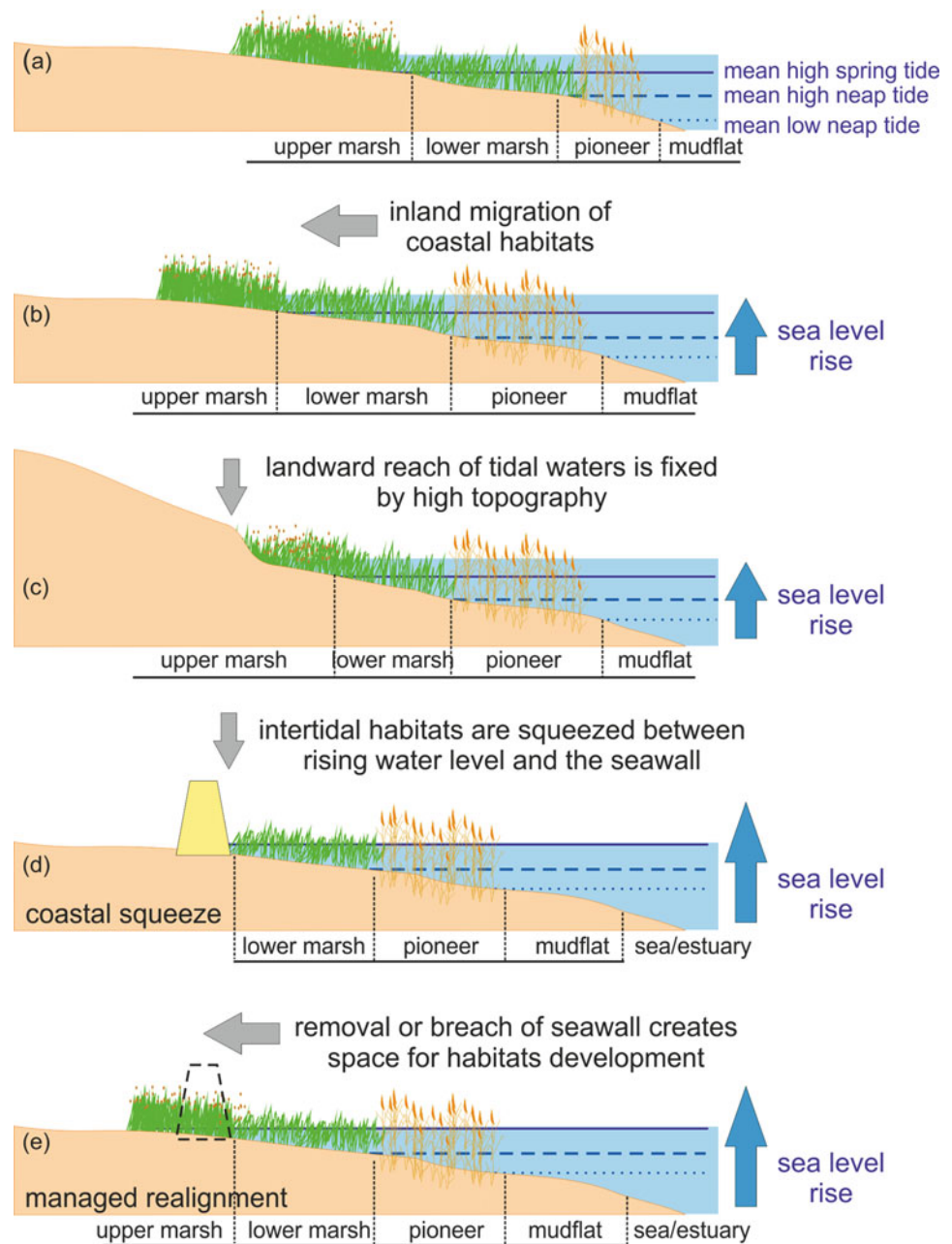
Coastal retreat or marine transgression may not be observed everywhere, even when sea level is rising. Marine regression (i.e., seaward displacement of the shoreline) will occur in areas where the rate of sediment accumulation outpaces the rate of sea-level rise (i.e., reducing the accommodation space). Increases in sediment supply arriving at the coast may be a consequence of sea-level rise or other factors. For example, sea-level rise may (a) change the base level of rivers, triggering an increase in fluvial sediment input to the coast, or (b) expose alluvium deposits to wave erosion creating a source of material to adjacent areas.

Human-Induced Coastal Erosion

Human activities have altered landscapes and the balance of natural processes in many ways and locations. Coastal erosion may result from activities taking place at the coast or further inland, causing cascading effects across temporal and spatial scales. These effects are so widespread and have been in place for so long that it may be difficult to separate the human and the natural contribution to coastal erosion. This differentiation is a complex task due to two main reasons: (1) coastal change results from the interaction of many factors which are naturally highly dynamic; and (2) our understanding of these

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Fig. 5 (a) The distribution of some coastal habitats (e.g., saltmarshes) is dependent on the elevation in relation to the tidal levels. (b) Seeking to occupy optimal elevations within the tidal range, coastal habitats tend to migrate inland as a response to sea-level rise. A reduction in intertidal area may occur when the landward migration is prevented by (c) high topography or (d) the presence of coastal engineering structures. (e) Managed realignment increases intertidal areas allowing for the development of intertidal habitats by artificially breaching or removing of coastal defenses. (Adapted from Esteves 2014)



interactions is impaired due to poor data availability (e.g., short-time series, limited spatial and temporal resolutions, poor geographical coverage). Often, the ultimate cause of human-induced coastal erosion relates to direct and indirect impacts on sediment supply and distribution. However, in some cases, the key impact may be due to changes in relative water levels (e.g., land subsidence). The intension here is to provide just a few examples that illustrate the widespread human contribution to coastal erosion across a range of scales.

Erosion in coastal areas adjacent to river mouths is often linked to reduction of sediment reaching the coast due to retention within reservoirs built along the river course. This is probably the anthropogenic interference that alone creates

the largest reduction of sediment supply to coasts worldwide. A global study using a combination of measurements and modeling estimates that (a) dams built in the past five decades are retaining over 100 billion metric tons of sediment, (b) about 26% of the total sediment flux is retained in reservoirs, and (c) this retention results in a net reduction of 1.4 ± 0.3 billion metric tons per year in the flow of sediment reaching the world's coasts when compared to prehuman times (Syvitski et al. 2005). These results include the increase in fluvial sediment load due to deforestation and poor soil management practices in catchment areas. In river-dominated deltas, retention of sediment in dams can greatly enhance the problem of land subsidence and local effects

of hard engineering structures, as reported in the Nile delta (e.g., El Banna and Frihy 2009). River diversion, water abstraction, and other activities that reduce sediment supply to the coast will also contribute to coastal erosion.

Extraction of sand and gravel from the beach or nearshore for construction works (including beach nourishment) can cause a direct impact in local sediment availability and an indirect impact by increasing the wave energy reaching the coast (due to increasing depth). At present, in most locations, such activities are regulated and would not be allowed if there is a risk of detrimental impact (usually identified through environmental impact assessment). For example, in the Philippines, the government act Batas Pambansa Blg. 265 approved in November 1982 prohibits the extraction of sand and gravel from beaches to prevent erosion or impact on their “natural beauty.” In the island of Jersey (in the English Channel), a similar law passed in 1963 prohibiting extractions of sediment (from clay to gravel) from beaches and rocks forming the coast, except under license from the government. However, in some locations, the impact exists either because these regulations have not yet been implemented or arrived

too late or illegal extraction still takes place. More information on sediment mining impacts in coastal systems can be found in the Coastal Care website (Sand mining, <http://coastalcare.org/sections/inform/sand-mining/>, last accessed on 18 Jan 2018.).

Stabilization of cliffs and bluffs aims to slow down cliff retreat but can lead to beach erosion due to a reduction in sediment supply. Revetments and seawalls are commonly used to protect the base of cliffs and bluffs from wave action. Structures parallel to the coast, such as revetments and seawalls, are known to cause beach lowering due to scouring and removal of sediment due to wave reflection. This effect occurs when the water level allows waves to reach the structures and becomes more critical if occurring frequently. The integrity of revetments or seawalls can be threatened if sediment deficit results in continued beach lowering. Therefore, by preventing cliff erosion, these structures create a positive feedback that enhances beach erosion and, ultimately, requires the implementation of a number of hard engineering structures with mixed results. In Suffolk (east England), a myriad of coastal protection works (Fig. 6) have been implemented over the last



Erosion of Coastal Systems, Fig. 6 Examples of cliff stabilization along the Suffolk coast (east England). In the village of Thorpeness, (a) a revetment from the 1970s was exposed by a storm in the summer of 2010, leading to the addition of (b) geobags that were also exposed by another storm in 2013; enhanced cliff erosion is observed north of the revetments (where profiles shown in Fig. 4 were measured). (c) Seawall,

rock revetment, and fishtail groyne (which replaced old wooden groyne) after increased cliff failure in 2013 in Hopton, where the beach no longer exists at high tide. (d) In Corton, 2 km south of Hopton, since the 1890s, different coastal protection works have been built in an attempt to slow down cliff retreat

100 years in an attempt to halt retreat of the soft cliffs with little avail. In addition to the reduction in sediment supply resulting from cliff protection, there is a general perception in the local communities that other human activities have accelerated erosion. For example, erosion at Hopton (Fig. 6c) is said to have worsened due to works at the Great Yarmouth harbor, 3 km to the north. In 2013, cliff failures at Hopton resulted in the construction of fishtail groynes, which in turn created fears of increased erosion 2 km to the south at Corton (Fig. 6d), where local residents are pledging for similar structures to be built along their coastline.

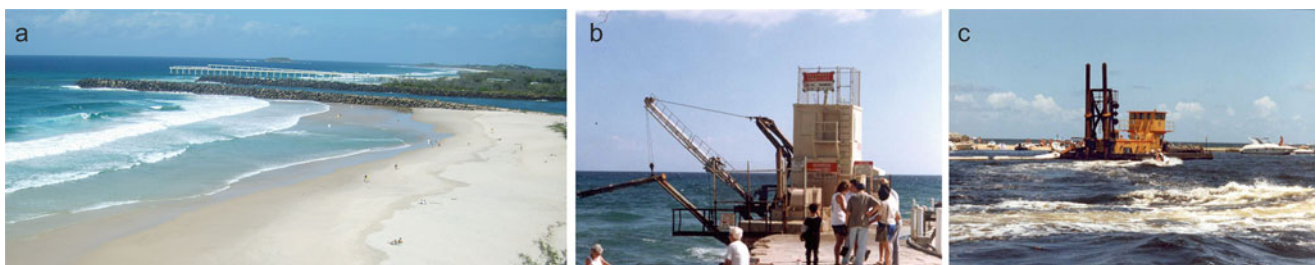
Besides lowering beach levels, stabilization of coastlines by hard engineering structures can result in coastal squeeze. Coastal squeeze is the loss of intertidal habitats due to rising sea levels along coastlines fixed by hard engineering structures (Fig. 5d). As explained previously, intertidal habitats tend to migrate inland following a rise in sea level (Fig. 5b). Similar to the effect of raised topography (Fig. 5c), engineering structures can prevent this migration and reduce the area for intertidal habitats to develop, resulting in coastal erosion, particularly due to vegetation loss. Managed realignment involves breaching or removal of coastal structures allowing tidal inflow into previously embanked land increasing space for intertidal habitats to develop (Fig. 5e). Managed realignment is increasingly being implemented in Europe as a more sustainable approach to achieve both environmental and coastal protection enhancements (Esteves 2014). In some cases, sea-level rise can facilitate the development of saltmarshes within embanked land, without active intervention (i.e., passive habitat restoration). In south Portugal, salt-water intrusion is enabling the colonization of saltmarshes in embanked reclaimed land where agricultural practices have been abandoned since the mid-1960s (Almeida et al. 2014).

Obstruction of the littoral drift is a well-known and widespread effect of coastal engineering structures extending into the surf zone, such as groynes and jetties. The sediment transported by longshore currents is retained in the updrift side of these structures, causing a sediment deficit and erosion downdrift. Inlets are stabilized to improve navigation safety

and, paradoxically, reduce erosion caused by inlet migration. Groins are built either to redistribute sediment, forcing accumulation to occur where desired, or to reduce the rate of erosion due to longshore transport (e.g., prolonging the lifetime of beach nourishment). Inlet stabilization and maintenance of navigation channels bring enormous economic benefits (transport of goods and people, tourism). Mitigating measures can reduce but do not eliminate the impact of these structures on sediment budget. The Tweed River Entrance Sand Bypassing Project (Tweed Sand Bypassing, <http://www.tweedsandbypass.nsw.gov.au/>, last accessed on 18 Jan 2018.) (Australia) is an example of an inlet management plan based on sand bypassing (Fig. 7a) that has been reasonably successful in maintaining navigation safety and restoring the amenity of beaches to the north along the southern Gold Coast. Sand bypassing involves moving sand from accreted beaches to nourish eroding beaches.

In Florida (USA), it was estimated that 72% of eroding beaches are adjacent to tidal inlets (Esteves and Finkl 1998), and near inlets (within 3.6 km on average) magnitudes of shoreline change tend to increase (Absalonsen and Dean 2011). Florida has a large number of inlets; most are stabilized by jetties. As a result from sand being retained updrift of jetties and lost to offshore areas due to dredging of navigation channels, a critical sediment deficit developed, especially in southeast Florida. The need for sand bypassing became crucial to manage beach erosion. In 1986 Florida legislation recognized the issue and passed the “Declaration of public policy relating to improved navigation inlets” (Section 161.142), which established that beach-quality sand should be bypassed to downdrift beaches on an annual basis in volumes equivalent to the net annual longshore sediment transport. The legislation has stimulated the implementation of sand bypassing measures (Fig. 7b, c) as part of inlet management plans, but compliance has been an issue.

In the past, coastal erosion induced by human activities was a result of poor understanding of both the environmental impacts these activities would cause and the temporal and spatial scales of coastal change. Knowledge on these two



Erosion of Coastal Systems, Fig. 7 Examples of sand bypassing systems. (a) Tweed River Entrance Sand Bypassing Project (Australia) uses underwater pumps to move sand from the accreting beach south of the inlet to the beaches further north in southern Gold Coast. (b) This fixed dredge at the South Lake Worth (Palm Beach County, Florida)

bypassing sand accumulated north of the inlet to the eroding beach in the south since 1937. (c) The resident floating dredge at Hillsboro Inlet (Broward County, Florida) bypasses sand from a sand trap inside the inlet to nourish the eroding beaches north of the inlet; a weir jetty (seen beyond the dredge) also contributes to sand bypassing

fronts has greatly improved, and many activities known to have detrimental impact on coasts are now being prohibited or reversed. Examples are varied worldwide and include bans of new hard engineering structures, removal of coastal protection works, and removal of dams. Many projects now exist that prioritize the renaturalization of coastal areas, rivers, and floodplains (e.g., through managed realignment) to improve coastal erosion and flood risk management with added environmental benefits. The twentieth century (and earlier in some locations) can be described as the era of engineered coastlines, which resulted in extensive interference in the natural functioning of coastal systems. The realization about the level of impact of such interferences has gradually led to shifts in management practices, which may see the twenty-first century to become the start of a de-engineering era aiming to restore the dynamic nature of coastlines.

Summary

Coastal erosion is both a natural process important for the maintenance of dynamic equilibrium of coastal systems and a major hazard to human occupation along dynamic shorelines. Most often, it results from a combination of natural and human-induced factors acting at a range of scales. The importance of each specific factor will vary geographically and will depend on the temporal scale in which the system is being analyzed. Considering natural processes, erosion is often a temporary state associated with phases of natural cycles that can vary from short (e.g., seasonal increase in wave energy) to very long durations (e.g., sea-level rise during interglacial periods). Poor understanding about the duration of these cycles or about the spatial scale of associated coastal changes has led to human interferences that often result in enhanced erosion. Human-induced erosion is becoming pervasive worldwide both due to *in situ* direct impacts (e.g., due to coastal engineering works) and indirect impacts of inland activities (e.g., retention of sediment by dams). In many locations, climate change may result in increasing storminess and sea-level rise, which are likely to enhance existing coastal erosion or trigger a shift toward erosion. This is a particular concern where urban areas or coastal infrastructure is occupying dynamic spaces (areas where natural coastal change is expected) in low-lying coasts. There is a growing trend toward renaturalization of coastal areas (e.g., managed realignment) as a more sustainable approach to manage flooding and erosion risks with added environmental benefits. However, a key challenge to coastal management remains the urgent need to ensure that existing issues are not made worse by ill-informed policies or management decisions driven by short-term gains (e.g., allowing occupation in erosion-prone areas or persisting with hardening of shorelines where other options, including retreat, are more sustainable albeit more complex).

Cross-References

- Beach Erosion
- Coastal Dynamics
- Coastal Erosion
- Coastal Erosion Management
- Coastline Changes
- Human Impact on Coasts

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Erosion Processes

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Erosional processes along coastlines include: (1) the direct effects of hydraulic action, wedging, and cavitation by waves; (2) abrasion (corrasion), using sand, gravel, and larger rock fragments as tools; (3) attrition of the rock particles themselves during this abrasive action; (4) salt weathering or fretting; (5) erosion by organisms (bioerosion); and (6) chemical attack, or corrosion, which weakens the rocks and accelerates erosion. The rates of erosion by these processes are a function of the exposure of a coast to wave attack (wave energy and length of time of exposure), and the resistance of the materials to erosion and weathering.

Waves are the most dominant force causing coastal erosion. Ocean waves typically break at depths that range from about 1 to 1.5 times wave height. Because waves are seldom more than 6 m high, the depth of vigorous erosion by surf is usually limited to from 6 to 9 m below sea level. This theoretical limit is confirmed by observations of breakwaters and other coastal structures that are only rarely affected by surf to depths of more than 7 m. The change in water level caused by tides, however, allows waves to attack a greater range of elevations on the coast.

Wave impact directly exerts great pressure and can be a highly effective agent of erosion. The pressures produced by breaking waves have been measured by dynamometers, with reported values as high as 250 kg/m^2 ($6,000 \text{ lb/ft}^2$). These measurements agree with theoretical estimates (Gaillard 1904) based on wavelength and celerity. When the waves are long and their celerities are correspondingly high, the greatest pressures are generated.

When waves break against a cliff face, water is driven into cracks and joints and the air within the cracks is greatly compressed. The compressed air acts as a wedge, widening the fractures and loosening blocks of rock. As the wave recedes, the compressed air expands with explosive force (a process called cavitation), effectively quarrying out large blocks (Barnes 1956). Such quarrying action has provided enough energy to move blocks weighing many tons. There are many reports of the movement of large blocks by wave action, including a 2,500-ton piece of concrete at Wick (Scotland), and a section of a breakwater weighing 1,700 tons that was completely overturned at Bilbao (Spain).

Another method of marine erosion is abrasion, the mechanical wearing and grinding away of rock surfaces by the friction and impact of rock particles. Also called corrasion, this process involves the action of sand, gravel, and large

rocks picked up by the waves and driven back and forth against the exposed shore. Potholes can develop where large clasts are rotated by swirling water in the surf or breaker zone. The abrasion process includes dragging (friction), rolling, bouncing (saltation), and the artillery action of large fragments, which can shatter solid rock when hurled against a cliff. At Tillamook Rock on the Oregon coast, for example, rocks weighing as much as 297 kg have been picked up during severe storms and thrown through the windows of the light-house, 44 m above mean high water. Similarly, the windows of the Dunnet Head Lighthouse on the north coast of Scotland, more than 100 m above the high water mark, have been broken by rocks propelled by wave action.

Concentration of wave attack at the cliff base accelerates erosion and notches the base, permitting collapse or movement of the overlying material (Sunamura 1992). Seepage of ground-water (especially along seams and fractures), frost action, and wind combine with the under-cutting to produce cliff recession (Nott 1990). Differences in the resistance of the rocks being eroded and episodic collapse may produce an irregular cliff face, subject to mass movements (Duperret et al. 2002; Benumof et al. 2000). The debris produced by cliff collapse is subsequently used in abrasion of the cliff base, and/or removed by wave action, allowing the wave attack to continue. Thus, cliff erosion is generated by two processes: notching at the base of the cliff by marine processes, and collapse and denudation of the entire cliff face by a combination of subaerial and marine processes (Duperret et al. 2002).

The rate of cliff retreat, combined with rock type, climate (especially rainfall) and vegetation will determine the nature of mass movement down the cliff face. An extensive examination of Japanese cliff erosion, found that long-term cliff retreat is logarithmically related to rock strength and unrelated to cliff height (Sunamura 1977, 1992). Benumof et al. (2000) also found that rock strength, rather than wave energy or slope, is the critical factor in coastal cliff retreat.

Mathematically, the major factors of basal erosion have been summarized in Sunamura's equation:

$$x = f(F_w, S_r, t),$$

where x is the basal cliff erosion rate, F_w is the wave induced force, S_r is the cliff material resistance, and t is the time. Belov et al. (1999) developed a mathematical erosion function (the so-called Belov–Davies–Williams (BDW) equation) which accounts for erosion forces and the strength of the rocks, and also for the exponential decrease in wave erosion intensity with height. Numerical models have been quite successful at simulating basal cliff erosion (Belov et al. 1999).

The level of planation of the cliff base will usually lie immediately above the level of permanent rock saturation, where continued wetting and drying weakens the rock, allowing accelerated wave erosion. This level will, however, vary depending

on wave energy, rock type and structure, platform width, and accompanying atmospheric processes (Stephenson and Kirk 2000). The abrasion along the shore works horizontally to form a wave-cut cliff and a wave-cut platform. Exposure may be an important factor, as studies of shore platforms in Japan have shown that the mean platform width is about 40 m on the sheltered Inland Sea, 50 m on the more exposed Japanese Sea, and 60 m on the exposed Pacific coast (Trenhaile 1997).

Attrition is the process of wearing down by friction, involving the abrasion, impact, and grinding together of clasts in motion, resulting in smaller and usually better-rounded particles (in Europe, abrasion is used as a synonym for attrition). The impact and rubbing together of the rock fragments during wave action, as well as the grinding of the blocks that fall as the cliff is undercut, cause the tools themselves to wear down to smaller sizes and become rounded. Studies of attrition of beach clasts include experiments and observations in natural settings. Kuenen (1964) experimentally correlated rate of attrition with relative resistance and median clast size. Early comparisons of natural sediment distributions were done by Landon (1930), but lack of control makes the conclusions questionable. The observed diminution of grain size along a coast could be also due to change in source material or to sorting by longshore transport.

Salt weathering or fretting takes place when salt crystals grow within pore spaces in rocks; the crystal growth expands the pore and may lead to fracturing of the rock (Wellman and Wilson 1965; Mottershead 1989). This may be the result of strong evaporation that draws salt solutions to the surface by capillary action. The salt-weathering process is most effective in areas exposed to salt spray and subject to alternative wetting and drying. Salt weathering can also produce a honeycomb pattern in the rocks called tafoni, which is most characteristic of tropical and subtropical semiarid to moderately arid climates. In chalk cliffs, basal notching has been attributed to salt weathering and repeated wetting and drying cycles (Duperret et al. 2002).

The chemical and solvent action of seawater may be especially important along limestone and chalk coasts. On some atolls, solution may be responsible for 10% of total erosion of the coast (Trenhaile 1997). However, even non-calcareous rocks may weather more rapidly as a result of the chemical action of seawater. In some cases, chemical dissolution and salt fretting may act together.

Additional factors that may contribute to mechanical weathering in certain areas include large diurnal temperature variations and biological activity. Weathering due to temperature variations mainly affects the surface layers. The temperature distribution in rocks is non-uniform, and differences in rock composition yield thermal conductivity contrasts and large temperature gradients. This can cause thermal cracking of blocks, resulting in displacement along existing joints and fractures.

Biological erosion is especially important in tropical or subtropical areas having limestone coasts. Algae are probably the most important bioerosional agent on rocky coasts. The substrate under algal mats is exposed to the products of metabolism and organic waste, which can directly etch the rocks. Algae also support grazing organisms such as gastropods and echinoids, which abrade rock surfaces. Bioerosion, combined with the chemical solution of limestone by algae, produces the intricate scalloping known as phytokarst (Bromley 1978). Boring and browsing organisms erode rocks most effectively in the inter-tidal zone, and bioerosion can eventually weaken the framework of reefs to the point where collapse takes place. Rates of bioerosion on horizontal and vertical limestone surfaces have been estimated at about 0.5–1 mm per year (Trenhaile 1997).

In cold regions, where unconsolidated but frozen, ice-bonded sediments are exposed to wave erosion, convective heat transfer between seawater and the melting surface of the frozen cliff sediment (thermal abrasion) combines with erosive wave action and slope failure to under-cut cliffed coasts (Kobayashi et al. 1999). Wave erosion removes the unfrozen sediment, which would insulate the frozen sediment from further melting. The combined thermal and mechanical processes can lead to rapid erosional retreat (>1 m per year) during periods of open-water conditions. Tidally induced frost action occurs in the inter-tidal zone where rocks freeze when exposed to air and thaw when covered by the rising tide. This mechanism is particularly effective because of the frequency of freeze–thaw cycles. In coastal Maine, for example, there are an average of 133 freeze–thaw cycles each year in the inter-tidal zone versus only 30 or 40 cycles inland (Trenhaile 1997).

Cross-References

- [Bioerosion](#)
- [Cliffs, Erosion Rates](#)
- [Cliffs, Lithology versus Erosion Rates](#)
- [Ice-Bordered Coasts](#)
- [Notches](#)
- [Rock Coast Processes](#)
- [Shore Platforms](#)
- [Wave-Dominated Coasts](#)
- [Waves](#)
- [Weathering in the Coastal zone](#)

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Erosion: Historical Analysis and Forecasting

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Definition

Historical erosion rates are determined by monitoring the location of a representative shoreline reference feature, along a prescribed shore-parallel distance, over a specified time frame. Shoreline reference features, such as the high water line (HWL), are digitized from historical maps and aerial photographs, and combined with more recent shoreline data obtained from sources such as Light Detection and Ranging (LIDAR) and Global Positioning Systems (GPS) surveys. In the USA, the use of historical National Ocean

Service (NOS) T-sheets often allow erosion rates to be determined over a period of 150 or more years. Long-term erosion rates are valuable for coastal land use planning in that they can be used to predict coastal areas susceptible to future erosion.

Introduction

The study of long-term coastal erosion has taken on increased importance as development along the world's coasts has risen dramatically over the past several decades. This is readily apparent, for example, in the United States where development has transformed many small beach villages into moderately to densely populated cities. This has resulted in increasing population at risk from coastal hazards, such as hurricanes, nor'easters, and tsunamis, and associated damages from wind, coastal flooding, and erosion. Almost 30% of the US population lives in "hurricane prone regions" (Crowell et al. 2010) as defined by the American Society of Civil Engineers (ASCE 2006). In addition, almost 3% of the US population lives in ocean and Great Lakes coastal flood hazard areas defined by the 1% annual chance flood (Crowell et al. 2013). A subset of this coastal flood-prone population is also subject to the risk of long-term erosion.

Erosion rates (The phrase "erosion rate" is used throughout this article as an equivalent to the more precise "recession rate." Long-term beach recession is caused by a combination of long-term "static" inundation from rising sea levels combined with long-term "dynamic" erosion) are calculated by monitoring the location of a representative shoreline reference feature or proxy, usually the high water line or bluff line, over a specified time frame. Rates are obtained by measuring the location of two or more shorelines from historical shoreline change maps. These maps are produced by digitizing shoreline change reference features from historical maps and aerial photographs, and combining and registering these data with more recent shoreline data from LIDAR, and GPS surveys. Historical shoreline change maps and datasets for the USA often contain four to eight or more shorelines and can span 150 or more years. Erosion rates are typically calculated from the maps by generating a line perpendicular to the multiple shorelines and measuring the amount of movement over a period of time defined (and constrained) by the dates of the digital shorelines.

Sources of Historical and Current Shoreline Location Data

There are four principal data sources used in the production of historical shoreline change data: historical maps, recent and

historical aerial photography, LIDAR surveys, and GPS surveys. Following are brief descriptions of these sources.

Historical Maps. In the United States, the primary historical map data source is NOS T-sheets. The first T-sheets date back to 1835, but maps of 1835 to 1845 vintage sometimes show inaccuracies in longitude. In 1846, NOS surveyors began using the telegraph as part of field mapping which significantly improved longitudinal accuracy to the point that maps of post-1846 vintage are typically accurate enough for use in modern historical shoreline mapping studies. NOS has periodically remapped coastal areas every 20–60 years. As such, two or three pre-1940s and one to two post-1940s T-sheets are generally available for most of the US East and Gulf Coasts (Anders and Byrnes 1991).

NOS T-sheets are an invaluable data source that expands the temporal range of the historical shoreline data from about 75 years (using only aerial photography, and/or GPS ground surveys and LIDAR) to 150 or more years. A major prerequisite, however, is that each map be rectified and corrected for error factors such as scale differences, rotation, and uneven map shrinkage. Map scale is obviously a major consideration – 1:10,000 or larger is preferred, although maps at scales as small as 1:20,000 are often usable (Anders and Byrnes 1991; Moore 2000).

Aerial Photographs. Vertical aerial photographs date back to the 1920s; however, aerial photography produced prior to the late-1930s is typically overly distorted and of insufficient resolution for use in historical shoreline mapping studies (see below). Moreover, it was not until after World War II that air photos on the US coasts were taken with any degree of regularity. Many of these sets of aerial photography were taken to document storm impact and, as is discussed later, are of questionable use for determining long-term rates of beach erosion (Douglas et al. 1998). Nonetheless, several sets of post-World War II aerial photographs suitable for historical shoreline mapping studies are often available for a given area.

GPS Surveys. Field-based surveys using GPS have been increasingly used within the past few decades (Morton et al. 1993; Hapke et al. 2011). Shoreline position data are collected by driving a GPS-equipped off-road vehicle directly along a shoreline reference feature, such as the HWL. Geo-coordinates are collected at specified intervals, enabling real-time digitization of the shoreline reference feature.

LIDAR Surveys. LIDAR technology is now routinely used to conduct beach mapping, including the updating of historical shoreline change data (Robertson et al. 2004). This technology is being extensively used by the United States Geological Survey (USGS) in their national shoreline mapping project (<https://marine.usgs.gov/coastalchangehazards/research/long-term-change.html>). The LIDAR sensing instruments only collect elevation data, which can be related to the mean high-water (MHW) shoreline (Hapke et al. 2011).

These aircraft-acquired data are of very high resolution with a vertical precision of 15 cm or better as verified by GPS kinematic beach surveys (Middleton et al. 2013). A large stretch of beach can be imaged in a few hours, providing coastal scientists and resource managers the capability of producing high-resolution beach maps much faster and less expensively than traditional beach mapping (<http://geodata.lib.ncsu.edu/fedgov/noaa/commvuln/htm/intro.htm>).

Special Note on Drone Technology. The advent of drone technology is revolutionary in its capability to conduct mapping using a high-resolution video-camera with still-frame capability. LIDAR units are now so light and small that they can be carried aloft by a quadcopter-type drone (<https://www.dronezon.com/learn-about-drones-quadcopters/best-LIDAR-sensors-for-drones-great-uses-for-LIDAR-sensors/#>). Measurements extending several kilometers can now be undertaken, and the range can be increased by using very short laser pulses in the invisible near-infrared; this enables higher laser power compared to continuous wave lasers.

Shoreline Change Reference Features (Proxies)

A number of potential shoreline reference features or proxies can be used to monitor historical shoreline changes (see, for example, Boak and Turner 2005). In many situations, the HWL has been shown to be the best indicator of the land-water interface for long-term historical shoreline comparison studies, especially those that incorporate NOS T-Sheets, as the HWL was the boundary line mapped by early NOS topographers. The HWL is usually easily recognizable in the field, it can often be approximated from air photos (Stafford 1971), and it represents a shoreline feature that has been mapped (albeit sporadically) since the mid-1800s. There are several factors, however, that must be considered or addressed when using the HWL as the representative shoreline proxy. For example:

- Variability of the HWL location must be considered. As an example, Dolan et al. (1980) monitored the short-term variability of the location of the HWL during a tidal cycle in NC and VA. He found that the position of the HWL could vary anywhere from 0.12 m to 5.8 m.
- Correct identification of the HWL is critical. Pajak and Leatherman (2002) found that other shoreline indicators, such as the swash terminus, can easily be mistaken for the HWL on poor-quality aerial photographs.
- LIDAR surveys can present inconsistent comparisons of shoreline reference features, as these datum-derived MHW shorelines are typically offset seaward from the HWL (Morton et al. 2004).
- The HWL is not always the most representative proxy for long-term shoreline movement, particularly in areas with

gently sloping beaches or where rates of erosion are low (Morton et al. 1993). In these and other situations, Morton notes that other morphological features, such as erosion scarp, crest of washover terrace, or vegetation line are more diagnostic of long-term change than the HWL. Use of these other indicators, however, severely limits the temporal span of source data.

For cliffed situations, the top edge of the bluff (employing the terms cliff and bluff interchangeably) is commonly used as a shoreline change proxy. Unfortunately, historical maps depicting accurate bluff positions are rare, thereby limiting historical source data to the temporal span covered by aerial photographs, and GPS and LIDAR surveys (e.g., 1930s to present). Other features are sometimes used to monitor historical bluff change. The State of Wisconsin, for example, has experimented with the mid-bluff contour, with the rationale that the middle of the bluff is typically the most stable portion of the slope because it is located below areas subject to frequent failure and above the associated debris or talus pile (Crowell and Leatherman 1999).

Rectification and Registration of Historical Maps and Aerial Photography

In order to use historical maps (such as NOS T-sheets) and aerial photographs in historical shoreline change exercises, one must be able to register these data to the same coordinate system used by other data sources, such as GPS and LIDAR surveys. At the same time, errors and distortions must be corrected. The methods for registering and correcting historical maps and aerial photographs have their differences.

Historical Maps. To register historical maps to a specific coordinate system, one must identify a number of permanent and stable points or features on the map for which accurate and current geographic coordinates are known. The features are known as primary control points. For computer applications, a minimum of four primary control points are required, although six to eight points are commonly used to improve accuracy and aid in error analysis. These points are used to calculate the transformation necessary to change from one coordinate system to another and to correct for error factors such as scale differences, rotation, and uneven map shrinkage. Selection of primary control points is trivial if the source map contains current geo-coordinates, such as latitude and longitude tick-marks. However, historical maps often have latitude-longitude lines based on obsolete datums. For example, latitude-longitude coordinates for US maps produced prior to 1927 must be updated to better standards such as the North American Datum of 1927 (NAD27), or preferably the more recent NAD83. If latitude-longitude coordinates are not available for the map or if the coordinates are present but with an

unknown datum standard, then other techniques must be used to update the map. One very useful technique is to search the map for survey stations or hard permanent features for which modern coordinates are known. In the case of NOS T-sheets, several triangulation stations are usually mapped for which current coordinates can be obtained (Crowell et al. 1991).

Aerial Photography. Since 1927, aerial photography and photographic methods have been used to provide topographic information along the coast, creating an important dataset for use in historical shoreline change studies. Air photos, however, are not map projections; corrections for distortion due to tilt, tip, and yaw (and sometimes relief) must be made prior to and/or concurrent with the compilation process so that shoreline positions obtained from air photos can be compared to those compiled from planimetric maps, GPS, and LIDAR surveys. Several computer programs are available that automate air photo rectification in a process called “soft-copy photogrammetry” (Moore 2000). The process involves creating photomosaics prepared from scanned and rectified aerial photographs. A series of ground control points are used to spatially reference the imagery to geo-coordinates. Pass points, which are identifiable landmarks common to two or more photographs (but not ground controlled), are supplemental and serve to reference the photos to each other (Coyne et al. 1999). In summary, the rectification process transforms the imagery to a map projection and corrects or minimizes map distortions, such as tilt, scale differences, relief displacement, and radial lens distortion.

Accuracy of Source Data

Accuracy of source data has been discussed in a number of publications (Shalowitz 1964; Morton 1974; Dolan et al. 1980; Leatherman 1983; Everts et al. 1983; Anders and Byrnes 1991; Crowell et al. 1991; Thieler and Danforth 1994; Gorman et al. 1998; Moore 2000; Hapke et al. 2011; Wernette et al. 2017). In general, most researchers acknowledge that historical shoreline maps and historical aerial photographs (combined with more recent LIDAR and GPS surveys), provide an important source from which to monitor shoreline movement and calculate rates of change.

NOS T-Sheets and Aerial Photography. Questions regarding accuracy of older source data, particularly historical NOS T-sheets, spurred Crowell et al. (1991) to devise a way to determine the maximum errors and measurement uncertainty one would expect to find using historical and current aerial photography and historical NOS T-sheets. Using rectification and registration methods described above, the authors approximated errors for each individual step in shoreline data compilation, including survey, digitization, mapping, and map and aerial photograph distortion. The square root of the sum of squares was used to obtain estimates of

measurement uncertainty in the digitized location of the interpreted shoreline (in this case, the high water line). Measurement uncertainties were calculated for the following data sources (all at 1:10,000 scale): 1846–1930 T-sheets prepared prior to the use of aerial photography: 8.4–8.9 m; post-1950s T-sheets compiled using aerial photography: 6.1 m. Crowell et al. (1991) also estimated measurement uncertainties of 7.5–7.7 m for aerial photography, albeit acknowledging that there is additional uncertainty resulting from misinterpretation of the HWL from poor quality aerial photography. These estimates were based on analysis of T-sheets and aerial photography from the northeast coast of the United States. Importantly, short-term process variability factors, such as daily and monthly tidal changes and storm effects, were not included in the error estimates. Comparable measurement uncertainties for T-sheets and aerial photography are presented in Hapke et al. (2011), who used similar statistical methods.

GPS Surveys. Field-based shoreline surveys using global positioning systems (GPS) have been used since the early 1990s (Morton et al. 1993). Shoreline position data are collected by driving an off-road vehicle equipped with a GPS unit directly along the HWL. Geo-coordinates are collected at specified intervals, allowing real-time digitization of the HWL. GPS surveys have an advantage over older plane-table, alidade, and rod surveys (as used, for example, by the field-based NOS topographers) in that there is little need to interpolate and sketch the HWL between rodded points. Indeed, GPS measurement of the interpreted position of the HWL are so accurate (within a meter) that measurement error is usually insignificant compared to the ambiguity inherent in the determination and identification of the HWL.

LIDAR Surveys. LIDAR surveys have an advantage in that it is possible to objectively determine the MHW shoreline based on tidal datums or elevation contours, rather than relying on an uncontrolled shoreline proxy, such as the HWL (Hapke et al. 2011). However, as noted by Morton et al. (2004) and Hapke et al. (2011), there are horizontal and vertical differences (offsets) in the location of the HWL determined from beach surveys, aerial photographs, and GPS surveys, as compared to the MHW shoreline determined from LIDAR surveys. The offsets can be assessed and corrected for by using methods described in Hapke et al. (2011).

Calculation of Long-Term Erosion Rates

A primary reason for calculating long-term erosion rates is for forecasting erosion hazard areas which are commonly used for land-use planning, ranging in purpose from providing information to property owners, to establishing regulatory setback lines for coastal construction. Usually building setbacks are based on the average annual erosion rate for the

area, multiplied by a specified number of years, typically 30 and 60 years. The computed setback is then measured landward from an erosion reference feature, whose movement would have the most impact on the structural integrity of a building. As such, the erosion reference feature is typically the top edge of a bluff, dune escarpment, vegetation line, beach scarp, or perhaps the HWL (The erosion reference feature for any given stretch of beach may not necessarily be the same feature or shoreline proxy used to determine the historical erosion rate). If the erosion rate at a site is, for example, 5 ft per year, then the 30-year setback is 150 ft and the 60-year setback is 300 ft, both of which would be measured landward from, for example, the top edge of the bluff.

The predictive accuracy and uncertainty of erosion forecasts have been discussed in a number of papers (Crowell et al. 1993; Crowell et al. 1997; Douglas et al. 1998; Galgano and Douglas 2000; Douglas and Crowell 2000; Honeycutt et al. 2001; Genz et al. 2007; Hapke et al. 2011; Fletcher et al. 2012; Del Río and Gracia 2013). While accuracy of the source shoreline data is paramount, there are other important matters to consider prior to calculating long-term rates of erosion. For example, a decision must be made regarding: (1) The temporal span of the shoreline data to be used in determining the rate – for example, short-term (<30 years), medium-term (30–60 years), or long-term (>60 years) data; (2) The statistical method or model that should be used to determine the rate – for example, end point rate, linear regression, modeling; and (3) whether certain shoreline data should be deselected from the historical shoreline change dataset.

Temporal Span. Crowell et al. (1993) and Galgano et al. (1998) demonstrated the importance of using data sets spanning 60–80 years or more to determine the long-term trend of erosion for forecasting future shoreline positions. Such long data spans are needed because the erosion/recovery cycle associated with severe storms can obscure the underlying trend of erosion for decades. However, in many cases, only the more recent sequences of historical shorelines should be used, particularly in areas where prolonged and perhaps permanent physical changes to the beach system have occurred (see below).

Statistical Methods. A simple method that is sometimes used by researchers and planners to calculate rates of erosion is the “end-point-rate” method. In this method, the analyst selects two representative shorelines, often the earliest and the most recent, to calculate the amount of shoreline movement. This distance is then divided by the time elapsed between successive shoreline positions. The advantages of this technique include simplicity in application and two shorelines suffice to obtain a rate. A major disadvantage is that the position of one or both of the endpoint shorelines may be aberrant, and so the rate of change can be misleading for use in determining future shoreline positions. Moreover, data between the end points, of value in determining the rate of

change, are not used and hence potentially useful information is ignored.

Another method popular among researchers is the use of linear regression. With this method, a best-fitting straight line is determined that minimizes the sum of the squares of the lengths of vertical line segments drawn from the individual data points to the fitted line. The advantages of this method are that: (1) all data points (except for poststorm shorelines) are used in the rate calculation, thereby reducing the influence of spurious data points; (2) linear regression is an easily understood statistical analysis tool that can be performed with most spreadsheet programs, and (3) summary and related statistical techniques are available to test and measure the quality of the straight-line fit and estimate the variance of the data (Kleinbaum and Kupper 1978). Other empirically based statistical methods have been proposed to calculate long-term erosion rates, but because of the scarcity and uneven sampling of the data, it is debatable whether these methods improve on the forecast (Crowell et al. 1997).

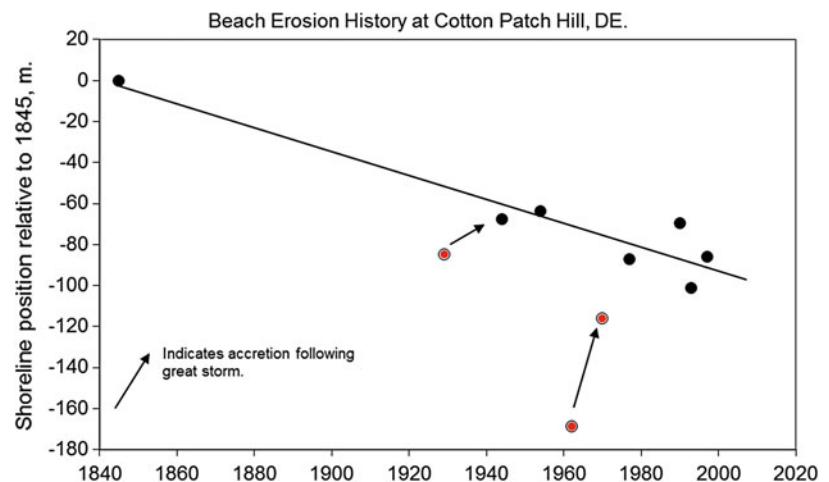
Deselection of Historical Shorelines. A decision whether to deselect certain shoreline datasets a priori is critical in determining a realistic historical erosion rate as certain historical shorelines may not be representative of the long-term trend. As an example, because of large seasonal fluctuations in beach width (particularly in the non-summer months), only summertime data should be used in historical shoreline change analyses (Zhang, et al. 2002). This is especially true when using historical NOS T-sheets in the shoreline change analysis, as T-sheets were almost always mapped during summer beach conditions. Combining a sequence of older “summer beach” NOS T-sheets with a sequence of more recent “mixed season” imagery/GPS shoreline data could lead to results biased towards higher rates of long-term historical erosion or lesser rates of long-term accretion. In addition, the emplacement of a major shoreline engineering structure (e.g., groins, jetties, and seawalls) may have a demonstrable long-term effect on the beach and make older

data unrepresentative of the long-term trend. Similarly, a major storm could open up an inlet, thereby altering the long-term trend on nearby beaches. In these situations, it may be prudent to deselect older, pre-event shoreline data.

A related question is whether immediate poststorm shoreline imagery should be included in the historical shoreline datasets. Insight into this question can be gained from the following example of an actual history of shoreline positions. Consider the data shown in Fig. 1 (after Douglas and Crowell 2000). The graph shows ten shoreline positions from 1845 to 1997. The shoreline location data are from the Atlantic coast of Delaware, just south of the Indian River Inlet near a location called Cotton Patch Hill. The 1929, 1962, and 1963 shoreline positions were photographed or mapped immediately following major storms and are shown as red circles on Fig. 1. Note that these three poststorm shoreline positions are inconsistent with the long-term trend line determined from the other seven shoreline positions. During a large storm, more erosion can occur in a few days than in the previous 50 years, which is often followed by an extended period of beach recovery. Note that accretion continued after the 1962 Ash Wednesday nor’easter for at least a decade. Therefore, determination of the long-term rate of beach erosion should not be undertaken without consideration of the role of severe storms on shoreline position.

The example above also serves to illustrate the proper statistical approach to the shoreline position forecasting problem. A correct methodology requires that the forecast model be in reasonable agreement with the actual physical situation. Figure 1 shows that there was a long-term trend of beach erosion, superimposed with episodes of severe erosion followed by extended periods of recovery. Use of poststorm shoreline positions in determination of the long-term trend will yield aberrant results.

Erosion: Historical Analysis and Forecasting, Fig. 1 Beach Erosion History at Cotton Patch Hill, DE



Newer Methods and Addressing Sea Level Rise

The above statistical methods for determining long-term erosion rates and future shoreline locations typically assume stationarity; that is, the forecasted rate of shoreline change is assumed to be linear and equal to the historical linear rate of shoreline change. As such, the method does not consider potential acceleration or deceleration caused by geophysical processes, such as changes in the rate of relative sea level rise. Long-term sea level rise, however, is an “enabler” of long-term beach erosion. As sea levels rise, land is inundated by the rising waters, and higher water levels allow increasing erosion. Accordingly, the location of a receding shoreline is dependent on both sea level rise-induced static inundation combined with dynamic erosion. If relative sea level rise is projected to accelerate, then in most cases long-term erosion rates should also be projected to accelerate (TMAC 2015).

The best known model to link sea level rise with erosion is the Bruun Rule. The Bruun Rule is a two-dimensional model that predicts a landward and upward displacement of the beach profile in response to sea level rise (Ranasinghe et al. 2012). The Bruun rule, however, has long been controversial, and while it may be suitable for regional scale assessments, it is generally recognized as unsuitable for localized studies which require reliable estimates of shoreline retreat (Ranasinghe et al. 2012). Over the past few years, however, new methods and models have been proposed that predict future shoreline locations in which the influence of sea level rise is considered. For example, Romine et al. (2009) developed polynomial methods that consider acceleration (both increasing and decreasing) in order to provide additional information regarding how erosion rates may vary through time. Anderson et al. (2015) combined historical erosion rates with a Bruun-derivative model (from Davidson-Arnott 2005) that incorporates projected acceleration in sea level rise. Ranasinghe et al. (2012) have developed process-based models that provide probabilistic estimates of long-term recession that incorporates sea level rise. Finally, Gutierrez et al. (2011, 2014) used a Bayesian network to produce probabilistic forecasts of future shoreline locations assuming accelerated sea level rise (TMAC 2015).

Conclusions

Many difficulties can be encountered in accurately determining the long-term behavior of shorelines based on sparse data sets. The complexity increases if engineering changes (such as construction of groins or jetties) have been made to the shoreline or if tidal inlets are opened as a result of a large storm. In addition, capes and spits often display a complex episodic cycle of erosion and accretion that is unrelated to storms. Thus the highest-quality forecasts of long-term

shoreline position need to consider meteorological and geomorphic factors and the role they play in the evolution of the shoreline position.

Cross-References

- [Beach Erosion](#)
- [Beach Features](#)
- [Cliffs, Erosion Rates](#)
- [Coastal Erosion](#)
- [Coastline Changes](#)
- [Erosion of Coastal Systems](#)
- [Erosion Processes](#)
- [Monitoring Coastal Geomorphology](#)

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Estuaries

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In the now voluminous coastal literature, the definition of “estuary” is highly variable, and frequently reflects the standpoint of the author. Pethick (1984) states that: “estuaries are among the largest and most complex of all landforms, so complex in fact that attempting to provide a basic definition has proved extremely difficult.”

Indeed one writer describes the definition of an estuary as rather like pornography – difficult to define, but you know it when you see it!

Central to the concept is that *estuaries are the tidal mouths of rivers* (Pethick 1984). The French geomorphologist Guilcher (1958) pointed out that the word estuary comes from the Latin *aestus*, meaning tide, but that in France it has a popular synonym, *etier*, and the two words denote that part of the river system where the tides (water level rise and fall) and tidal currents make themselves felt. This is essentially the definition adopted by Fairbridge (1980) who defined an estuary as: “an inlet of the sea reaching into a river valley as far as the upper limit of tidal rise.” This concept is also accepted by Caspers (1967) who argues that “the upper limit of the estuary is determined not by salinity but by tidal forces. In other words it is determined hydrodynamically rather than hydrochemically.”

Terry R. Healy: deceased.

Thus, in principle an estuary can be defined from several different perspectives, including from the point of view of the geomorphology, sedimentation processes, tidal influences, the physical oceanography/salinity structure, or the ecology. In simple terms the definition is perhaps best considered from the standpoint of either “form” or “function.”

From the geomorphic perspective, an estuary is defined in the general sense as: “the mouths of rivers, widening as they enter the sea” (Bird 1972). On a global scale they typically form in drowned river valleys. The Webster Dictionary (2nd edn) describes an estuary as “a passage, as the mouth of a river or lake, where the tide meets the river current; more commonly, a narrow arm of the sea at the lower end of a river.” However, the authoritative Glossary of Geology, edited by Jackson (1997) defines an estuary variously as:

(a) the seaward end or the widened funnel-shaped tidal mouth of a river valley where freshwater mixes with and measurably dilutes sea water and where tidal effects are evident; e.g., a tidal river, or a partially enclosed coastal body of water where the tide meets the current of a stream. (b) A portion of an ocean, as a firth or an arm of the sea affected by freshwater; e.g., the Baltic Sea. (c) a drowned river mouth formed by subsidence of land near the coast or by the drowning of the lower portion of a non-glaciated valley due to the rise of sea level.

In physical geography terms the Webster Dictionary (2nd edn) defines an estuary as: “a drowned river mouth, caused by the sinking of the land near the coast.” Stamp, in his *Glossary of Geographical Terms* (1966), defines an estuary as a “tidal opening, an inlet or creek through which the tide enters; an arm of the sea indenting the land,” but Stamp states that this definition is rare in modern use.

Cameron and Pritchard (1963) provide a definition of estuaries perhaps the most widely accepted by coastal scientists, as: “a semi-enclosed coastal body of water which has a free connection with the open sea and within which sea water is measurably diluted with fresh water from land drainage.” This definition appears in most of the modern textbooks such as Pethick's (1984): “*An Introduction to Coastal Geomorphology*” and Black (1986): “*Oceans and Coasts*.” Such a definition leads to a classification of estuaries as a function of their salinity structure, focusing on the oceanographic properties determined by the physical dimensions in relation to river flows, salinity distribution, and tidal conditions. Another influencing factor is the Coriolis effect which becomes important for wide estuaries. Based upon these physical characteristics, estuaries may be classified as follows (Bowden 1967): (1) *Salt wedge estuaries* occur where river flows dominate the salinity structure. In such cases, the river waters flow seawards as a layer of freshwater separated from the underlying saltwater by a sharp density interface. This reduces the turbulence and mixing between the two separate water bodies, and the saltwater extends as a down-sloping wedge along the bottom into the head of the estuary. (2) *Two*

layer flow with entrainment. Again two distinctive water bodies of different salinity and density are apparent in the vertical, but when internal waves at the interface break, salt-water is entrained upwards into the surface freshwater, thereby increasing its salinity. To compensate for this entrainment there is an upstream directed residual flow along the bottom. It is this current that facilitates transport of fine suspended material upstream into the upper reaches of the estuary. (3) *Two layer flow with vertical mixing*. In relatively shallow estuaries the turbulence generated by the tidal flows generates mixing through a greater proportion of the water body. Fresher water is mixed downwards and saline water upwards, but there is no sharp density interface boundary. However, there are still two layers of flow and a bottom landwards residual current. (4) *Vertically homogeneous estuaries*. When tidal currents are strong in relation to the freshwater input, or in shallow estuaries where form drag induces turbulence throughout the flow, the vertical mixing becomes complete throughout the water column, and there is no measurable variation in salinity from the surface to the bottom. There is, of course, a horizontal gradient of salinity from the estuary head to the mouth.

Wind may have an important influence on the estuarine density driven circulations. Wind stress on the water surface produces a net transport of surface water, and the waves generated increase the intensity of vertical mixing. These processes on occasions can override the normal estuarine density circulation.

Physiographically estuaries are of four identifiable types (Healy and Kirk 1982). These are: (1) *Fjords*, characterized as former valley glaciers in mountainous terrain, with steep sides, a deep basin, and in many cases a sill at its mouth. Such features occur in Norway, Scotland, southern Chile, and southwest New Zealand. But there are also more complex lowland fjords of the Baltic type, formed originally as meltwater depressions under the Pleistocene ice sheet, and flooded with the postglacial sea-level transgression. (2) *Drowned river valleys* or *rias* occur widely along mid-latitude coasts, and occur in their purest form as meandering river valleys along coasts with hard crystalline basement rocks with low sedimentation rates. In many cases, complete valley systems were drowned resulting in a dendritic pattern of arms. Along active sedimentation coasts the rias are often modified as they connect with the open sea, typically exhibiting sediment infill of the drowned river valley to form modern tidal flats (Fig. 1) and low alluvial plains developed over the estuarine sediments. They may possess barriers across the mouth of the embayment, such as occurs on the Oregon and New Zealand coasts (Fig. 2). Where major faulting occurs the rias may become fault-aligned as in New Zealand's Marlborough Sounds. (3) *Barrier enclosed estuarine lagoons* occur along the low sandy barrier coasts of the world, such as the Atlantic seaboard of the United States or the eastern coast of Australia (Fig. 3). The older term

Estuaries, Fig. 1 Drowned river valley with active infill sedimentation forming muddy tidal flats, Raglan Harbour, New Zealand (Photo: T. Healy)



Estuaries, Fig. 2 Estuary of the Waioeka River mouth, Bay of Plenty, New Zealand, illustrating a Holocene sandy barrier formation across the river mouth (Photo: T. Healy)



“bar built” estuaries is genetically misleading, and should be avoided, as few barriers originate from “bars.” The enclosing barriers typically comprise Holocene dune barrier-islands and spits as is typical along the Atlantic and Gulf coasts of the United States, but that is not always the case. Some enclosing barriers may comprise high (up to 200 m) Pleistocene dunes, as for the very large harbors on the west coast of the North Island of New Zealand, and sometimes the enclosing barriers may be of gravel, and even boulders, as for example at Nelson, New Zealand. These lagoonal-type estuaries are characterized by shallow depths and frequently broad expanses of intertidal flats become exposed at low tide. Where they occur along littoral drift coasts the entrances are characterized by narrow

tidal inlets with ebb and flood tidal delta systems. Because of the overall shallow lagoon, the estuaries are well mixed in terms of their salinity structure, although there may be salinity variation from the mouth to the head of the lagoonal estuary. In arid coasts, the lagoonal waters may become hypersaline. (4) *Structural and tectonically induced estuaries* are typified by downthrown grabens (San Francisco Bay), tectonic depressions (Botany Bay, NSW), or flooded calderas (Banks Peninsula, New Zealand). These tend to be large-scale features (Fig. 4).

Clearly, there is a geomorphic gradation both within and between large estuaries. For example, a large barrier-enclosed estuarine lagoon may possess arms in its upper reaches that

Estuaries, Fig. 3 The shallow estuarine lagoon of Maketu, Bay of Plenty, New Zealand, enclosed by a Holocene barrier spit. The entrance is a tidal inlet system (Photo: T. Healy)



Estuaries, Fig. 4 The down faulted structural graben of the Firth of Thames, New Zealand, into which flows the large Waihou River. The name “Firth” is a geomorphic misnomer, as this structural estuary did not originate from ice scour (Photo: T. Healy)



exhibit the geomorphic characteristics primarily of a drowned river valley, and salinity characteristics of a salt wedge and two-layer flow. Conversely, a drowned river valley estuary with overall dendritic outline may possess barriers across its mouth (e.g., Chesapeake Bay, or Raglan Harbour in New Zealand).

However, all of these geomorphic categories of estuaries feature an element of “drowning” from rise of the postglacial sea level, which reached its approximate present level some ~6,000 years ago (Russell 1967). However, continued drowning of an estuary is rarely identified in modern times, despite the present global warming and its consequent sea-level minor rise.

On the contrary, active sedimentation within estuaries overall outweighs any modern sea-level rise effect. This is

especially so for high sediment load major rivers such as the Amazon, the Yellow River of China, and the Mississippi, where rapid sedimentation and the growth of deltas has greatly modified the estuary over the past ~6,000 thousand years. Indeed, estuaries are well recognized as natural sediment traps, and are ephemeral features of the geological record. As Russell notes, most estuaries have changed considerably from the active sedimentation during the past few millenia, and many estuaries are headed for extinction. Rapid sediment input into estuaries from the inflowing river catchments, acts to fill them up from the land. But (especially for lee coast estuaries) diabathic onshore transport across the shelf provides sediment at the coasts that sweeps into the estuary mouths, and acts to infill estuaries from the sea. In addition,

the saline density circulation within estuaries and mixing of mud-laden freshwaters from the catchment with the saline marine waters, leads to flocculation of fine particles, which residual bottom currents tend to transport toward the head of the estuary, where the muds deposit and accumulate.

Within most estuaries tides are the major energy source for mixing fresh and saltwater, and for resuspending sediment from the bed and transporting those sediments seawards or landwards. Frictional effects dissipate the tidal wave energy within the estuary, and deform the tidal wave. But this is offset by convergence of the tidal wave as it progresses up the estuary. Depending upon the balance between tidal wave convergence compared with frictional dissipation effects, the tidal range within an estuary can be larger than at the open sea before decreasing landwards toward the river (a hypersynchronous condition), or the frictional effects can dominate and the tidal amplitude decreases landward from the mouth, a hyposynchronous condition. Most ria type estuaries are a hypersynchronous and attain maximum current strength in middle or landward reaches, whereas for the barrier enclosed estuaries the maximum current strength occurs at the narrow tidal inlet.

In many estuaries, there exists an essentially closed circulation system for fine suspended particles, which causes them to become entrapped. This occurs due to the formation of "turbidity maxima" and from the "settling lag" effect. In the latter process, there is a time lag between the turn of the tide (when current velocity is zero) and when the concentration of fine suspended particle concentrations become minimum. This settling lag is due to the time it takes for the fine material to settle through the water column after the reduction in flow velocities (and reduction in turbulence), and thus the particle is carried some distance beyond the point where it starts to sink. Thus for a flood tidal current near high water, a certain amount of fine sediment settles toward the landward areas. The lag effect can explain why in many estuarine tidal flats there is a much greater proportion of mud particles than in the adjoining open sea, and why moving landwards up the estuary the particle size gradually fines (Postma 1967).

A "turbidity maximum" of suspended sediment concentration often occurs in the upper reaches of many estuaries, particularly in the partially mixed types, where suspended sediment concentrations may reach 10 to 100 times greater than either seaward or further up the river, and may reach 1–10 g/L. This feature occurs in a wide variety of estuaries over the world, and occurs because fine particle settling is trapped in the null zone where river current converges with the landward estuarine flow near the bottom (Nichols and Biggs 1985).

The formation of highly concentrated fluid mud layers at the bottom may also be part of the sediment dynamics. Fluid mud is a highly concentrated sediment suspension

with a sediment concentration of between 10,000 and 300,000 mg/L, and can originate from the turbidity maximum at slack water. It may take the form of mobile or stationary suspensions (Kirby 1988). But fluid mud can also form from wave action (Healy et al. 1999).

A typical feature of shallow estuaries is the estuarine "turbid fringe." This is generated by the wave action as the tide rises over the intertidal flats and reagitates the fine particles into suspension. Thus even in days of very light wave action a zone of brown turbid waters is observed around the shore of the estuary. Fine sediment transport in the estuary is also affected by specific processes such as aggregation of the fine particles, deposition, erosion, and consolidation. Classically, the aggregation of fine sediment particles was explained by the so-called electric double-layer theory. In recent decades, additional mechanisms have been identified including organic aggregation, bioflocculation, and pelletization.

Cross-References

- [Barrier](#)
- [Barrier Island Landforms](#)
- [Estuaries: Anthropogenic Impacts](#)
- [Muddy Coasts](#)
- [Ria](#)
- [Salt Marsh](#)
- [Tidal Creeks](#)
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- [Tidal Prism](#)
- [Tides](#)
- [Vegetated Coasts](#)
- [Volcanic Coasts](#)
- [Wetlands](#)

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biogeochemical cycling of elements. They therefore can strongly influence the environmental quality of coastal waters. Estuaries also protect coastal watershed areas, buffering infrastructure from the damaging effects of storms, floods, waves, and erosion. Because more than four billion people live within 60 km of the world's coastlines (Kennish et al. 2008), many inhabiting estuarine watersheds, these buffering effects are extremely important for sustaining coastal communities.

Various industries, aside from fisheries, depend heavily on estuaries for their successful operation. Most notable in this regard are aquaculture, electric power generation, oil and gas recovery, marine biotechnology, tourism, transportation, and shipping. Together, these industries inject trillions of dollars into the world economy each year. They also employ millions of people.

Estuaries: Anthropogenic Impacts

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Definition

Anthropogenic impacts are defined as marked adverse effects of human activities on estuarine environments (Kennish 2015a).

Introduction

Estuaries are coastal environments with exceptional ecological, recreational, and commercial value (Day et al. 2012; Kennish 2015a). They provide a wide array of ecosystem services, including diverse habitats (e.g., open waters, submerged aquatic vegetation, unvegetated bottom sediments, tidal flats and creeks, and fringing wetlands) that serve as nursery, feeding, and refuge areas for numerous estuarine, marine, and terrestrial organisms. The rich food supply in estuaries supports many biotic communities. Numerous marine species of recreational and commercial importance utilize estuaries during their life; for example, these coastal systems play a significant role in the production of marine fisheries. Adjoining wetland habitats are particularly important in the life history of fish, shellfish, migratory birds, and other wildlife.

Within the land-river-coastal shelf continuum, estuaries serve a number of vital functions, including the filtering of contaminants, transformation of nutrients, and

Anthropogenic Impacts

Kennish (1992, 1997, 2002, 2005, 2015b) and Kennish et al. (2008) have examined the effects of human activity on estuaries. They have identified a wide array of anthropogenic stressors on these ecosystems which can be organized into 12 major categories. These include: (1) habitat loss and alteration; (2) enrichment (nutrients, organic carbon, and thermal loading); (3) sewage and pathogenic inputs; (4) chemical contaminants; (5) human-induced sediment/particulate inputs; (6) dredging and dredged-material disposal; (7) human-altered hydrological regimes; (8) invasive/introduced species; (9) overfishing; (10) climate change effects; (11) coastal subsidence; and (12) floatables/debris.

Habitat Loss and Alteration

Estuaries are particularly susceptible to habitat loss and alteration caused by human activities. Examples are physical changes in bayshore areas linked to the construction of lagoons, bulkheads, retaining walls, docks, piers, boat ramps, marinas, houses, roadways, and other infrastructure. Waterfront development which creates permanent structures, and excessive recreational and commercial use which alters bottom habitat especially in shallow waters due to boat engine prop scarring of seagrass beds and shellfish dredging of bay bottoms, have had profound consequences. Conservation zones have been established in some of the more sensitive and valuable estuarine habitat impacted by human activity. Dredging of lagoons, channels, and other estuarine areas disrupt benthic habitats for months or even years. Dredged-material disposal at selected sites in estuarine basins causes longer term alteration of habitat, albeit in restricted areas. Revetments constructed for bank stabilization along shorelines to mitigate erosion from coastal storms and currents and to protect against sea-level rise and inundation are increasing

in many regions. Other protective features are being installed or reconstructed in hazard zones, including storm walls, flood walls, and jetties. The damaging effects of major storms such as Superstorm Sandy in the mid-Atlantic region and Hurricane Katrina in the Gulf of Mexico have led to the application of innovative marine engineering strategies to increase the sustainability and resilience of coastal communities, land areas, and estuarine ecosystems.

Wetland habitat destruction has declined appreciably in the USA over the past 50 years as a result of strong regulatory controls, habitat restoration, and education programs. However, the loss of wetlands has continued unabated in other countries, particularly undeveloped ones. The losses have been acute in some countries. For example, most wetland and mudflat habitats along the mainland coast of the Dutch Wadden Sea have been lost due to reclamation, diking, levee construction, and conversion to polders. Parallel grid ditching, canal and levee construction, impoundment dikes, and open marsh water management (OMWM) have physically altered many salt marsh systems along the Atlantic and Gulf Coasts of the USA, their hydrological regimes, and biotic structure (Kennish 2001a, 2001b; Rochlin et al. 2012; Gedan 2015). Tidal flooding and drainage have been modified in many salt marshes, often reducing sediment input to the marsh surface and decreasing vertical accretion which could hasten submergence of these systems in response to climate-change driven sea-level rise.

In the USA, ~50% of the original salt marsh habitat was destroyed before 1970 (Kennish 1997, 2001b). Adam (2002) noted that considerable salt marsh area in Northern Europe has been altered as a result of land reclamation, pollution, tidal barriers, salt production operations, insect control, and introduced species. Burd (1992) indicated that 10–44% of the salt marshes in South East England were destroyed between 1973 and 1985/1988. The substantial losses of salt marshes along the lower Mississippi River have been attributed in large part to land subsidence together with a reduction in sediment load in the river due to dam construction upstream as well as extensive levee systems that preclude nutrient and sediment inputs to the marsh surface (Day et al. 2012; Mitsch and Gosselink 2015). Kennish (2001b) and Adam (2002, 2015) have provided a detailed account of the natural and anthropogenic stressors impacting salt marsh systems.

Similar losses have been reported for mangrove forests. For example, human activities have accounted for more than 50% of the mangrove forests lost in Southeast Asia and 75% of the mangrove forests lost in Puerto Rico (Alongi 1998; Kennish 2015b). Mariculture and silviculture operations are responsible for much of these losses. Mariculture ponds have replaced nearly 50% of the mangroves in the Philippines (Hopkins et al. 1995; Kennish et al. 2008).

Aside from bayshore construction and shoreline habitat modification, development of coastal watersheds has

converted extensive natural habitat to compacted soils and impervious cover that decrease infiltration of rainwater while increasing runoff, erosion, and nonpoint pollution to estuarine water bodies. Land use and land cover changes in these developed areas often translate into higher nutrient and sediment loads that often affect water and sediment quality of estuaries. A watershed impact threshold is exceeded when the amount of impervious surface cover is greater than 10% (Arnold and Gibbons 1996), and increasing water quality degradation of influent systems in coastal watersheds occurs with progressively greater impervious cover. Sprawl development accelerates the fragmentation of natural habitats in watersheds, creating an impediment to colonization and migration of wildlife. Freshwater diversions, dam construction, and impoundments alter hydrological and salinity regimes in influent systems and contiguous estuarine waters, affecting the structure and function of biotic communities. The channelization of streams and rivers for flood control enhances freshwater discharges to estuaries. However, dams reduce the downstream flow as well as inputs of sediments and other particulates. The diversion of freshwater for agriculture, municipal, and industrial uses also reduces freshwater inputs. This is exemplified by conditions surrounding San Francisco Bay, where more than 50% of the freshwater inflow is diverted elsewhere, mainly for agriculture use (e.g., irrigation) which accounts for more than 80% of the diverted water volume (Kennish 1997, 2004). Mining and silviculture usually raise sediment loads to estuaries, as well as metals and other contaminants transported to the coastal zone. These human activities have been problematic in many developing countries.

Enrichment

Enrichment, as used here, refers to the input of excessive amounts of nutrients, organic carbon, and heat (i.e., calefaction or thermal loading from electric power plants) to estuarine and coastal marine systems leading to ecological impacts. Nutrient enrichment, primarily nitrogen and phosphorus inputs from point and nonpoint sources, promotes algal blooms in estuaries and can cause anoxic and hypoxic conditions detrimental to biotic communities and habitats (Kennish 2002; Diaz 2015). The eutrophication of estuarine waters often has devastating consequences, at times culminating in heavy mortality of benthic and nektonic organisms, including recreationally and commercially important forms, and causing significant changes in estuarine community structure and function. Phytoplankton and macroalgal blooms (including toxic and nuisance algal blooms) stimulated by nutrient enrichment decrease light transmission to the benthos, and the shading effect typically reduces the production and areal distribution of seagrass and other favored habitats. The habitat change commonly results in a shift from seagrass-dominated systems to phytoplankton-

dominated systems in shallow coastal bays and lagoons (Kennish and Paerl 2010).

Nixon (1995) defined eutrophication as the increase in rate of production and accumulation of organic carbon in a system. The consequences of eutrophication in an estuarine waterbody are numerous and varied and include such impacts as low dissolved oxygen, harmful algal blooms (HABs), loss of essential habitat (e.g., seagrass and shellfish beds), lower biodiversity, reduced harvestable fisheries, imbalanced trophic food webs, declining system stability and resilience, diminished ecosystem services, and impacted human use. The effect of escalating eutrophication is potentially permanent alteration or loss of biotic communities and habitats and great ecosystem-level degradation (Kennish and de Jonge 2011). Influxes of nutrients from farmlands, animal feedlots, stormwater runoff, industrial wastewater discharges, sewage inputs, groundwater seepage, atmospheric deposition, and other sources have contributed to nutrient loading concerns in many estuaries. Increasing population growth and development in coastal watersheds have accelerated reactive nitrogen inputs to estuaries and the occurrence of estuarine eutrophication (Howarth 2008; Anderson et al. 2010).

Waste heat discharged by electric generating stations sited in the coastal zone has caused the calefaction of estuarine waters, affecting the physiological and behavioral responses of organisms most acutely in near-field regions of station outfalls (Kennish 1992, 1997). Biotic impacts often observed in these receiving waters are decreased primary production, altered reproduction, decreased growth, and increased mortality (e.g., heat-shock and cold-shock mortality of fish). Behavioral modification may be dramatic, with many nektonic organisms exhibiting avoidance or attraction responses to the heated effluent. Significant change in species composition, abundance, and diversity of biotic communities commonly occur in discharge canals of electric generating stations. Other more serious impacts of electric generating stations are the impingement and entrainment mortality of estuarine organisms, which usually far exceed the adverse effects of waste heat discharges.

Sewage and Pathogenic Inputs

Inputs of sewage and other organic wastes can exacerbate estuarine eutrophication problems by delivering excess nutrients and organic carbon to estuaries. These wastes may enter estuarine systems via malfunctioning septic systems, stormwater, domestic wastewater discharges, industrial effluents, farmlands, mariculture, wildlife, livestock and fish processing operations, dredged materials, marinas, and other sources. Organic carbon inputs from external sources augment the organic carbon pool generated within estuarine systems, including vascular plant subsystems (seagrasses, salt marshes, and mangroves), macroalgae, animal remains, and biodeposits (feces and pseudofeces).

Organic carbon levels are frequently highest (>100 mg/l) in estuaries receiving the largest volumes of sewage wastes. Overenrichment of sewage wastes in estuaries is linked to elevated biochemical oxygen demand and reduced dissolved oxygen levels. Hypoxia and anoxia of estuarine and shallow coastal marine environments have increased worldwide over the past 50 years largely due to anthropogenic activities (Diaz and Rosenberg 2008; Diaz 2015). Chronic organic carbon enrichment degrades benthic habitats and communities, altering species diversity, dominance, and density, while favoring proliferation of opportunistic forms (Diaz and Rosenberg 1995).

Pathogens (bacteria, viruses, protozoans, and helminths) increase in estuaries receiving sewage wastes. In these polluted waters, they pose a health risk to swimmers and humans consuming contaminated shellfish. Consumption of raw, viral-tainted shellfish can cause hepatitis and serious gastroenteritis. Pathogenic bacteria in estuaries are particularly threatening to human health. For example, *Vibrio cholerae* causes cholera-like infection (diarrhea, dehydration, and vomiting), and other pathogenic bacteria (*Shigella* spp. and *Salmonella typhi*) are responsible for dysentery and typhoid. *Escherichia coli* causes gastroenteritis and other maladies. Indicator bacteria (enterococci, fecal streptococci, and total coliform bacteria) have been used in monitoring water quality at bathing beaches and in shellfish growing waters to reduce health risks to humans. Helminths and protozoa in estuaries also raise potential health concerns. Among problematic helminths are hookworms, roundworms, tapeworms, and whipworms that are commonly associated with untreated sewage and sludge. *Entamoeba histolytica*, a pathogenic enteric protozoan that can impact human health, is also transmitted via raw sewage contamination (Kennish 2004). Aggressive management programs have been instituted in many coastal communities to better control pathogenic contamination of estuarine waters, including identifying and restoring function of malfunctioning septic tanks and sewage treatment facilities, replacing combined sewer overflows, as well as improving agricultural and livestock management, stormwater runoff, and boat and marina sanitary protocols.

Chemical Contaminants

Bottom sediments of estuarine and coastal marine environments are repositories for chemical contaminants, particularly near heavily populated metropolitan centers and other urbanized areas. Examples are the bottom sediments of Boston Harbor, Newark Bay, Raritan Bay, upper Chesapeake Bay, and San Francisco Bay (USA). However, even in suburbia, where land development and industrial activity have increased during the past several decades, contaminant input to estuaries can be significant. Bottom sediments accumulate chemical contaminants because many of the contaminants are particle reactive, sorbing to grain surfaces, and settling to the

estuarine floor. Some chemical contaminants concentrate in estuarine organisms or in the water column. In coastal watersheds, both point and nonpoint sources of chemical contaminants may occur, although nonpoint sources have become increasingly more important through time because of greater effectiveness in controlling the point source inputs to receiving waters. Principal delivery systems to estuaries include urban and agricultural runoff, municipal and industrial discharges, groundwater inputs, river inflow, and atmospheric deposition (Kennish 2015b). Some polluting substances enter estuaries from the weathering and erosion of rocks, soils, and ore-bearing deposits as well as from volcanic activity and atmospheric deposition.

Metals, halogenated hydrocarbons, and polycyclic aromatic hydrocarbons are major groups of chemical contaminants found in estuaries (Kennish 1992, 1997). When present in elevated concentrations, they are potentially damaging to estuarine organisms and habitats. Oil spills from vessels have severely impacted some estuaries, with the released aromatic and aliphatic hydrocarbons causing impairment of vital habitats (e.g., estuarine floor and beaches, mudflats, and salt marshes) and biotic communities for decades. Oil entering estuaries in wastewater discharges from coastal watersheds and fixed installations has also been detrimental. Volatile organic compounds (VOCs), while less threatening over the long term, can be acutely toxic as well.

Metals pose a potential danger to estuarine organisms because of their acute or chronic toxicity at higher concentrations, tendency to bioaccumulate, and persistence in the environment. Metals of concern include the transition metals (e.g., copper, cobalt, iron, and manganese), metalloids (e.g., arsenic, cadmium, lead, mercury, selenium, and tin), and organometals (e.g., methylmercury, tributyltin, and alkylated lead) (Kennish 1997, 1998; Kennish et al. 2008). Important anthropogenic sources of metals to aquatic systems are smelting facilities, tanneries, solid waste dumpsites, municipal wastewater systems, fossil-fuel combustion, dredged spoils, refineries, electroplating, and dye manufacturing. The uptake of metals by estuarine and marine organisms occurs via the removal of the contaminants from solution, from bottom sediments or interstitial waters, and from ingestion of food and suspended particulates. Some of these organisms have the capacity to internally regulate the concentrations of metals through activity of metallothioneins and lysosomes, which play a significant role in the sequestration and detoxification of the contaminants. Metals are potentially toxic to estuarine organisms above a threshold bioavailability; exposure of the organisms to elevated concentrations of metals may lead to a range of physiological and pathological responses, such as growth inhibition, aberrant reproduction and development, tissue degeneration, neoplasm formation, and genetic derangement. Other processes that can be adversely affected are respiratory metabolism, feeding, and

digestive efficiency which, when seriously impaired, cause an increase in mortality. Abundance and distribution of these organisms are often altered in the estuary as well. The effects are manifested most conspicuously in benthic communities that inhabit sediments with high metal burdens.

Halogenated hydrocarbons are some of the most persistent, ubiquitous, and chronically toxic contaminants found in estuaries. They mainly originate from agricultural and industrial sources and include an array of biocides (e.g., insecticides, herbicides, and fungicides), low-molecular-weight compounds (e.g., chlorofluorocarbons), and high-molecular-weight chemicals (e.g., chlorinated aromatics and chlorinated paraffins). Some of the most notable synthetic organic compounds belonging to this group of contaminants (e.g., PCBs, DDT, and non-DDT pesticides) rank among the most toxic constituents in estuaries (Kennish 2004). They are suspected etiologic agents which have been linked to a number of disorders in estuarine animals such as reproductive abnormalities, altered endocrine physiology, aberrant developmental patterns, skin and liver lesions, and cancer. DDT and PCBs are particularly hazardous substances because they biomagnify in food chains (Kennish 1997). Therefore, they pose a potentially serious threat to humans who consume contaminated seafood products.

Polycyclic aromatic hydrocarbons (PAHs) are also widely distributed contaminants, being found in estuarine environments worldwide. They largely derive from fossil fuel combustion, land runoff, municipal and industrial wastewater discharges, and oil spills (Kennish 1992). Natural processes (e.g., volcanic activity) also generate PAHs, but their inputs to estuaries are low relative to human sources. PAHs are a problematic class of contaminants because the higher-molecular-weight forms are predominantly carcinogenic, mutagenic, and teratogenic, and the lower-molecular-weight forms can be acutely toxic (Kennish 2004). PAHs, when readily bioavailable, produce hepatic neoplasms (e.g., hepatocellular carcinoma, hepatocellular adenoma, and cholangiocellular carcinoma) and other disorders in demersal fish and shellfish (Kennish 1997). They are also responsible for a range of sublethal effects manifested as biochemical, behavioral, physiological, and pathological aberrations in organisms that can contribute to changes in the structure of estuarine communities.

Human-Induced Sediment/Particulate Inputs

Land use and land cover changes in coastal watersheds coupled to development of shore communities facilitate sediment and other particulate inputs to estuaries, contributing to biotic and habitat impacts. The removal of natural vegetation and soils during construction of buildings and other structures promotes erosion and delivery of sediments to streams, rivers, and estuaries. Even more dramatic is the loss of large volumes of sediment to aquatic systems via deforestation associated

with silviculture operations. Sediment delivery to estuaries via these processes typically increases water column turbidity, light attenuation, and shading of the estuarine floor, thereby decreasing production of seagrass beds and other benthic habitat that supports numerous faunal populations including many commercially and recreationally important finfish and shellfish species. Moore et al. (2012, 2014) showed that elevated turbidity in some areas of the Chesapeake Bay system is responsible for the loss of extensive seagrass beds which once supported benthic communities. The degradation of these biotopes can significantly reduce production of an estuarine system.

Dredging and Dredged-Material Disposal

There are both direct and indirect impacts of dredging and dredged-material disposal in estuaries. The most profound direct impacts relate to the removal of bottom sediments and destruction of the benthic habitat and communities. The dredging action causes mechanical injury and smothering of organisms entrained in the sediments extracted from the bottom, typically causing mass mortality. Successful recruitment and recolonization of the dredged site by benthic organisms may not occur for a year or more. The disposal of dredged material at restricted locations results in similar impacts on the benthos (Kennish 1997, 2015a).

Dredging increases turbidity levels substantially in an estuary, although these effects are ephemeral. Elevated turbidity reduces light penetration in the water column which can adversely affect the growth and production of phytoplankton and submerged aquatic vegetation. The roiling of sediments at the dredged site also releases nutrients and chemical contaminants from the bottom sediments, remobilizing them to other areas of the system. Thus, water quality can be adversely affected as well.

On the positive side, dredging increases circulation in an estuary. Deepening of channels promotes recreational and commercial usage of the system. Sediment removed by dredging is often used for beach and bayshore nourishment to mitigate erosion effects and enhance land habitats.

Human-Altered Hydrological Regimes

Both natural and anthropogenic factors affect freshwater inflow to estuaries that modulates their physical-chemical and ecological characteristics. Freshwater inflow to estuaries varies seasonally and annually with the flux of precipitation, catastrophic storms, and major flooding events. Local and regional drought conditions – stochastic natural disturbances – often cause substantial reduction in surface and groundwater inflow that decreases estuarine flushing and increases water residence time and salinity. During other periods of increased precipitation, opposite conditions occur, with greater estuarine flushing and reduced water residence time and salinity. The vagaries in these natural drivers of change are exacerbated by anthropogenic

factors. For example, coastal watersheds that are heavily developed with impervious cover exceeding 10% have greater surface runoff to rivers, estuaries, and the coastal ocean than less developed coastal watersheds. Other system modifications, such as channelization and the loss of salt marshes and other wetlands, increase freshwater flows to estuaries which concomitantly increase inputs of nutrients and chemical contaminants commonly stored in soils and deeper sediments. In contrast, freshwater diversions in coastal watersheds for agricultural, municipal, and industrial needs decrease freshwater inputs to estuaries, as do dams and reservoirs in upland drainage basins. The effects of the reduced freshwater inflow are compounded during dry years. Excessive groundwater pumping for farmland irrigation and domestic uses has resulted in saltwater intrusion in some regions. Flood control structures installed to mitigate storm surge caused by hurricanes and major coastal storms also affect water-flow regimes in estuaries. With the increase in sea-level rise driven by climate change, there is growing awareness and concern about the effects of installing these structural features to mollify flooding and other bayshore impacts (Kennish 2002, 2015a; Kennish et al. 2008).

Changes in the circulation and salinity regimes of an estuary due to natural and anthropogenic fluxes in freshwater inputs can significantly affect the structure and function of biotic communities. The species composition, abundance, and distribution of floral and faunal communities will change with shifts in the estuarine salinity gradient. Changes in freshwater inflow will also alter an estuary's capacity to dilute, transform, and flush contaminants accumulating in the system (Kennish 2004, 2015a).

Introduced/Invasive Species

Introduced and invasive species have altered biotic communities in estuaries worldwide. These nonnative species often adapt very successfully to their new environments, outcompeting and excluding many native species and disrupting the trophic organization. In some cases, the ecological balance of the system is altered by acute reduction in biodiversity and increase in dominance by the nonnative forms. The adapted habitats of the introduced and invasive species commonly lack natural controls, enabling these forms to rapidly dominate the native plant and animal communities and potentially impacting the ecological and economic components of the system. New pathogens and parasites may enter estuarine water bodies with these nonnative species, leading to the spread of infectious diseases and harmful parasites. Recreational and commercial fisheries can be adversely affected as well, as can wetlands and other habitats (Kennish 2005, 2015b).

Many nonnative species have been deliberately introduced into estuaries for recreational or commercial purposes, an example being the introduction of the striped bass (*Morone saxatilis*) to San Francisco Bay (USA). In fact, San Francisco Bay is one of the most heavily “invaded” estuaries in the

world, with nearly 250 species of plants and animals being nonindigenous (Cohen and Carlton 1998). More than half of the fish species in the delta region consist of exotic, nonnative forms, and most macroinvertebrates along the inner, shallow water habitats of the bay are introduced species (Kennish 2000). Among the prominent introduced species in this estuary are the smooth cordgrass (*Spartina alterniflora*), oyster drill (*Urosalpinx cinerea*), and striped bass (*M. saxatilis*). The Asian clam (*Potamocorbula amurensis*), an introduced bivalve that has outcompeted native clam species (e.g., *Macoma balthica* and *Mya arenaria*), has seriously decimated the phytoplankton community of Suisun Bay. Dramatic changes in the native biotic communities of San Francisco Bay have been well chronicled (Kennish 2000, 2004).

The number of introduced and invasive species is on the increase in estuaries for several reasons. First, many species are expanding their geographic range as the planet warms in response to climate change, enabling the species to extend their distribution to higher latitude regions and estuarine water bodies previously uninhabitable for them. In addition, greater marine shipping and transportation provides an avenue for global transport of potential invasive species. Furthermore, more mariculture and other estuarine and coastal marine commercial ventures enable additional nonindigenous species to adapt to new estuarine environments.

Overfishing

The escalating human population, particularly in the coastal zone, is placing greater pressure on fisheries resources in estuarine and marine waters. The societal demand for seafood has led to the overexploitation of many recreationally and commercially important finfish and shellfish species and the collapse of some of them. Excessive fishing pressure on some species has been allayed by the expansion of aquaculture in many countries. Government regulations have been formulated as well to protect viable fisheries.

Overfishing not only depletes finfish and shellfish stocks but also impacts the structure and function of biotic communities. For example, the collapse of a major piscivorous fish population often results in the rapid increase of its prey species and potential imbalances in community structure and altered ecosystem function (Pinnegar et al. 2000; Kennish 2005). The fishing pressure on many marine species is unsustainable (Pauly et al. 1998). The trend therefore has been to “fish down” marine food chains, rendering affected ecosystems impoverished and less valuable (Williams 1998). The global catch has shifted from large piscivorous species to smaller planktivorous forms (Boehlert 1996). Since many recreationally and commercially important marine finfish species utilize estuaries at some point during their life history, estuaries are critically important to the long-term viability of these species.

Climate Change Effects

Climate change is rapidly becoming one of the most serious environmental, economic, and societal concerns around the world. The effects of climate change are multifaceted, impinging on agriculture, manufacturing, energy systems, fisheries, transportation, shipping, tourism, emerging technologies, and other major sectors of developing and developed nations. The earth is warming, and the consensus among climate scientists is that anthropogenic activities account for much of it (Oreskes 2004; Anderegg et al. 2009; IPCC 2013). Increasing global temperatures, ascribed in large part to carbon dioxide emissions, have been linked to greater frequency and severity of damaging storms, ongoing sea-level rise, coastal flooding, droughts and fires, and other hazards projected by climate forecasting models for the twenty-first century (IPCC 2013). Carbon dioxide levels in the atmosphere now exceed 400 ppm, the highest in centuries. Extreme climate events and sea-level rise will be increasingly hazardous to coastal communities worldwide. The damaging effects of major storms such as Superstorm Sandy in the Mid-Atlantic region (USA) and Hurricane Katrina in the Gulf of Mexico (USA) have drawn much attention to the seriousness of these phenomena.

Eustatic sea-level rise is attributed to thermal expansion of the oceans and the melting of mountain glaciers and continental ice sheets. During the twentieth century, the mean global temperature increase was 0.8 °C (1.4 °F), being responsible for <50% of the average eustatic sea-level rise of 1.8 ± 0.3 mm/yr. Global sea-level rise is currently 3.3 ± 0.3 mm/yr. The rate of sea-level rise will continue to increase during the twenty-first century, as demonstrated by climate change models, with the total sea-level rise expected to range from 52–98 cm (IPCC 2013).

Relative sea-level rise, however, is much greater in some regions, such as the Mid-Atlantic region of the USA. As noted by Milne et al., (Milne et al. 2009) and Miller et al. (2013), relative sea-level rise is affected by several important elements: (1) global average sea level; (2) regional and local subsidence or uplift (including thermal subsidence, sediment loading, flexural, and glacio-isostatic adjustment effects), gravitational, rotational, and flexural effects due to changing ice sheets; and (3) oceanographic effects (dynamic topography and tide-range change effects). Hence, during the twentieth century, the relative sea-level rise in the Mid-Atlantic region (USA) was nearly twice that of the global sea-level rise.

During the twenty-first century, climate change effects will significantly impact low lying coastal areas, including estuaries. However, because of the multiple environmental stresses from a wide range of local and regional anthropogenic activities, identifying the climate change impacts on estuaries could be complicated (Cronin 2015). Despite this complexity, climate models indicate that the frequency of coastal storms and storm surge will increase along with precipitation,

accelerating estuarine shoreline erosion and flooding of bayshore areas. Rising sea level and coastal inundation will also lead to substantial loss of some coastal wetlands, converting them to open water habitat, eliminating their protective buffer, and leaving coastal communities more vulnerable to extreme weather events (Kennish et al. 2008). While some wetland systems will be able to migrate inland with rising sea level, others will be subject to coastal squeeze due to human development behind them, causing the habitat to be submerged by the encroaching sea. For example, ~30% of the coastal salt marshes in New Jersey (USA) cannot migrate inland because of development.

Human-induced climate change will alter freshwater inputs, temperature and salinity regimes, biogeochemical processes, and the structure and function of biotic communities in estuaries (Kennish 2002, 2015a). Higher temperatures will result in latitudinal shifts in the distribution of many estuarine plant and animal species, and the migration patterns of others will be significantly altered. Estuarine basins will be modified, widening and deepening in response to sea-level rise. Shifts will occur in nutrient and sediment supply as well as freshwater inflow. In addition, tidal prisms and tidal ranges will change in many systems, altering basin circulation. The upstream penetration of saltwater will destroy extensive plant and forest vegetation in coastal watersheds and cause saltwater intrusion of potable water supplies. The ecological landscape of lower watershed areas will change considerably. Carefully crafted mitigation and adaptation plans will be needed for coastal communities to effectively deal with these coastal zone changes.

Coastal Subsidence

Human activities are contributing substantially to coastal subsidence in some areas; they often exceed the effects of natural processes. For example, oil and gas extraction (e.g., Galveston, Texas, USA) and groundwater withdrawal (e.g., Atlantic City, New Jersey, USA) have exacerbated the effects of eustatic sea-level rise by increasing relative sea-level rise and leading to the submergence and loss of estuarine and wetland habitats. Natural processes – compaction of subsurface sediments, tectonic (crustal) movements, and sinkhole formation – augment these effects. In coastal areas of the Mid-Atlantic region, relative sea-level rise effects are significant, with the subsidence of thick Cretaceous to Holocene unconsolidated sands, silts, clays, and gravels being driven by thermoflexural (the combination of thermal subsidence and sediment loading offshore, producing flexural subsidence onshore) and compaction effects (Kominz et al. 2008; Miller et al. 2013); Boon et al. (2010) reported subsidence rates of –1.3 to –4.0 mm/year at Chesapeake Bay. Acute subsidence is also evident along the northern Gulf of Mexico (i.e., Louisiana coast) where sediment compaction and downwarping of the underlying crust contribute greatly to rapid coastline

retreat. Biotic effects of coastal subsidence in estuarine environments are often most evident in fringing wetland habitats.

Floatables/Debris

Estuaries are environments which accumulate large amounts of floatables and other debris because of their close proximity to coastal communities and the heavy recreational and commercial utilization (Kennish 2001a). This is a worldwide problem. Litter is discarded from boats or shoreline areas, and much debris also enters estuaries via stormwater discharges and river inflow. Plastics are particularly detrimental because they can damage power plant intake systems and other engineered infrastructure, seriously foul habitats, and pose a hazard to estuarine organisms. Fish, birds, turtles, and marine mammals can be injured or killed when entangled in fishing line, nets, packing bands, and other synthetic debris. Plastics, when ingested, commonly obstruct the digestive systems or air passages of these animals, causing their death. Furthermore, the plastics are not biodegradable and can remain entrapped in estuaries for many years, threatening biotic communities indefinitely. Because the use of plastics has increased dramatically over the past several decades (Kennish 2001a), they have created serious pollution problems in estuarine and coastal marine environments (Ribic 1998; Kennish 2015b).

Conclusions

Estuaries are affected by many anthropogenic factors that can compromise the structure and function of biotic communities, the integrity of habitats, and the condition of entire ecosystems. Twelve major categories of anthropogenic stressors have been identified in estuaries. These include: (1) habitat loss and alteration; (2) enrichment (nutrients, organic carbon, and thermal loading); (3) sewage and pathogenic inputs; (4) chemical contaminants; (5) human-induced sediment/particulate inputs; (6) dredging and dredged-material disposal; (7) human-altered hydrological regimes; (8) invasive/introduced species; (9) overfishing; (10) climate change effects; (11) coastal subsidence; and (12) floatables/debris. Anthropogenic impacts on estuaries are increasing because of ongoing human population growth and development in coastal watersheds and the heavy recreational and commercial use of these coastal systems. In addition, interactions between two or more anthropogenic stressors can be synergistic, creating more serious impacts together than when acting alone.

Estuaries are sites of intense human activity, and thus sound management strategies must be formulated to protect them. Best management practices are required to minimize anthropogenic impacts, and education programs must be developed to inform the public of their stewardship responsibilities to protect and improve estuarine systems. There also

needs to be a strong outreach initiative to interpret scientific findings and connect with coastal communities on the condition and health of estuarine environments. Finally, effective restoration projects must be instituted to revitalize degraded habitat and improve biotic communities.

Cross-References

- [Changing Sea Levels](#)
- [Dams, Effect on Coasts](#)
- [Demography of Coastal Populations](#)
- [Dikes](#)
- [Dredging of Coastal Environments](#)
- [Environmental Quality](#)
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- [Water Quality](#)

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Europe, Coastal Ecology

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Introduction

Europe occupies a little less than 7% of the total area of the world land surface. For its size its coastline is long, estimated at 143,000 km (Stanners and Bourdeau 1995). The European figure is derived from a map of 1:3 million scale and represents a conservative figure. The coastlines of Great Britain and Greece, for example, are 18,838 and 15,000 km, respectively, when the many islands are included. Though the figure for Great Britain is taken from measurements made at a scale of 1:50,000 and not directly comparable with the European figure they give an indication of the scale of the resource.

Europe's landmass is bounded by the Atlantic Ocean in the west and includes several enclosed seas (Fig. 1). The basic fabric of the coast is formed from rocks which span the whole of the geological timescale from the Precambrian period (3,000–570 million years ago) to the rocks laid down in the Cretaceous–Tertiary period (140–2 million years ago). Superimposed on this structure, in the north, is the effect of the ice-ages of the Quaternary period (2 million to 10,000 years ago) and the glacial deposits, which were left behind as the ice melted. In the south, tectonic activity associated with volcanoes and earthquakes has also had an impact, though on a more limited geographical scale.

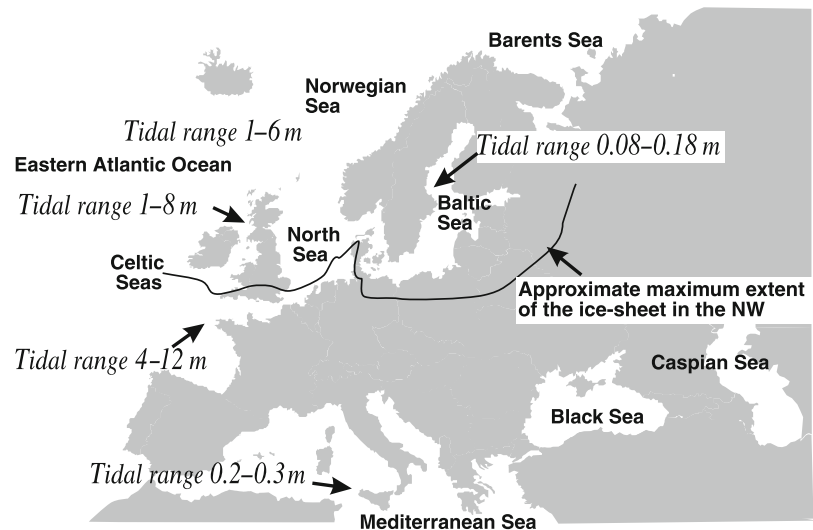
A combination of oceanographic factors (such as tides and sea-level change) as well as climatic effects, force the landward or seaward movement of sediments derived from erosion of the land or seabed. This provides opportunities for

new coastal areas to develop as the material is deposited on or adjacent to the land. A variety of other variables, including the ameliorating effect of the warm Atlantic waters and westerly winds help to define a west to east climatic gradient, which is most clearly seen in the nature of the vegetation. Human activities have also been important, though more recent in their impact, causing modification or loss of habitat and “squeezing” the coastal zone (Doody 2001a). The description that follows provides an overview of the nature of Europe's coastline from Norway to the eastern Mediterranean and the flora and fauna which occur there.

The “Nature” of Europe's Coast

The coastline and marine waters of Europe are areas of great contrast with a high diversity of wildlife and landscape features. The underlying geology helps to define the coast and its landscape. Today, in many areas where the coastline is composed of “hard” rocks resistant to erosion, it may appear unchanging. However, during the 2 million years or so of the Quaternary Period the growth (and retreat) of the ice sheets eroded mountains and scoured river valleys. In the north and west from Norway to the Atlantic coasts of Ireland and western Britain, the combined effects helped to mold the landscape, exposing some underlying structures and covering others. This glaciated northern landscape exists today as a highly indented coast with exposed sea cliffs, coastal inlets (including fjords, fjards and firths, and a few, usually small estuaries), rocky shores, and pocket beaches. By contrast, the coasts of southern England, France, northern Spain, and Portugal include “softer” though resistant limestone molded by more recent events (such as erosion from the sea and land slip) to form cliffs and headlands. Hard igneous rocks of Precambrian age and resistant granites also abound in the northern Baltic. Here, the lie of the land is gentler and cliffs are much less evident. Isostatic rebound continues to create small flat offshore islands. At the southern limits of the ice sheets, in the southern North Sea and the Baltic Sea particularly (Fig. 1), major deposits of glacial sediment left behind as the ice retreated northwards have helped to create the present low-lying landscapes. These provided abundant soft sediments for reworking during the Holocene epoch of the last 10,000 years. Today eroding glacial till cliffs, sand dunes, salt marshes and tidal sand, and mud flats are abundant. These combine to form large estuaries in the meso-to macro-tidal southern North Sea and deltas on the micro-tidal shores of the southern Baltic. Within the matrix of sandy spits and bars and tidal flats, promontories of harder limestone rocks help form embayments in a few locations. Sub-tidal areas also tend to be dominated by softer sediments on the sea bed.

The Mediterranean Sea has both hard and soft rocks, the former pre-dominating as karstic limestone cliffs, and more

Europe, Coastal Ecology,**Fig. 1** Europe's seas

gently sloping rocky shores. Generally, the land rises uninterrupted from the sea. Where cliffs occur these can reach a height of over 150 m, as on the southern coast of Spain. Over the last 4000 years or so erosion of the mountains has also helped to create large coastal plains as their river catchments have delivered sediment to the coast. In the micro-tidal areas of both the Mediterranean and Black Seas large deltaic systems such as the Ebro Delta in Spain, the Po Delta in Italy, and the Danube Delta in the Black Sea have grown partly in response to deforestation in the hinterland (Figs. 2 and 3).

Sea-Level Change, Tidal Range, and Wave Action

Long-term sea-level change is a major determining factor in the development of the coastline and coastal habitats. From around 7000 years ago, when sea level began to stabilize at its present level, many of the coastal formations that are present today began to develop. Their evolution then and now depends on a combination of factors, including the availability of sediment and the influence of sea-level change, tidal range, and wave action.

With global sea levels rising today at about 1–2 mm per year (Warrick 1993) there appears to be a general trend towards erosion and mobilization of sediments. Exceptions occur where isostatic rebound of glaciated areas is rising faster than sea level, causing a general rise in the land relative to the sea in areas such as Norway, the Gulf of Bothnia and around the margins of the Central Highlands of Scotland. Where the land continues to sink or is stable, relative sea-level rise tends to push coastal habitats toward the land. When this migration is restricted by resistant, rising landforms or artificial structures built to defend the land from erosion and/or flooding, the shore is squeezed into an ever narrower zone. This has important implications not only for the habitats themselves, but also the ability of the coastal landforms to

provide a flexible response to storms and extreme tidal events. It may also impact on human use and infrastructure at the coastal margin.

Tides and tidal range provide a further mechanism for change. Where large amounts of material are available for transport extensive sedimentary habitats can be created. In the macro-tidal areas (spring tidal range 4 m) of the west, including the Celtic Seas, the Bristol Channel (where the tidal range, at 12 m is the second highest in the world) and the Atlantic coast of France, sediments are moved inland to fill inlets and bays, creating estuaries. In meso-tidal (spring tidal range 2–4 m) or micro-tidal areas (spring tidal range 2 m) the tides have progressively less influence. At the lower end of the tidal range the sediment movement is offshore, with rivers tending to exert a more dominant role. Waves and wave energy help determine the final configuration of the deltas as in the Baltic and the Mediterranean.

Climate

The coasts and seas of Europe encompass a wide range of climatic conditions from the Arctic Circle to the deserts of north Africa. Summer temperatures rarely rise above 10 °C in most northern latitudes. This gradient is further influenced in the west by the marine waters, which are warmed by the Gulf Stream, creating an extreme maritime climate, which helps to define the Atlantic biogeographical zone encompassing much of northwest Europe (Polunin and Walters 1985).

Unlike the areas further north the climate of the Mediterranean is very different. The winters are warmer (averaging 6 °C) and the summers dry and hot. The range of temperature is much greater than on the Atlantic coast. The combination of thin limestone soils, dry climate and the long history of grazing and burning of the natural vegetation, has helped to create low-growing shrubby, drought and grazing resistant vegetation, which covers large areas of the Mediterranean.

Europe, Coastal Ecology,**Fig. 2** Distribution of the “cliffed” coastlines of Europe**Europe, Coastal Ecology,****Fig. 3** An elevated coastline (the Old Man of Hoy, Orkney Islands, Scotland)

Since storms and waves are generated over a smaller sea area than the Atlantic, extreme exposure to gales is less evident than on the Atlantic coast.

Elevated Landscapes: Sea Cliffs

Sea cliffs help to create some of the most spectacular and remote landscapes in Europe. They may have vertical or sloping faces, and in exposed locations are often drenched in salt spray, especially where they face the full extent of the Atlantic Ocean, and in the northern North Sea. Their orientation (and hence degree of exposure) and the nature of the underlying rock helps to determine the type of vegetation growing there. Although not considered in detail here stability is also an important factor helping to perpetuate shallow skeletal soils.

Sea Cliff Vegetation

Three main types of vegetation have been identified in the European Union Habitats and Species Directive, (European Commission 1999). A brief description of the first two is given below.

Vegetated sea cliffs of the Atlantic and Baltic coasts. In exposed areas with hard rock cliffs, some of the best examples of maritime cliff vegetation in Europe are to be found. These include communities dominated by roseroot *Sedum rosea* and Scot's lovage *Ligusticum scoticum*, on the salt-spray drenched cliffs. Above this spray zone, the cliffs support communities that include arctic-alpine plants including purple saxifrage *Saxifraga oppositifolia*, moss campion *Silene acaulis*, mountain avens *Dryas octopetala* and in Scotland the endemic Scottish primrose *Primula scotica*. In some areas, the extent of the truly maritime zone can be small. In sheltered locations, such as the east coasts of Ireland and Scotland, the cliffs are only marginally exposed to the full power of the wind, breaking waves and salt spray. Here, the maritime influence is reduced and plants less resilient to exposure and salt spray grow in communities more typical of inland locations. Further south from south west Ireland, Wales and south west England to southern Portugal, spectacular west facing cliffs also occur. These are less resistant than the harder rocks further north and in some areas rapid rates of erosion (1 m or more per year) can be measured as, for example, in the chalk cliffs of the Channel coast. Although still exposed to the Atlantic Ocean and its gales, climatic amelioration also allows a richer flora to develop. In limestone areas of south west Britain, northern France, and Spain the cliffs provide important refuges for warm loving species, such as spider orchid *Ophrys sphegodes*, formerly typically found in inland grasslands. In some areas, cliffs provide the most spectacular natural rock gardens, as at Cape St. Vincent in Portugal, where the endemic *Cistus palinhæ* is one of the many species present.

Vegetated sea cliffs of the Mediterranean coasts with endemic Limonium spp. Unlike the corresponding cliffs in the north and west, the incidence of exposure and hence the maritime nature of cliff vegetation in the Mediterranean is much reduced. Exceptions include the megacliffs of Croatia (Velebit) which represent a coastal habitat of exceptional biological importance. The habitats present range from emerging rocks with a calcifying algal community to sea water-drenched hyper-saline communities, including *Arthrocnemum glaucum* and *Atriplex portulacoides*, to exposed windswept cliff tops from 460 to 1200 m high. Interspersed in more sheltered areas are woodlands and scrub. There is little elsewhere in the literature concerning these habitats though reference is made to similar cliffs in Albania and southwest Turkey (van der Maarel 1993). These areas should probably be considered alongside the Atlantic woodlands of the northwest and the laurel forests of the Canary Islands for their conservation importance. The vegetation of all of these is to one degree or another dependent on its presence of moisture-laden air derived from strong onshore winds.

Sea Cliffs and Seabirds

Because of the close proximity to the rich waters of the Atlantic Ocean and the North Sea, coastal cliffs often support large colonies of nesting seabirds. The most common and conspicuous species are guillemot *Uria aalge*, kittiwake *Rissa tridactyla*, fulmar *Fulmarus glacialis* and razorbill *Alca torda*, which nest on ledges. Steep-sided granite cliffs may support large populations of gannet *Sula bassana*. On sloping cliffs where soil has accumulated, species which prefer to nest in burrows are found in high numbers such as puffin *Fratercula arctica* and manx shearwater *Puffinus puffinus*. Sea cliffs also provide nesting sites for a rare birds of prey such as the peregrine *Falco peregrinus* and the raven *Corvus corax*.

Seabird colonies do occur on cliffs and islands on more southerly coasts, though the populations, especially of cliff-nesting species are much smaller. Birds of prey are important and of these Eleonora's falcon *Falco eleonora* is of special significance. It is a small falcon feeding mostly on small birds. It migrates northwards from Madagascar where it is widely dispersed feeding on a range of insects and in a wide variety of habitats. These include coastal cliffs and cultivated areas, rice fields, lightly wooded slopes or coastal plains. It breeds in colonies on steep rocky coastal cliffs and islands, mainly in the Mediterranean. The breeding season is later than other species and timed to coincide with passerine migrations, ensuring a supply of food for its young.

"Soft" Rock Landscapes: Unconsolidated Cliffs

Cliffs derived from unconsolidated deposits of clay (including glacial material), loam or other easily eroded

sediments occur throughout the area. Landslip is common in sand and clays in the southern North Sea, southern Baltic Sea and on the south coast of England and in the sedimentary deposits of the Black Sea. Continuous erosion can result in the formation of high sloping unstable cliffs with little or no vegetation other than ephemeral communities. Where erosion is more episodic, the slopes can support scrub and even woodland on ledges remaining stable over periods of one hundred years or so. In these situations, some of the least “maritime” of the coastal cliff vegetation is found and may be indistinguishable from inland types.

The release of sediment during periods of erosion is an important factor in the development of coastal systems elsewhere. Preventing erosion in order to protect buildings or other human land uses can have consequences for habitats along the shore. As less sediment is available for movement by longshore drift opportunities for the creation of new beaches, themselves a form of coast protection, is restricted. In extreme cases adjacent habitats may be so deprived of sediment that they begin to erode.

Complex Sedimentary Systems

In contrast to cliffed landscapes, sedimentary landscapes are characteristically low-lying (Fig. 4). They can include a variety of very different habitats and are often highly unstable. Tidal flats and salt marshes, sand dunes, and gravel are the main formations that help to define the ecosystem. Coastal wetlands may include some or all of these habitats though the tidal habitats, especially sand and mud flats, have a special significance and are

described first. The short-to medium-term influence of tides and waves, freshwater inflows and sediment movement, or longer-term trends associated with sea-level change can result in the systems being very complex both spatially and temporally. Included within this section are the other components that make up the complex coastal sedimentary habitats, including the more terrestrial sand dunes and gravel structures.

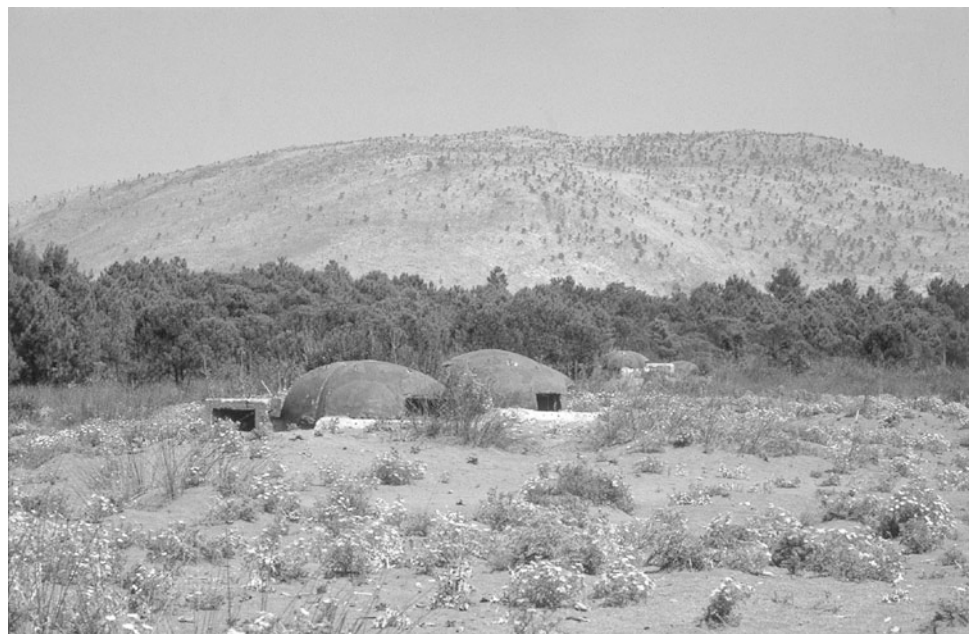
Coastal Wetlands

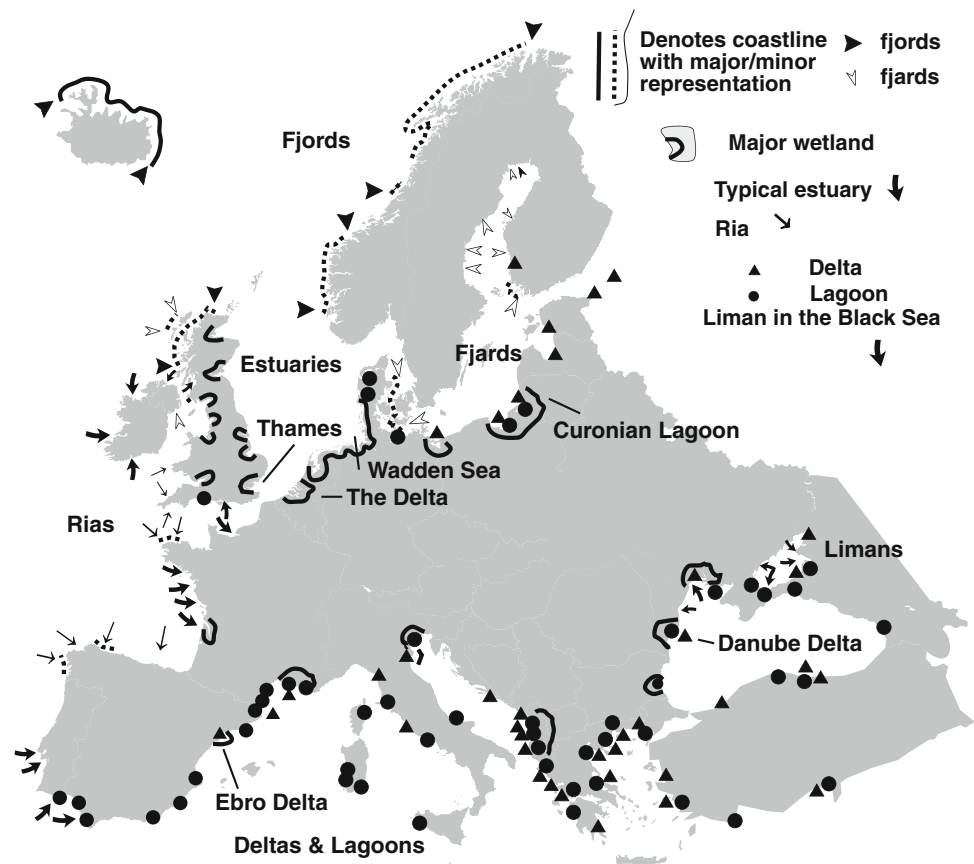
Coastal wetlands occur throughout Europe (Fig. 5). Wolff (1989) estimates that intertidal areas cover 9300 km², excluding small wetland areas in Iceland, Russia, Norway, Spain, Portugal, and Italy where little or no information is available. This implies that intertidal wetlands cover only 0.09% of Europe’s land surface, a figure clearly highlighting the scarcity of the habitat type. They are most extensive in the major estuaries of the southern North Sea and the deltas of the southern Baltic and Mediterranean Seas. The Wadden Sea in Holland, for example, has approximately 123,900 ha and the Ebro Delta in Catalonia, one of the five biggest in the Mediterranean/Black Sea area has some 350,000 ha of tidal and former tidal land.

Estuaries. Estuaries, deltas, and lagoons lie within a range of coastal wetlands which have specific characteristics defined partly by the geographical location and degree of sediment input. In the north and west fjords and fjards, usually considered to be a special form of estuary, are specifically associated with glaciated areas and have limited inputs of sediment. The former are drowned glacial troughs, often associated with major lines of geological weakness. They have a close width–depth ratio, steep sides and an almost

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Fig. 4 Low-lying sedimentary coastal plain derived from sediments eroded from the hills of northern Albania



Europe, Coastal Ecology,**Fig. 5** Distribution of coastal wetlands in Europe. (Adapted from Doody 2001b)

rectangular cross section. Sediment deposition is mostly restricted to the head of the fjord where the rivers usually discharge. The latter are glacial relicts on lowland coasts; ice-scoured rock basins, bars and small islands are characteristic features. Rias also have limited sediment and are formed when river valleys were inundated at the end of the last glacial period. They occur on the exposed western shores of Europe in macro- or meso-tidal areas. They are characteristically steep-sided with wooded slopes and occur in south west England and northern France, Spain, and Portugal.

Estuaries usually have major inputs of sediment. They are “partially enclosed bodies of water, open to saline water from the sea and receiving fresh water from rivers, land run-off or seepage” (Davidson et al. 1991). They are complex systems, subject to a usually twice daily tidal rise and fall, they may have sand dunes or gravel structures enclosing sheltered shallow mud and sand shoals and salt marshes. They are particularly extensive in lowland areas with abundant sediment and a macro-meso-tidal range. Particularly large sites occur in the southern north sea, (in and around the Thames Basin, the delta region of Holland and in the Wadden Sea).

Estuarine fauna. A key component in the development of the ecological value of coastal wetlands, especially tidal estuaries, is their high productivity. This is derived from in situ

primary production of salt marshes and in estuarine waters, as well as inputs from the land and the sea. This abundant source of food can result in very large numbers of a few dominant organisms. Typical densities on muddy shores of the mud snail *Hydrobia ulvae* can be up to 120,000 individuals present per square meter. The burrowing amphipod *Corophium volutator* is also numerous with up to 60,000 individuals per square meter. The segmented worms *Nereis diversicolor* (a ragworm) and on more sandy shores *Arenicola marina* (lugworm), are also often abundant and of particular importance as prey for wintering wading birds. In this context, estuaries in north west Europe are especially valuable and frequently support nationally and internationally important numbers of water birds (wildfowl and waders). Some indication of the significance of these estuaries can be gleaned from the location of the main wintering populations of the knot *Canutus canutus islandica* (Fig. 6).

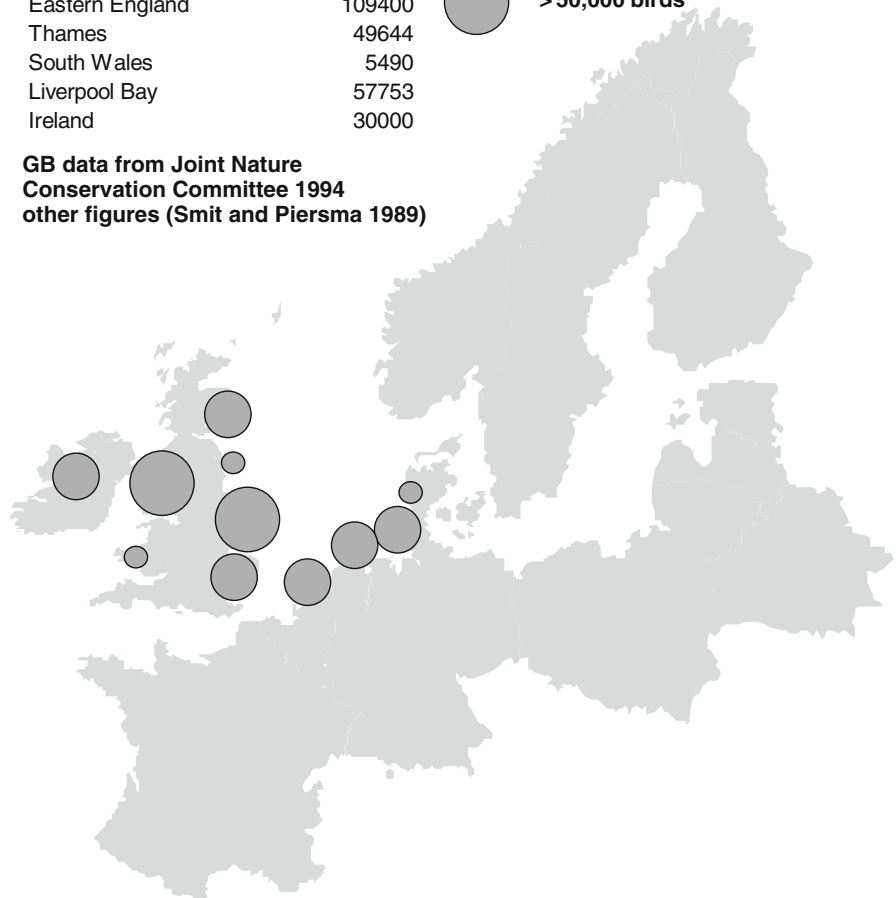
While these wintering bird populations are often the most obvious visible manifestation of the productivity of a wetland system a range of other species occur. These include sea fish using the shelter of the estuary to spawn, or as nursery areas. Diadromous fish are specialists of the estuarine environment. Catadromous species such as *Anguilla anguilla* (the common eel) move from freshwater to the sea where they spawn, passing on the way the salinity gradients in the estuaries and deltas.

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Fig. 6 Distribution of the main wintering sites of the knot *Canutus canutus islandica*. (Adapted from Doody 2001b)

Geographical Area	Nos	Major locations with:
Wadden Sea (Denmark)	200	up to 5,000 birds
Wadden Sea (Germany)	8000	
Wadden Sea (Netherlands)	50000	5–50,000 birds
Delta (Netherlands)	15000	
SE Scotland	11794	
NE England	3574	> 50,000 birds
Eastern England	109400	
Thames	49644	
South Wales	5490	
Liverpool Bay	57753	
Ireland	30000	

GB data from Joint Nature Conservation Committee 1994
other figures (Smit and Piersma 1989)



In contrast an anadromous species such as *Salmo salar* (salmon) develops from eggs shed upstream in rivers. A major part of its life is spent outside the home rivers at sea. Important feeding areas are in the Atlantic along the west coast of Greenland, above the Faroe Islands, in the Norwegian Sea, and possibly in the Barents Sea. In the Baltic, the Curonian Lagoon has 46 freshwater, migratory, and marine fish including the Baltic salmon, reflecting its importance for these species.

Deltas and Lagoons

Coastal deltas and lagoons are most characteristic of micro-tidal regions (with a tidal range < 2 m), and hence in Europe are particularly abundant around the shores of the Baltic, Mediterranean, and Black Seas. Deltas are sedimentary coastal plains built up from material brought down by rivers to cover shallow offshore areas creating sedimentary features which protrude beyond the normal limit of the coastal margin. They are best developed in micro-meso-tidal areas

where wave action is the dominant force in building the sand dunes, barrier islands bars, and salt marshes. Lagoons are shallow, virtually tideless, pond-or lake-like bodies of coastal salt or brackish water that are partially isolated from the adjacent sea by a sedimentary barrier, but which nevertheless receive an influx of salt water from that sea (Barnes 1980; Guélorget and Perthuisot 1992). Although they are mostly found in micro-tidal seas, they are also present on some macro-tidal coasts – such as the northern North Atlantic – where offshore deposits of gravel are to be found. Gravel then replaces the sand of micro-tidal seas as the barrier material. Europe possesses the smallest proportion of lagoonal coastline of any continent, with only 5% of its coast in this category (Cromwell 1971). Several deltas, for example that of the Danube (and Nile), have developed within – and have now largely or completely obliterated – former lagoons enclosed by barrier-island chains. The limans of the Black Sea coast receive their salt input via seepage.

Fauna. Deltas and lagoons are also important for a variety of water birds. They support waders, wildfowl, grebes, and similar marshland birds many wintering or on passage, as well as a considerable number of nesting species. Among these black-winged stilt *Himantopus himantopus*, avocet *Recurvirostra avosetta* and herons, including purple heron *Ardea purpurea* and large populations of a few species of breeding duck are significant. However, it is the spectacular flamingo *Phoenicopterus ruber roseus*, which is most readily appreciated by anyone visiting the Camargue or the Ebro Deltas in the western Mediterranean. Numbered in thousands in the Camargue it is a species dependent on wetlands which periodically dry out and become hypersaline. These conditions naturally occur in North Africa and the Mediterranean but it is the juxtaposition of “natural” lagoons with “artificial” salinas which provide especially suitable nesting conditions (Sadoul et al. 1998).

The lagoons also support invertebrate and fish faunas. These may include essentially freshwater or marine/estuarine species capable of withstanding a degree of brackishness as well as specialist lagoon species. Among the marine fishes are sea bass *Dicentrarchus labrax* and sole *Solea vulgaris* while carp *Cyprinus carpio* and pike *Esox lucius* live in freshwater. Eel *A. anguilla* and mullets (Mugillidae) bridge the salinity gap.

Salt Marshes

Salt marshes include all halophytic plant communities and their associated animals, which are influenced by sea water. Typically they are regularly inundated by the tide, have sometimes rapid accumulations of fine sediment and in the absence of enclosure, include transitions to nontidal vegetation. They are at their most extensive on flat tidal plains in sheltered locations where abundant sediment is deposited and wave action restricted (estuarine marshes). These conditions are most favorable to the establishment of the pioneer salt marsh plants such as *Salicornia* spp., *Suaeda* spp. and *Spartina* spp. which are among the most frequent colonists in northern latitudes. Lagoonal marshes occur in association with lagoons and are most extensive in the micro-tidal areas. Salt marshes also occur on beach plains behind exposed or partially protected beaches and in the lee of barrier islands. In some areas where sediment is restricted or (as is the case on some rocky shores and cliff tops) absent, inundation is intermittent and dependent on storms or extreme high water levels. In these situations, “perched” salt marshes on exposed cliffs and narrow saltings of rocky shores “beach head” and “loch head” salt marshes have vegetation similar to the typical salt marsh but without their rapid sedimentation (Doody 2001a).

Distribution in Europe. Salt marshes occur throughout Europe and are particularly associated with the major macro-tidal estuaries of the southern North Sea and south and west England (Dijkema 1984). Elsewhere in the northwest they occur as scattered small sites associated with rocky coasts. In the Baltic, they are more restricted, partly because

of the very low salinity within this sea (Dijkema 1990). In the Mediterranean, all the major deltas have extensive salt marsh communities. Elsewhere they again occur as scattered sites, though here less frequent than in the north. Their distribution is closely related to that of the coastal wetlands.

Vegetation succession. A key element in the development of salt marshes in sediment rich areas is the process of succession. A sequence of communities tolerant of different degrees of tidal submergence may develop, together with transitions to other, terrestrial vegetation. Transitions may be to reed swamp sand dune, gravel, freshwater, or woodland. This natural transition may be truncated by the construction of sea walls or other coastal defenses, with the loss of these upper salt marsh transitions and their associated wildlife. In some areas where a saline influence continues landwards of sea walls on enclosed unimproved salt marshes, coastal grazing marshes with brackish ditches may develop.

The early stages of colonization are represented by two main community types (European Commission 1999), namely: ***Salicornia* and other annuals colonizing mud and sand** and ***Spartina* swards (*Spartinion maritimae*)**. These are typically referred to as pioneer marsh colonizing the lower levels of the tidal flats, most frequently inundated by the sea. Under these conditions, areas of mud, exposed for several days, are required for the early colonizers to become established. The communities which develop are composed of a few, mostly annual species occurring in mono specific stands, or in simple combinations. These typically include *Spartina* communities, most frequently *Spartina anglica* stands of *Salicornia* spp., *Suaeda maritima* and *Aster tripolium* with *Puccinellia maritima* dominant. They are most frequently found in the macro-meso-tidal areas of the northwest.

The next stage in the successional process is typically referred to as mid-marsh. The distinction between low and mid-marsh is blurred but as the tidal height of the marsh increases additional species occur. These are dominated by the shrub *A. portulacoides* in the absence of grazing with increasingly *Festuca rubra* on grazed marshes. There is a wide variation in the composition of these communities and *Juncus gerardii* is another characteristic species. The presence of abundant *Armeria maritima* gives many marshes in the northwest a visually distinct appearance with a close-cropped, pink sward in summer.

The development of upper-marsh in the absence of enclosure provides some of the richest plant communities. Mid-upper marshes in the northwest are collectively classified within community type. **Atlantic saltmeadows (*Glaucopuccinellietalia maritimae*)**. The presence of a number of species such as *Suaeda vera*, and *Frankenia laevis* in southeast England, more frequently found in Mediterranean regions suggest an overlap in distribution with. **Mediterranean salt meadows (*Juncetalia maritimi*)** and

Mediterranean and thermo-Atlantic halophilous scrubs (*Sarcocornetea fruticosi*). Transitions to swamp also occur in the absence of enclosure and are usually dominated by *Scirpus maritima* and *Phragmites australis*, which can be extensive in the upper reaches of estuaries. These communities form important transitions and may be particularly rich, especially where freshwater seepages or transitions to other habitats such as sand dunes occur. In the Baltic, the low salinity and micro-tidal range of the sea restricts the presence of the pioneer communities and extensive swamps dominated by *Phragmites* are more prevalent (Dijkema 1990). Similarly in the Mediterranean the extent to which a true succession develops is restricted and narrow zones of a few mid-marsh species quickly merge into more extensive swamp. Some of the best examples of these occur in the still relatively undeveloped coastal deltas of northern Albania.

Fauna. Salt marshes present difficult environments for colonization by invertebrates due to changes in salinity and humidity caused by periodic tidal immersion. The fauna is a mixture of marine, freshwater and terrestrial species, which are adapted in various ways to the stressful environment. Marine species tend to occur lower down the marsh and often burrow to avoid desiccation. Terrestrial and freshwater species occur mostly in the upper marsh and transition zones, and have adapted to, or avoid immersion in saline water. The presence of individual invertebrate species is often closely associated with the type of vegetation which occurs.

Successful breeding by a variety of birds occurs on the upper levels of unenclosed salt marshes following the high spring tides in April and May. Although early nests may be destroyed by high tides, at least one brood is usually possible. The structural diversity of the marsh is also of crucial importance to some species such as redshank *Tringa totanus* and particularly high populations may be present in the best habitats. In winter, many of the most numerous species of wader feed almost exclusively on the invertebrates in the mud flats at low tide and are not dependent on the salt marshes as their main food supply. However, some also use the salt marsh for roosting at high tide, or as supplementary feeding around the pans and bare areas on the marsh (e.g., dunlin *Calidris alpina* and knot). Only the redshank and snipe, of the wading birds, use salt marsh in preference to the exposed tidal flat, probably because they like to be sheltered from view by the cover of vegetation. For these species the salt marsh represents the primary feeding location throughout most of the tidal cycle.

Other species, mainly grazing ducks and geese use the salt marsh as a source of food. The geese seek out open areas with close-cropped or low vegetation. Even here there are distinct preferences, and while the brent goose feeds on Eel-grass (*Zostera* spp.) in the low marsh, the stronger grey-lag goose feeds on the tougher grasses of the marsh itself. Of the ducks the wigeon feeds on both low and high marsh, normally at the

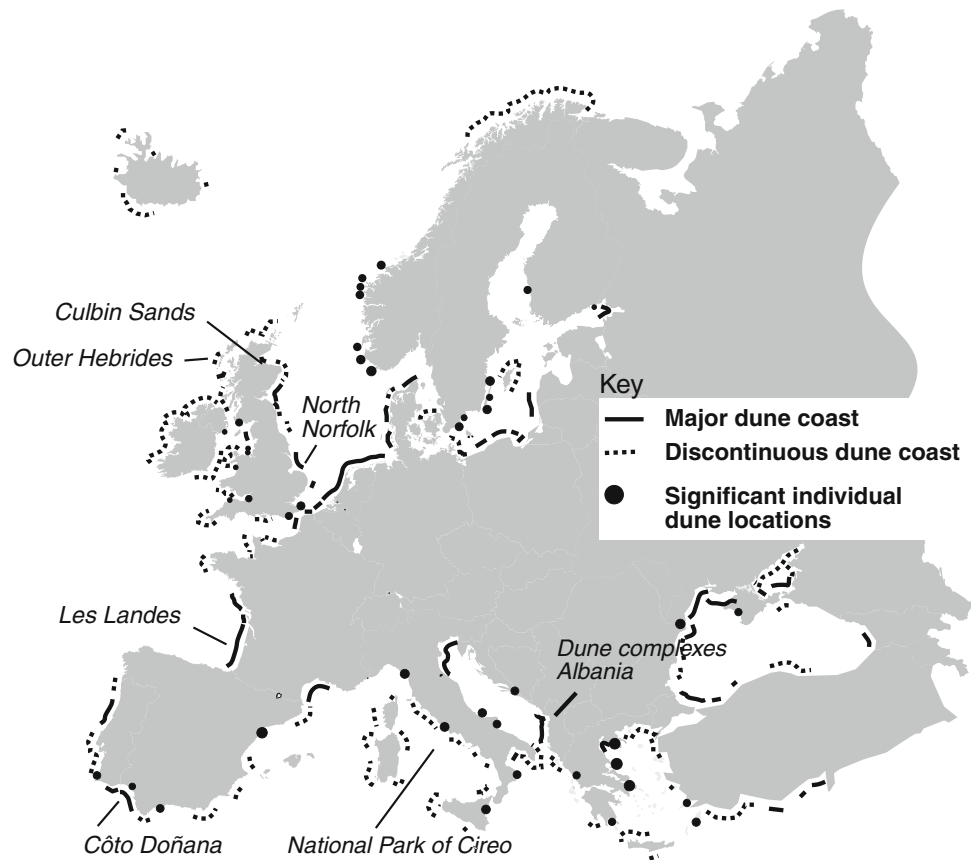
waters edge. For most of these species the more palatable grasses favored under high stocking densities of cattle and sheep, are important to the continued use of the salt marsh by the grazing ducks and geese.

Sand Dunes

Sand dunes develop wherever there is a suitable supply of sediment (within the size range 0.2–2.0 mm) which is moved onshore by the tide and then blown inland to form accumulations varying from a few centimeters to 40 m or more thick. The type of sand dune which develops depends on the underlying topography and climate. In most locations in the temperate regions of the world, vegetation plays an important role in the growth of the typical dune landscape, facilitating the accumulation of sediment (Ranwell and Boar 1986). Bay dunes are the most common form and are associated with indented coastlines, often between headlands. Backshore dunes can be large and occur when both dominant and prevailing winds blow sand inland. On especially exposed sites, coastal cliffs may be covered with a veneer of sand blown up over the cliff to create “climbing dunes.” Barrier islands tend to occur in exposed locations and may be built on gravel bars. Ness dunes, including cusped forelands, occur where prevailing and dominant winds are in opposition. Sometimes progradation may be quite rapid and a series of low dune ridges form, interspersed with wet hollows. These are also found in areas where sea level is rising relative to the land as in southeast Norway, northeast Scotland, and Northern Ireland. Spit dunes occur when sediment is deposited at the mouth of an estuary. They take a variety of complex forms depending on the action of waves (longshore drift) and the force of the river flows to the sea. These also include the dune forms associated with deltas in micro-tidal seas.

Distribution in Europe. Sand dunes are distributed throughout Europe (Fig. 7). In the areas where sediments are restricted there are a relatively large number of small sites, as in northern Norway and on the rocky, cliffed coasts of the exposed shores of the Atlantic from western Ireland to Spain, the karst coasts of the Mediterranean, and in the Black Sea. In the west where there are abundant coastal sand deposits, prevailing westerly winds blow the sand grains onshore to create large backshore systems such as those of the Outer Hebrides in Scotland. These include some of the best and largest examples of the extensive cultivated sandy plain or machair, also present in the west of Ireland. In northern Denmark massive accumulations of sand are forced onshore under the action of the prevailing winds from the North Sea. In France on similarly exposed coasts the dunes stretch many kilometers inland. Here, the area known as “Les Landes” has a special significance, and is one of the most extensive dunes in Europe.

On the North Sea coast and in the northern Baltic, dunes develop on a coastline, which is rising relative to sea level.

Europe, Coastal Ecology,**Fig. 7** Distribution of sand dunes in Europe. (Adapted from Doody 2001b)

Here, they are generally prograding with ridges becoming established as a sequence lying “parallel” to the coast. They are sometimes interspersed with damp hollows in which dune slack vegetation develops. In eastern Scotland some large examples occur, as at Culbin Sands a system with in excess of 4000 ha, one of the largest areas of blown sand in Great Britain.

In the southern Baltic and southern North Sea where glacial sediments are particularly abundant, sand dunes are also extensive, for example, they make up 80% of the coastline of Poland. The predominant dune type is represented by sand bars and spits, which lie parallel to the coast (Poland and Wadden Sea). In southern Portugal and Spain, the dune systems also include barrier islands and spits. Rivers bring the sediment to the sea, which is then reworked by wave and tidal action and blown onshore. One of the most important dune systems in Europe, lies within the Doñana National Park in Southern Spain. The park consists of beaches, foredunes, high mobile dunes (up to 30 m), and stabilized dunes which enclose a major wetland.

On the Mediterranean and Black Sea coasts dunes are often found in association with deltas. In Italy, dunes probably originated during the thermal optimum after the last glaciation (5000 years ago). They have since been broken by rivers and their subsequent development has been a product of natural

erosive forces and human activity. Today dunes are present around the whole coastline but only in a few protected areas, like the National Park of Circeo, can natural development be seen. Along the Greek coast there are many places where sand dunes cannot develop because the hills or mountains outcrop as rocky coastlines near to the sea. Sand dunes tend either to occupy a narrow fringe bordering flat areas of land or exceptionally form extensive dunes up to 10 m height, as in Western Peloponnesus.

Vegetation succession. The vegetation of sand dunes in temperate regions such as Europe can be grouped according to the successional sequence from the beach inland. These have been defined for the north west coastal areas as follows:

Strandline: This includes vegetation along the high tide line.

It is usually ephemeral, salt tolerant and composed of a limited number of species;

Foredune: The first stage in sand deposition occurs here. The vegetation is limited in species diversity, dependent on its ability to withstand the influence of salt spray and trap moving sand;

Yellow dune: This represents the main and usually most rapid phase of dune growth. *Ammophila arenaria*, the main dune forming species, is able to withstand rapid burial by sand;

Dune grassland: Stabilized dune vegetation dominated by species of grass and herbs develops under the influence of grazing and in a moist climate. It is mostly found in northwest Europe;

Dune heath: Occurs on sand dunes where the calcium carbonate content of the soils is low, either because the original sand has a high proportion of silica or where leaching has taken place;

Scrub: Areas with low shrubs. These include woodland understorey species in the northwest and the “maquis” and other similar vegetation in the Mediterranean;

Dune slacks: Vegetation which develops under the influence of high water table and is often completely flooded in winter;

Woodland: Natural forest with various pine species or deciduous trees such as oaks. In most areas natural woodland is scarce.

The European Habitats Directive recognizes three major types of vegetation (European Commission 1999). Of these the dunes of the “Atlantic, North Sea and Baltic” include at least 11 distinct types and those of the Mediterranean coast, 7. Most of the sand dunes described for northwest Europe appear to have a sequence of vegetation types which potentially includes all the more important successional communities from strandline (driftwalls) to yellow and grey dune, dune pasture and scrub, described above. (In areas where beach erosion is occurring some of the early stages of succession may be absent.)

The sequence of vegetation on the mid-Atlantic coast is similar to that occurring in the northwest. In the north of France, some of the botanically richer areas occur where dunes are composed of calcareous sand and lie against to chalk cliffs. The further south the dunes are formed, the more the southern elements of the flora appear.

On the Mediterranean coast the zonation described above is much less obvious. While *A. arenaria* still plays an important part in dune establishment, communities with dry grasslands and scrub are more prevalent. The following are recognized from the Manual of European Union Habitats (European Commission 1999).

Crucianellion maritimae fixed beach dunes
Dunes with *Euphorbia terracina*
Malcolmietalia dune grasslands
Brachypodietalia dune grasslands with
annuals
Coastal dunes with *Juniperus* spp.
Cisto-Lavanduletalia dune sclerophyllous scrubs
Wooded dunes with *Pinus pinea* and/or
Pinus pinaster

Fauna The dune fauna is not especially rich or abundant though the variety of slope, aspect and vegetation structure, coupled with the dynamic nature of many systems, make dunes one of the most suitable habitats for a range of

invertebrates requiring open conditions. They are also frequented by warmth-loving species, which bask on the dry warm slopes even in the more northern latitudes. Hymenoptera (especially wild bees, digger wasps, spider wasps, and other solitary wasps) are able to build nests easily in the soft sand and are often found dwelling in sand dunes (Haeseler 1989, 1992).

The dune avifauna includes breeding colonies of terns such as the sandwich tern *Sterna sandvicensis* and the little tern *Sterna albifrons* which nest on open sand. Other species such as the eider duck *Somateria mollissima* and the shelduck *Tadorna tadorna* nest in the ground vegetation in the north. Further south (including eastern Spain) tree nesting species of heron, such as the squacco heron *Ardeola ralloides* and the purple heron *Ardea purpurea* can be found.

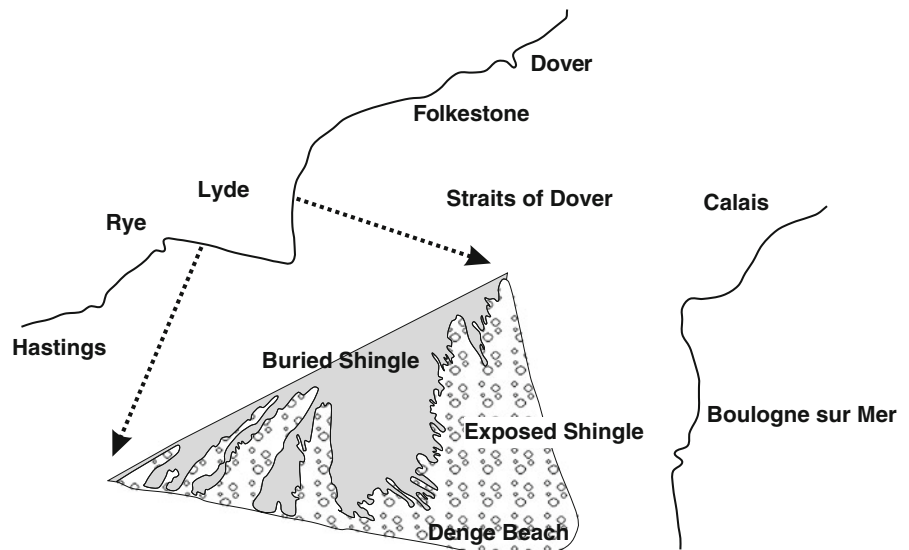
Rabbits *Oryctolagus cuniculus* occur in abundance in all areas and are stock food for fox *Vulpes vulpes* populations of which inhabit mainland dunes. Polecat *Mustela putorius*, weasel *Mustela nivalis*, several species of bats, voles, mice, and shrews also occur in most of the areas, while the roe deer *Capreolus capreolus* and red squirrel *Sciurus vulgaris* are less frequent. Among reptiles and amphibians are sand lizard *Lacerta agilis* natterjack toad *Bufo calamita* and the edible frog *Rana esculenta*.

Gravel

Gravel is the term applied to sediments larger in diameter than sand but smaller than boulders. A predominant particle size of over 2 mm separates gravel from sand (King 1972) whereas at much over 200 mm diameter, boulders approximate to cliff habitat. Four environmental factors are responsible for the growth of a gravel beach. There must be an available supply of sediment in the size range mentioned above. At the same time waves, wind, and tidal currents should be favorable for its movement. Since this coincidence is unpredictable, considerable periods of time may exist when no movement takes place, these are interspersed by times of marked activity resulting in stable and mobile gravel habitats varying both in time and space.

Geomorphologists and ecologists have recognized five categories of gravel structures which vary in their oceanicity and therefore in their ecology. The simplest and commonest type is the fringing beach forming a strip in contact with the land along the top of the beach. Gravel spits form where there is an abrupt change in the direction of the coast. They often contain recurved hooks and a recurved distal end, a result of deflection of waves by refraction. Bars or barriers are effectively spits which have formed across estuary mouths or indentations in the coast. These three structures are basically foreshores, regularly washed by spray and storm waves and therefore possessing only limited or ephemeral vegetation over much of their area. Cuspate forelands and offshore barrier islands, on the other hand, are larger structures

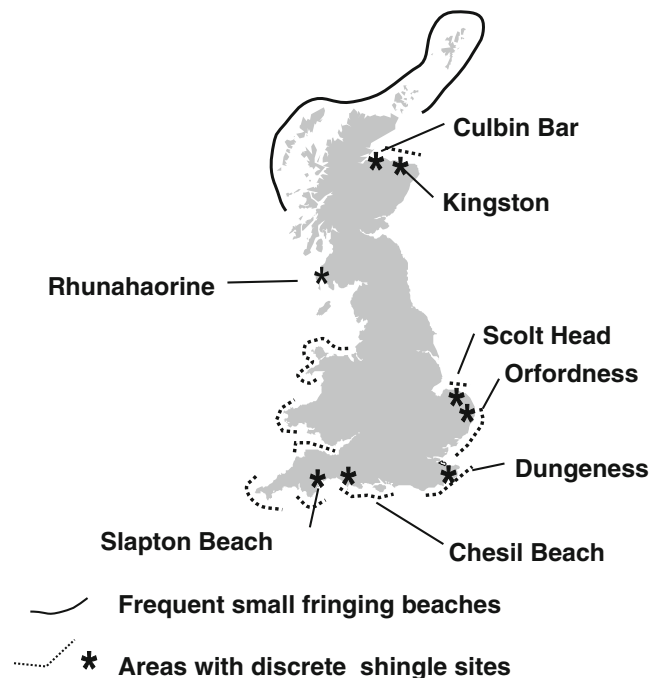
Europe, Coastal Ecology,
Fig. 8 Dungeness southeast
 England – a cusped foreland



and are more terrestrial in nature. The former develop when gravel is available in large quantities and piles up in front of fringing beaches or spits and is then driven landwards by storm waves to form a series of roughly parallel ridges. Where wave approach can from two directions only, such apposition beaches will form into cusped forelands (Fig. 8). Large masses of gravel may also form offshore barrier islands under conditions of shallow water and low-energy environments such as in the North Sea and the Baltic.

Distribution. Gravel deposits occur throughout Europe, especially in the north and west. Fringing beaches are widely distributed, though often small and narrow. Larger gravel structures occur as isolated and discrete entities. (Doody 2001b). In the north, the fjords and sea lochs in both the meso-tidal and micro-tidal areas of the Atlantic, North Sea, and Baltic Sea support long stretches of gravel foreshores. Individual, more substantial structures are scattered throughout the area. The situation in the United Kingdom illustrates the nature of this occurrence (Fig. 9). This is derived from extensive surveys undertaken in the 1990s and organized by the statutory conservation body, the Nature Conservancy Council.

Vegetation. Two plant communities identified in the European Union Habitats Directive (European Commission 1999) help to define these formations: **Annual vegetation of drift lines** and **Perennial vegetation of stony banks**. Gravel shore vegetation includes *Mertensia maritima* prevalent in the north and *Glaucium flavum* and *Matthiola sinuata* in the south. Mobility is an over-riding consideration and the species which colonize are able to withstand periodic disturbance. These include *Lathyrus japonicus* and *Crambe maritima* colonizing the gravel just above the high water mark where the shore may remain stable for periods of 3–4 years. *Crambe maritima* is constant in the three sub-types of **Perennial**



Europe, Coastal Ecology, Fig. 9 Distribution of gravel shores in the United Kingdom, some key sites are named

vegetation of stony banks, for the Baltic, Channel coast, and the Atlantic, respectively. Besides stability (or its converse mobility) beach composition and especially the amount of fine material present in the gravel matrix is another principal factor influencing the type of vegetation namely:

1. Gravel without matrix (relatively uncommon) – Vegetation is limited to encrusting lichens and the most tolerant angiosperms such as *L. japonicas*;

2. Gravel with a sand matrix – Dry yet stable conditions, contains species such as *Lotus corniculatus* and *Honckenya peploides*;
3. Gravel with a silt–clay matrix – Ecologically related to salt-marshes but with freer drainage. The vegetation is similar to that of saltmarsh levées with *A. portulacoides* or *Glaux maritime*;
4. Gravel with an organic matrix – Rotting seaweed is nutrient rich and supports populations of species that arrive as seed within tidal debris. Species include *Beta maritima* or *Atriplex* spp. where there is little drainage from the land, or with more humic soils reed *Phragmites communis* or *Iris pseudacorus* swamp may be present.

Fauna. The fauna associated with gravel structures largely consists of invertebrates (Shardlow 2001) and avifauna. In the case of the former large numbers of species, such as bumble bees are found. Gravel beaches and structures provide suitable habitat for a small number of other species. These are generally ground nesting birds (Cadbury and Ausden 2001) which make a simple scrape where they lay their camouflaged eggs. The importance of gravel as a habitat for breeding terns has led to the conservation of several sites and the need to create areas protected from human trampling affects. Few vertebrates other than domestic herbivores are found on gravel sites except hares and rabbits, though in some places foxes are important predators of ground nesting birds.

Human Influences

This entry along with many text books describe coastal habitats in terms of their ecological or geomorphological development. Salt marshes and sand dunes, in particular are described as exhibiting primary succession. Classic studies in Great Britain covered salt marshes and sand dunes on the North Norfolk coast. All emphasized the natural status of the vegetation in areas where ecological processes appear to have been relatively free from human influence. The vegetation was thus described as a series of community types progressing from the early pioneer stages to more complex forms related to the physical factors influencing their development. While this is in part true, it is also the case that much of Europe's coastline has been influenced in one way or another by human activity. A few examples are discussed below.

Sea Cliffs and Slopes

Apparently inaccessible steep-sided sea cliffs and slopes are often seen as virtually unaffected by human activity. Certainly difficulties of access, low productivity, associated with a harsher climate mean that the harder rock of the north are less built on than their counterparts in some parts of the

Mediterranean. However, grazing is and has been a traditional activity and even cliffs on more inaccessible islands have been and in a few cases continue to be plundered for sea birds and their eggs.

Sand Dunes

Major tourist development, especially along the Mediterranean coast, has destroyed many sand dunes. Building directly onto the beach in some cases has completely obliterated some areas while sand extraction for use nearby has also taken its toll. Another activity is planting with non-native trees, especially conifers, to stabilize them. This has affected many systems throughout Europe and while the dune form is retained most of the dune plant and animal species are lost. One exception to this general picture is the large dune systems which have developed in association with major deltas on the coastline of Albania and in some parts of south east Turkey. Here, many different dune forms are present with a maximum height of about 50 m and width of up to 4.3 km, including huge beach plains with embryo dunes, parabolic dunes, blowouts, dune slacks, lakes, secondary barchans, and dune fields. They also include complete transitions to pine woodland with native *P. pinea* and *Pinus halipensis* as the forest tree. They are among the most important dune areas in Europe.

Secondary Habitats Coastal Grazing Marsh and Salinas

The partial enclosure of salt marsh by the erection of a small earthen bank was probably one of the first human activities directly affecting coastal habitats. The aim was to extend the period of summer grazing by livestock (cattle and sheep), by preventing over-topping on spring tides. As the early techniques of enclosure improved, larger clay or earth banks were constructed enclosing consolidated salt marsh excluding the tide from larger areas. Historically the land was subsequently used as pasture, with little or no further "improvement". These areas of coastal grazing marsh are defined by the presence of permanent and semi-permanent grassland, drainage ditches, and enclosing earth dikes. The wildlife interest has developed alongside the traditional use of the area for agricultural.

The production of salt is one of the oldest industries known to man. Industrial salinas are large, artificial, and complex lagoon systems and a major source of salt produced today. The principal aim of any salt company is to produce a maximum tonnage of salt per annum to meet the demands of commerce and industry. However, these artificial ecosystems are in fact, wetlands of international importance and areas with a rich and diverse flora and fauna, vertebrate and invertebrate communities. They host important breeding populations of aquatic birds, besides rare and endangered species. The special nature of Mediterranean salinas make them vitally important resting and refuelling sites for many

thousands of shorebirds migrating between the Palearctic breeding grounds and their winter quarters in Africa (Sadoul et al. 1998).

Grazing and Other Management

A key feature of many habitats throughout Europe is the long history of exploitation by human kind, including grazing by domesticated stock. This has had a profound influence on the type of vegetation which has developed. In the northwest grazing (by rabbits cultivated in warrens) helped to create species rich calcareous dune grassland and heathland, preventing the natural progression to scrub and woodland. In areas formerly used for grazing by domestic stock a reversion is taking place to secondary mixed scrub, broad-leaved woodland and pine forest. This is particularly prevalent in Finland today where invasion of open land by pine forest with *Pinus sylvestris* is common. Prior to this, open vegetation was maintained by the grazing of domestic animals.

Grazing and burning are intimately bound up with the development of the low, thorny scrub vegetation which predominates in the Mediterranean. It is also clear that many of the major deltas would not be as extensive without the influence of deforestation in the hinterland. Thus it can be seen that the rich diversity of habitats and associated species, which exist along the coastline of Europe are influenced to a greater or lesser extent by human activities. At the other extreme major developments including tourist growth in the Mediterranean has greatly reduced the ecological and landscape diversity of many areas to the detriment of wildlife and cultural heritage. The way in which this has altered the coastline of Europe and the approaches being adopted to monitoring the impact of development and policy initiatives designed to protect it is the subject of the entry on Monitoring, Coastal Ecology.

Cross-References

- Coastal Climate
- Dune Ridges
- Estuaries
- Europe, Coastal Geomorphology
- History, Coastal Ecology
- Monitoring Coastal Ecology
- Rock Coast Processes
- Salt Marsh
- Vegetated Coasts

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Europe, Coastal Geomorphology

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well as coastal cementation like eolianites and beachrocks. Because of the rather long and complicated geologic and tectonic history of the European continent, the coastlines are greatly differentiated. Their total length in natural scales is several 100,000 km. Many of the scientific terms in coastal sciences derive from European examples, like ria, fjord, canale, tombolo, or Thyrrenian, Sicilian, and Calabrien for Pleistocene sea-level highstands.

Definition

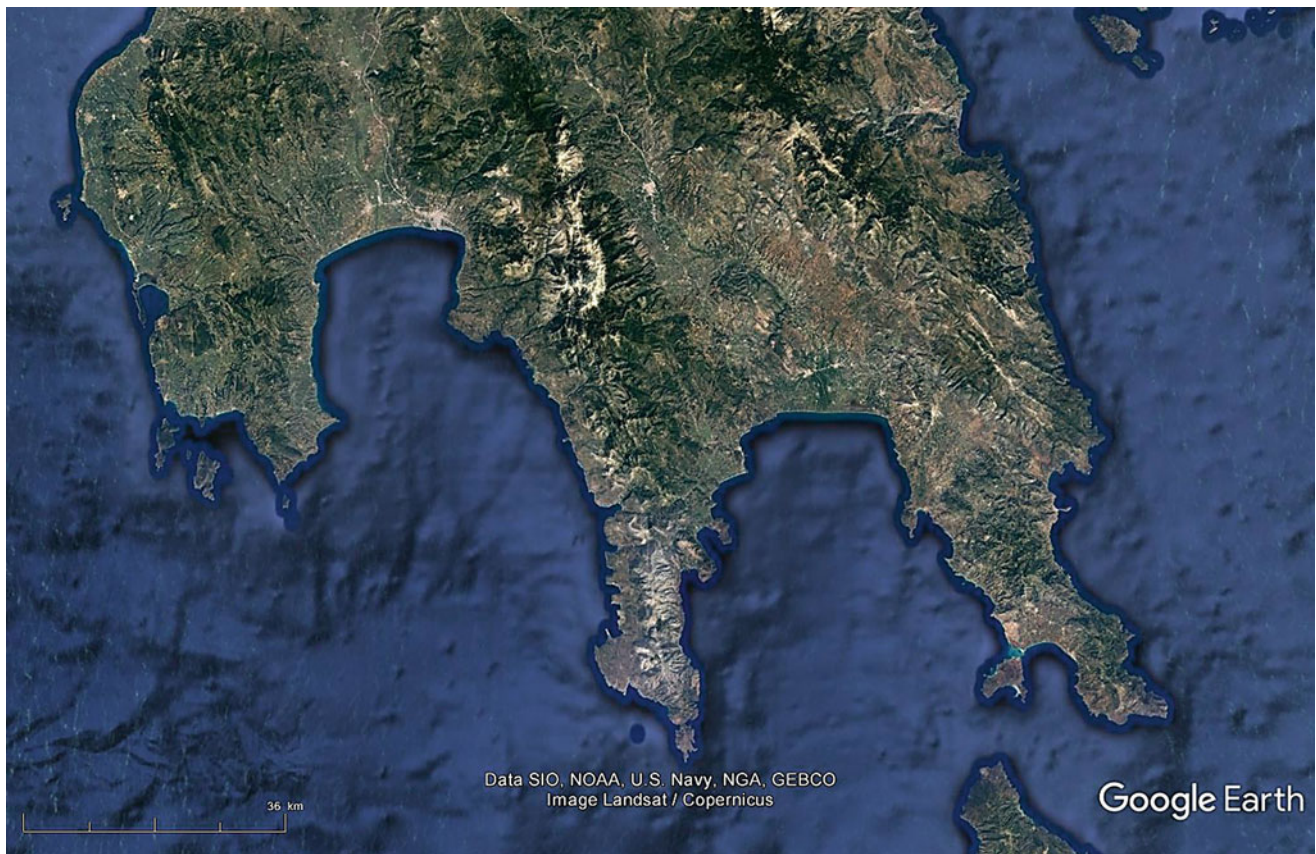
The coastlines of Europe cover a latitudinal range of 45° reaching from 80°N in Spitsbergen to 35°N in southern Crete, Greece. Thus, nearly all coastal types except for mangroves and coral reefs are present in Europe, including arctic ice coasts with calving glaciers, drifting sea ice contact and permafrost decay along cliffs on Spitsbergen, or young volcanic coastlines on the Aeolian Islands of Italy or Santorini/Greece. Processes of salt weathering associated with tafoni are present at the Costa Brava in Spain and the islands of Corsica and Sardinia or SW Turkey, and biogeneously formed coastal forms (Kelletat 1997) including vermetid boiler reefs and microatolls are characteristic on Crete islands (Greece) as

The Framework of Coastal Forming in Europe

The geology of Europe varies from very old structures, the Proterozoic shields of granite and gneiss in Scandinavia, to rather young like the vivid neotectonic areas in the eastern Mediterranean (Pirazzoli et al. 1996). The orogenic belts of the Caledonides are still visible in the coastal configuration of Scotland (Fig. 1), the Variscids in southern England, and the imprints of the alpidic orogenesis, can be observed along nearly all coastlines of the Mediterranean (Figs. 2 and 3). Areas of longer subsidence are marked by large embayments like the North Sea basin, the Bay of Biscaya, or the intracontinental sea of the Baltic and the intercontinental sea of the Mediterranean. Extremely resistant silicate rocks



Europe, Coastal Geomorphology, Fig. 1 Ingression in geologic structures: Caledonian fault in Scotland (Credit: Google Earth)



Europe, Coastal Geomorphology, Fig. 2 Horst and graben structures dominate the coastal configuration along the southern part of the Peloponnese/Greece (Credit: Google Earth)

survived thousands of years of wave attack, whereas Pleistocene moraines or Holocene dunes suffer from strong erosion and abrasion. Vertical movements influence coastal development in many regions, but by different causes. Among them are long-lasting and ongoing glacio-isostatic uplift in Scandinavia (nearly 300 m, with a modern maximum of 9 mm/year, and Fig. 4), Iceland and Scotland (both with more than 120 m), salt intrusion in small places of the North Sea basin (e.g., Heligoland), active horst and graben structures in the eastern Mediterranean (Fig. 5), sediment load in large deltas, or local subsidence by gas and water drilling in the Venice area. All processes cause their respective characteristic coastal shape.

Furthermore, exogenic factors of coastal forming are manifold and here in particular climate and its influence on the ocean. The spectrum ranges from frost climates of average annual temperature below -2°C and sea ice cover of 9 months to nearly 30°C of surface water temperature in summer. Extreme humid conditions with over 2,000 mm of precipitation are typical for NW Europe, but less than 200 mm/year of precipitation is measured in eastern Crete (Greece), where evaporation exceeds precipitation by a factor of 10. The salt content of seawater ranges from nearly 0% in the inner Baltic Sea to at least 3.9% in the eastern

Mediterranean. Moreover, wind and wave spectra are very different: in the west, strong winter storms may cause severe destruction and flooding (Clarke and Rendell 2009; Weisse et al. 2014; Masselink et al. 2016), while in small bays of the Baltic or the Mediterranean wave action is insignificant. Tidal range and the strength of tidal currents differ along a west-east gradient: spring tide range is highest around southwestern England and the Brittany coast of western France (Fig. 6), reaching more than 10 m, but in the Baltic and the Mediterranean it usually is less than 0.5 m to zero. Nevertheless, low air pressure and extreme winds may cause storm surges up to 3 m above datum in these enclosed basins. Even tsunami impacts can be detected such as along the Algarve coast of southern Portugal, southern Atlantic Spain, Mallorca Island, Apulia, Greece, Cyprus (Kelletat and Schellmann 2002), or in marshes in Scotland and in coastal lakes of western Norway by the Storegga event about 8,100 BP (Bondevik et al. 1997).

Specific Coastal Regions

European coastal regions with a specific environmental or geomorphic factor are illustrated in Fig. 6: northern Europe



Europe, Coastal Geomorphology, Fig. 3 The Adriatic coast along Croatia is a typical area of parallel anticlines from Mesozoic folding, drowned by subsidence and postglacial sea-level rise to form canale and vallone coastlines (Credit: Google Earth)

including Scotland and Iceland present coastlines of emergence by glacio-isostatic uplift between near 0 and 285 m. Nevertheless, those coastal forms are predominantly marked by ingression, so that fjords (Fig. 7), fjärds, bodden, förden (Fig. 8), or partly submerged roches moutonnées as skerries, as well as partly drowned eskers, end moraines, or drumlins (Fig. 9), are characteristic (Duphorn et al. 2011). Glacial erosion has carved bedrock far below sea level, and therefore the late- and postglacial transgression is not compensated by late- and postglacial uplift. The area of glacio-isostatic uplift mostly corresponds with glacially sculptured coastal landscapes, in which the forms from the last glaciation are much clearer than those from the penultimate glaciation, which are smoothed by periglacial processes during the Weichselian. Another *specificum* of the northern sections in Europe is frost action along the coast including drifting sea ice contact, in particular, in the inner and eastern Baltic as well as in the Barents Sea and in some inner parts of northern Norwegian fjords, where Gulf Stream influence is minimal and freshwater input is important.

Pleistocene glaciations have left behind specific materials along the coasts: moraines are washed out and show boulder beaches as in the Åland archipelago between Sweden and

Finland or along the southern Baltic coastline, partly pushed into ridges by drifting sea ice. These boulders are an important natural protection against surf beat and erosion. Glacial outwash and fluvioglacial deposits are characterized by gravel and sand, which dominates the southern coasts of Iceland as well as the west coast of Denmark or the northwestern coast of Germany and the Netherlands. If sand supply from the ice ages is abundant, barrier islands with extended beaches are developed, accompanied by coastal dunes. Similar conditions are also found along the southern Baltic coastlines (Baker et al. 1990).

In contrast to formerly glaciated landscapes, fluvial-formed landscapes are dominant in the south of Europe, where rivers have carved into older bedrock during low sea levels. Postglacial transgression has transformed these valley mouths into rias as in southern Ireland, England, Brittany in France, and in NW Spain, some of them widened by tides and waves into estuaries as in the mouth of the rivers Elbe (Germany), Thames (England), or Seine (France).

A third representative coastal area in Europe is formed due to faulting processes associated with horst and graben structures such as southern Greece including the Aegean Sea (Fig. 5). A vivid neotectonic is still dislocating coastal



Europe, Coastal Geomorphology, Fig. 4 Glacio-isostatic uplifted beach ridge systems (back into late-glacial times), Varanger Peninsula, NE Norway (Credit: Google Earth)

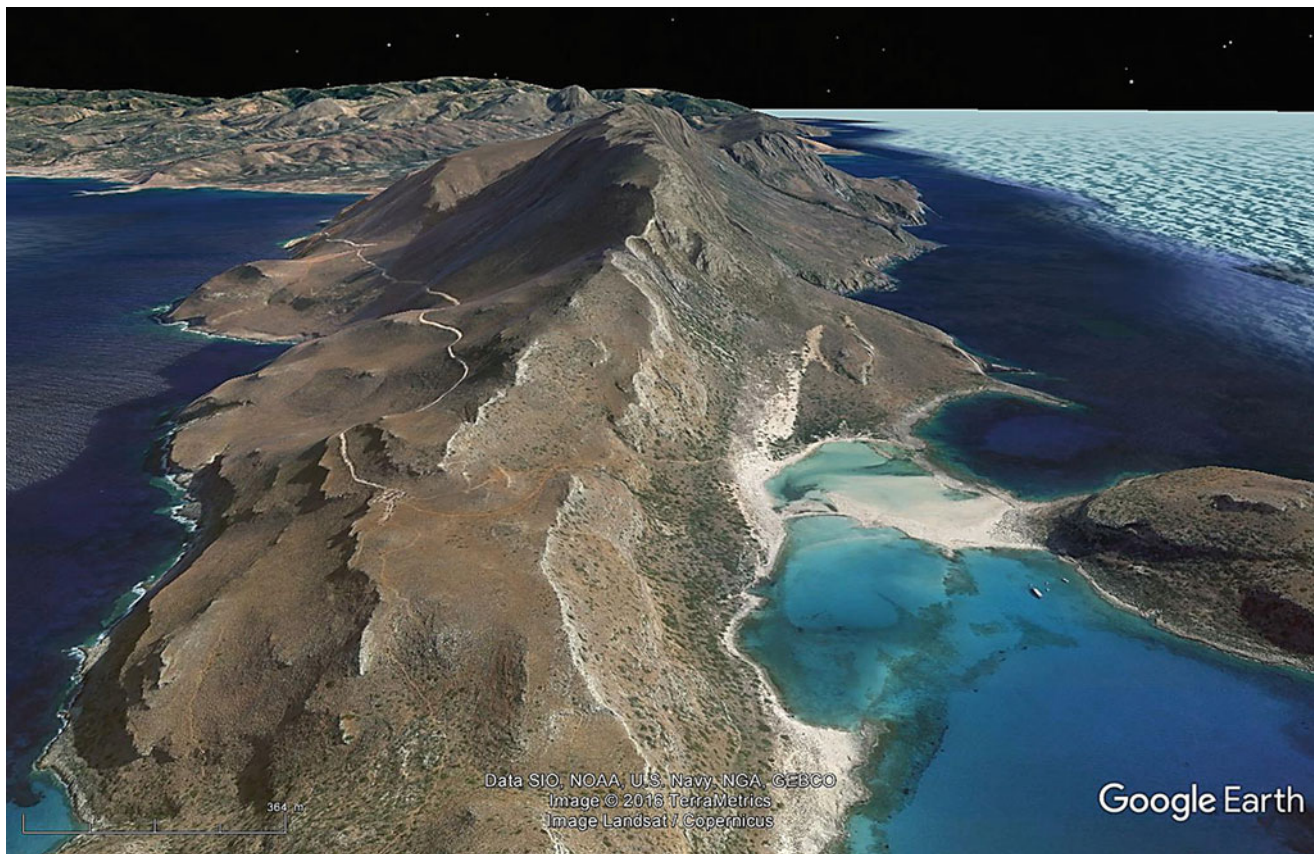
sections vertically (Pirazzoli et al. 1996, see also Fig. 10). In more downward-moved parts, coastlines of ingress with drowned medium-to-steep slopes are common (Fig. 11). Here, canals (Fig. 3) and short rias, transformed by karst and marine erosion into calas of different shapes, are developed, and as well drowned karst features like dolines or poljes dominate (Fig. 12). In contrast, cliffs and shore platforms (Figs. 13 and 14; Erdmann et al. 2015; Gomez-Pujol et al. 2014; Robinson and Lageat 2006) and coastal terraces are typical for the uplifted parts of the landscape. The areas of (Pleistocene) uplift in southern Europe often correspond with young orogenic structures, representing excellent examples of Pleistocene marine terraces up to several 100 m above sea level today.

As limestone is the predominant rock type in the Mediterranean area, coastal bioerosion with notches (mostly by grazing *Patella*, Fig. 9) and rock pools (by grazing *Littorina neritoides*) and bioconstruction by calcareous algae (e.g., *Neogoniolithon notarisii*) and vermetids (e.g., *Dendropoma petraeum*, Fig. 15) are widely common in this part of the European coast. Another specification is the extended distribution of younger Pleistocene eolianites along the Mediterranean coasts, whereas Holocene beachrocks are mostly restricted to the hot and dry sections of the eastern Mediterranean (Briand and Maldonado 1997).

There is only one area of high tides in Europe, located between southern England and the Brittany coast of France where the spring tide range reaches around 10 m, diminishing to the inner German Bight and Scotland to medium values of 2–4 m. These areas may show extended tidal flats, with a sandy matrix where wave action and tidal currents are strong and muddy in regions of weaker water movements (Fig. 16; Ehlers 1988; Hillen and Verhagen 1993). The growth of green and brown macro-algae (*Fucus* sp., *Laminaria* sp., and others) along the rocky shores is characteristic for the cooler European coastlines from northern France including the British Isles and Iceland to northernmost Norway (but excluding the Baltic). They partly give protection against strong surf beat, whereas in the Mediterranean, sea grass may form a thick carpet on beaches after winter storms. Along parts of the Swedish, Finnish, Polish, or German coastline with a low salt content of seawater, coastal freshwater swamps with *Phragmites* sp. occur.

Coastal Forms and Sediments

The differentiation of coastal forms along the European coastlines is visualized in Fig. 17 giving an impression of



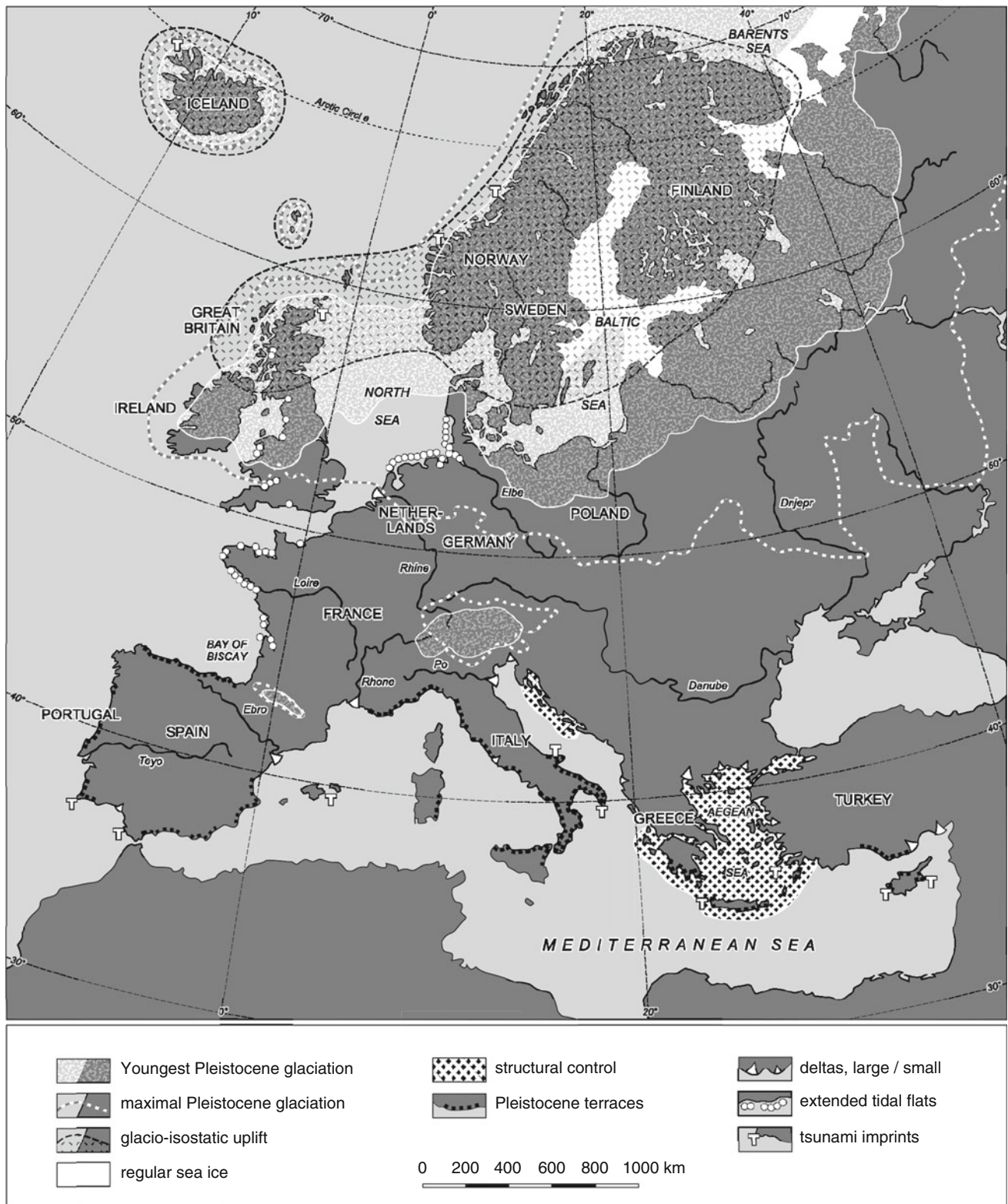
Europe, Coastal Geomorphology, Fig. 5 Stepwise uplifted Gramvousa peninsula as a horst structure in NW Crete (Greece), looking to the south. Main fault line to the west (right), with Balos lagoon area,

emerged +5 m on June 21st, 365 AD by a strong earthquake, with large tsunami boulder deposits (Credit: Google Earth)

the general form of bays: rias and estuaries by fluvial action as in Fig. 18; fjords, fjärds, förden, and bodden by glacial sculpturing; canale by folding and faulting control; cala by karst and abrasion; or lagoons with barriers and spits up to 90 km in the eastern Baltic by longshore drift). The map also represents the different coastline types: rocky coastlines with cliffs (and intertidal platforms), sandy shores with beaches (including areas with beachrocks), muddy ones in tidal flats, deltaic by fluvial progradation (as in the areas of low tides and surf of the Mediterranean), or biogenic structures along limestone coasts with poor sediment supply.

Deeply incised fjords are the main coastal features of northern Iceland, northwestern Scotland, and nearly all of Norway except of the easternmost parts where cliff sections around the North Cape and the Barents peninsula occur. In Scotland and the central north coast of Iceland, wider rocky embayments in a low and hilly landscape smoothed by Pleistocene glaciation reveal more aspects of fjärds, which are typical for the Swedish coastline along the Baltic and some parts of Finland. Here, an old peneplain is slowly dipping seaward. Numerous skerries decorate all rocky Scandinavian coastlines, whereas to the periphery of former

glaciation, morainic sediments and fluvioglacial deposits dominate. Thus, the landscape is either characterized by partly drowned eskers, drumlins, and end moraines, which are mostly well preserved in areas with low surf energy conditions, such as in the island archipelago of southeastern Denmark, or fast receding cliffs may have been developed like along the southern shores of the Baltic. As a rather unusual feature in the Scandinavian coastal landscape the so-called strandflat exists as an undulating rocky plain surrounding most of the coasts of Norway but occurring likewise in northwestern Iceland. The form originates from a glacially smoothed old peneplain now situated around sea level with a medium width of 16 km and a maximum one of 40 km. Roches moutonnées are the dominant features, and old inselbergs may rise from the peneplain to several 100 m of relative height. Outwash plains (sander), fed by meltwater and glacial-burst floods, are typical for the southern part of Iceland. Therefore, the coasts show beaches and even barriers and lagoons. Along the southern shores of the Baltic spits fed by longshore drift from the west (Fig. 19), lagoons and intercalated cliffy sections occur. Along the Polish coast 52% are sandy beaches, 28% are freshwater swamps, and only 20% show cliffs. At the German Baltic



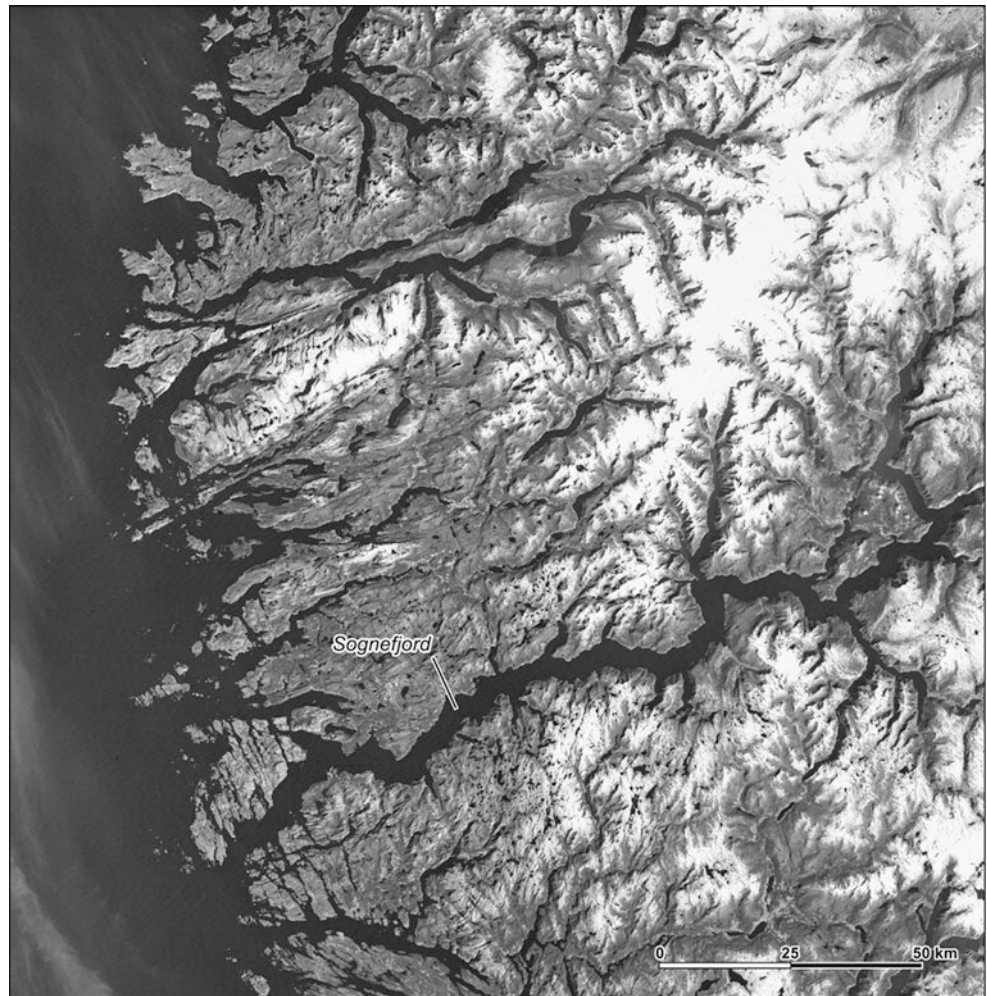
Europe, Coastal Geomorphology, Fig. 6 Map of the European coastlines showing special environmental parameters

coastline, this type gradually changes to ingressive forms from glacial and fluvio-glacial origin like bodden and förden (Fig. 8).

The longest section of muddy and sandy tidal flats in Europe with barrier islands and spits can be found from western Denmark all along the German North Sea coast and the

Europe, Coastal Geomorphology, Fig. 7

Satellite image of Sognefjord in western Norway, the deepest (more than 1,200 m) and longest fjord in northern Europe with about 200 km of length and a width of 1–8 km. Around the mouth of the fjord structural control of the coastline in very old crystalline rocks, worked out by several glaciations (© EOSAT, GAF, AAF 1985)



Netherlands to Belgium and northernmost France (Fig. 20). Furthermore, inland marshes are typical (Dijkema 1987), and along broader sandy beaches, coastal dunes are well developed (Baker et al. 1990; Clemmensen et al. 2009; Jackson and Cooper 2011). Other areas of tidal flats are present in many ria embayments from Ireland to Great Britain and western France. They partly are rocky, exhibiting a rich variety of algal belts and hard encrustations of barnacles, mussels, etc. Northern England shows a great variety of spits, marshes, tidal bays, bars (Fig. 21), forelands with wide beach ridge systems (Fig. 22), and uplifted coastal terraces, whereas the southern part is dominated by the world famous high chalk cliffs along the English Channel and rias of different length and steep rocky headlands to the southwest. The southwestern coast of France along the Bay of Biscay shows the longest coastal dune belts of Europe with shifting sands more than 100 m height (Fig. 23).

The coastline of Northern Spain along the Bay of Biscay is rocky with cliffs and caves cut into the limestone. To the west narrow and deep rias can be observed (Fig. 18), which change to the Portuguese coast into wider and shallower fluvial

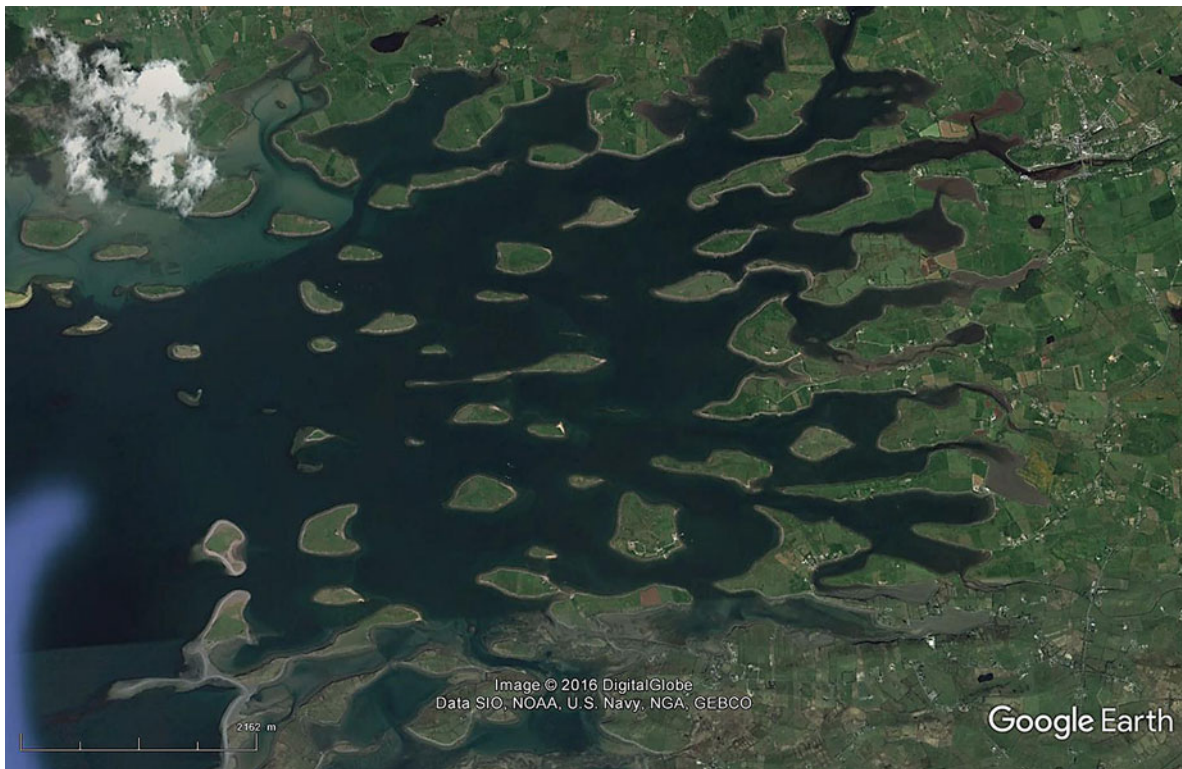
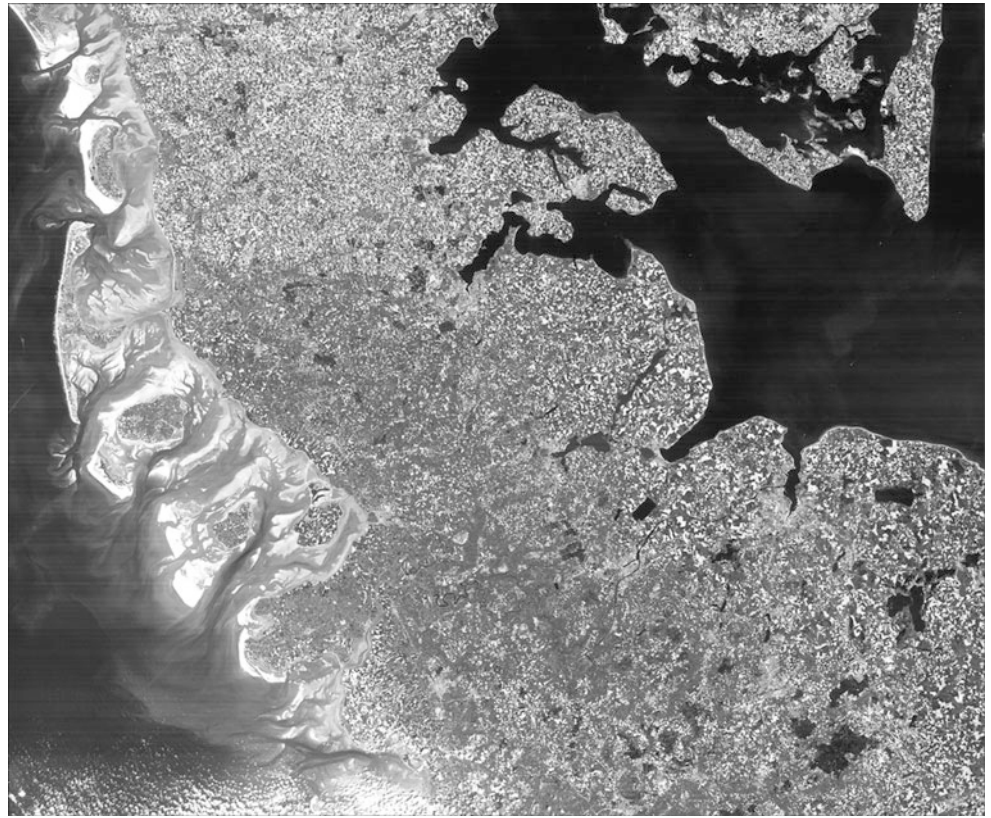
embayments. The coasts of Portugal, however, are mostly sandy with dune belts and lagoons, except of some rocky headlands and the high and steep cliffs developed in the hard limestone at the southwestern most point of Europe. They are constituted of soft rocks with numerous small landslides at the south-facing Algarve coast. Continuing to the Strait of Gibraltar the eastern Portuguese coastline and the Spanish one are mostly low and bordered by lagoons and dune belts along sandy beaches. Only the last 50 km exhibit rocky coastlines, headlands, and cliffs with smaller sandy beaches.

The volcanic archipelagos of Madeira, the Azores, and the Canary Islands are mostly steep, some with megaciffs up to more than 500 m height. They are not singularly formed due to abrasion, but in addition they represent head scars of huge submarine slides from the collapse of the high and growing volcanic edifices.

The coastlines of the Mediterranean are limited in their coastal forms because tides are nearly absent, and longshore drift is less efficient because of the smaller ocean areas. Thus, the limited tides and wave energy conditions are responsible

Europe, Coastal Geomorphology, Fig. 8

Satellite image of northern Germany and southern Denmark. To the west tidal areas with different types of islands (Pleistocene morainic hills, younger Holocene marshes, very young spits and dunes), sandy bars, and muddy flats with tidal rivers and ebb deltas occur. To the east at the Baltic coastline ingressions into morainic and fluvioglacial relief on the Danish islands or into ria-like *förden* is typical in an environment without tides (© ESA, Eurimage 1992, GAF 1995)



Europe, Coastal Geomorphology, Fig. 9 Partly drowned drumlins in Newport Bay, western Ireland (Credit: Google Earth)

Europe, Coastal Geomorphology, Fig. 10

Bioerosive notches in limestone at the east coast of Rhodes (Greece) show steplike neotectonic uplift during the last 2,000 years to a maximum of about 2.5 m



Europe, Coastal Geomorphology, Fig. 11 Coastal contours pointing to subsidence and ingression: Greek island of Astypalaia, (16 km long), Aegean Sea (Credit: Google Earth)

for many deltas, among them the very large ones of Ebro (Fig. 24), Rhône, or Po (Fig. 25) and numerous others along the Italian and Albania coasts. Headlands with cliffs in limestone including bioerosive notches alternate with longer

beach sections at the Spanish Mediterranean coastline, whereas in southern France the western section consists of beaches, barriers, beach ridges plains, and the Rhône delta and of rocky steep coastlines in the east with calanques and



Europe, Coastal Geomorphology, Fig. 12 Drowned poljes on Lesbos Island, Greece (Credit: Google Earth)



Europe, Coastal Geomorphology, Fig. 13 Cliff and slightly inclining abrasive rock platform at the south coast of England (Credit: Google Earth)



Europe, Coastal Geomorphology, Fig. 14 Over 40 m high cliffs in chalk, with caves, basal rock platform, and boulder beach, southern England (Credit: Google Earth)

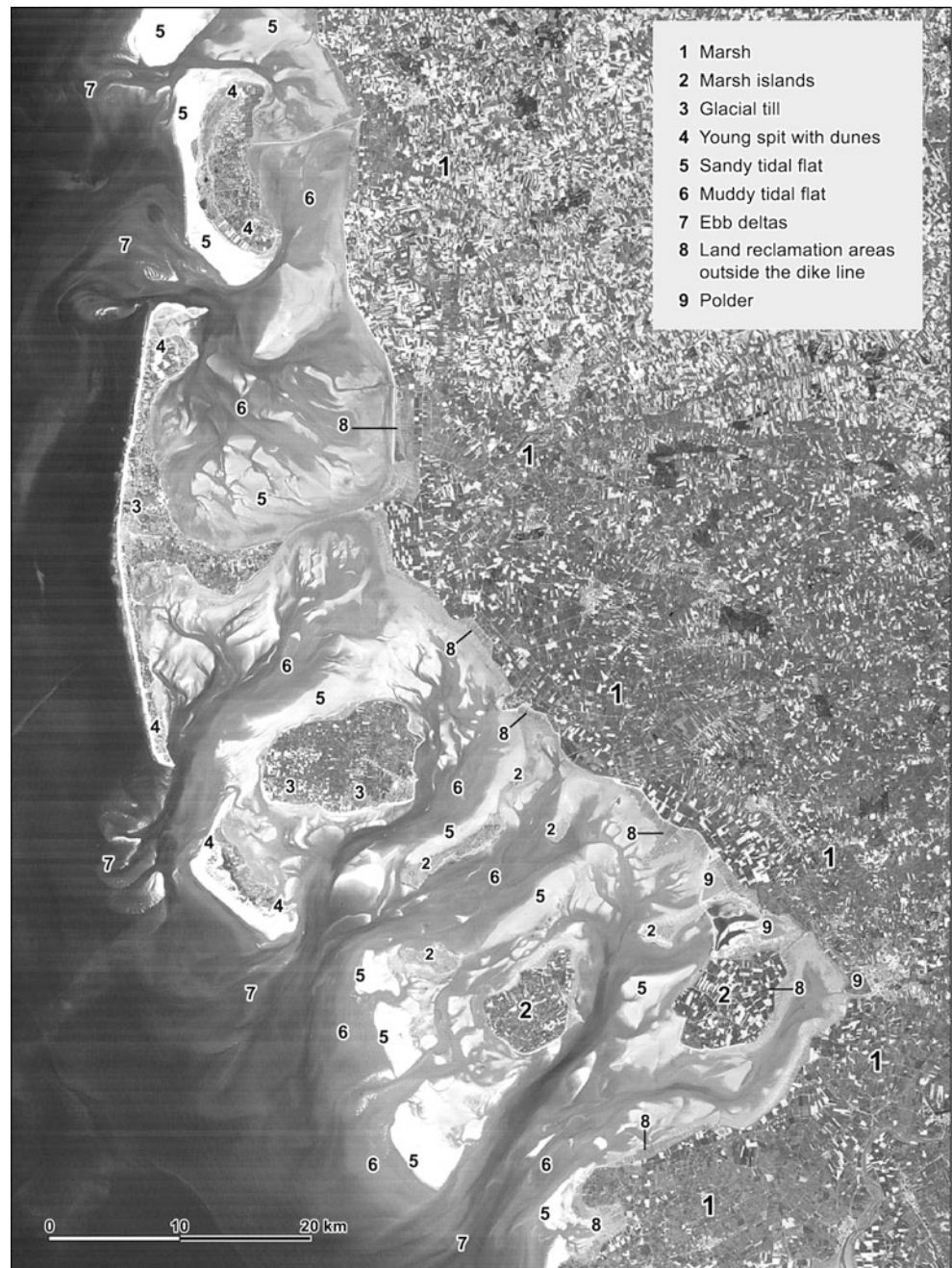
Europe, Coastal Geomorphology, Fig. 15

Living bioconstructional microatoll by vermetids (*Dendropoma petraeum*) giving a very reliable sea-level indicator along the west coast of Crete, Greece



Europe, Coastal Geomorphology, Fig. 16

Satellite image of the tidal areas and islands along the west coast of northern Germany (© ESA, Eurimage 1992, GAF 1995, modified)

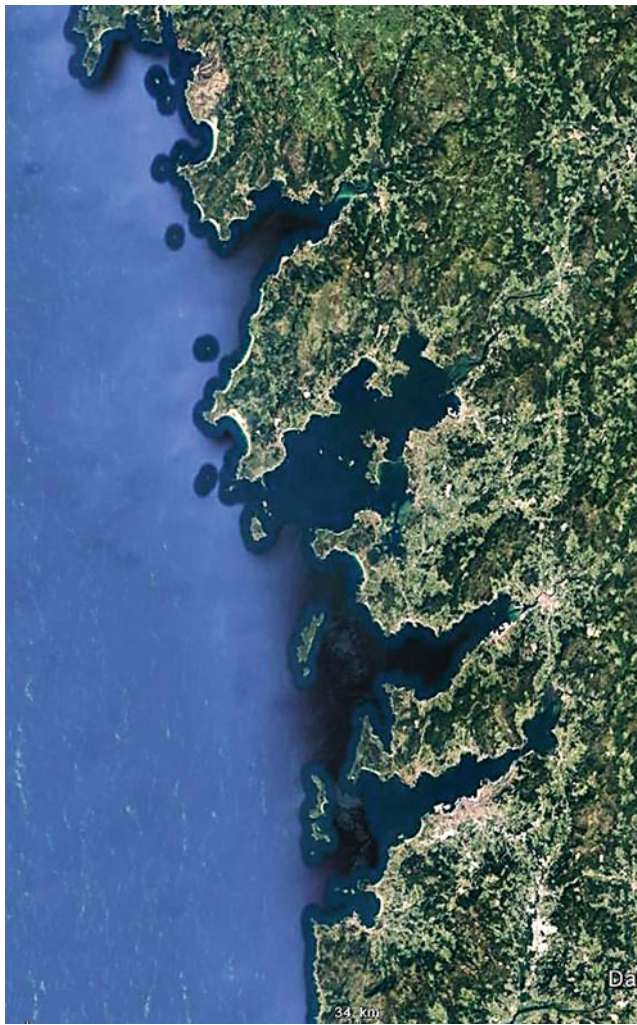


calas, which are dominant on the Balearic Islands, as well. The most extended granitic coastal landscapes decorated by huge boulders from deep weathering during pre-Pleistocene times and Holocene tafoni can be found around the islands of Corsica and Sardinia. Here, also longer sections of coastal barriers and lagoons exist. Except for the young volcanic Aeolian Islands, Italy has beaches along nearly 70% of its coastline, whereas in the southeast and on Sicily, cliffs and Pleistocene coastal terraces are widespread. From Istria to

northern Albania, drowned landscapes with typical parallel canale features occur (Fig. 3); beaches are very rare. Another type of an ingressive coastal landscape can be found along the Greek and Turkish coasts. It is the various change of structural embayments as grabens and horst-like long peninsulas (Figs. 2 and 5), mostly showing drowning as the predominant process with medium-to-steeper slopes but a less amount of real cliffs. Limited sections of uplifted coastal terraces alternate with beaches along barriers and deltas.



Europe, Coastal Geomorphology, Fig. 17 Forms and sediments along the coastlines of the European continent



Europe, Coastal Geomorphology, Fig. 18 Ria coast in NW Spain (Credit: Google Earth)

The islands of the Aegean Sea again are drowned remnants of an older land bridge (Fig. 10), sometimes with karst features (Fig. 11). Due to the absence of longer rivers and the rather dry climate, beach material is scarce. Neotectonics are an important process in coastal forming, as well as bioconstruction at the warmest easternmost parts of the Mediterranean, where many of the small beaches are cemented by Holocene beachrocks.

The maps do not show areas of artificial coastlines mostly due to the need for protection against storm floods in winter by groins, beach nourishment, sea walls, dikes, or land reclamation, most widespread around the North Sea as well as along Mediterranean bathing beaches and harbors.

Many of the European coasts are endangered (Caputo et al. 1991; Bartolini and Carobene 1996). The primary cause hereby is the accelerated erosion along beaches – at least 75% of all European beaches suffer from this process – marshes, and coastal dunes, which as a consequence costs billions of Euro/

year for maintaining bathing beaches, coastal settlements, or farmland. The reason is a small general sea-level rise and a local accelerated one caused by deltaic sedimentation, downwarping of a ring-like structure around the glacio-isostatic uplifted areas, or pumping of oil, gas, or groundwater from below the sea. Some of the reasons of the accelerated erosion may be located not directly along the coasts but far inland. The predominant cause is the increasing number of reservoirs holding back sediment and the mining of sand, gravel, and cobble from river beds, in particular, in the Mediterranean area. Consequently, millions of tons of sediment are not transported to the coasts any longer. This is in striking contrast to a period of progradation and delta buildup during late Holocene and historic times, in particular, during the Greek and Roman antiquity and early Medieval times, in some places even starting during the later Neolithic. For several 1,000 years progradation was in particular responsible for the development of barriers, spits, or marshes in spite of the still slowly rising sea level. The reason was the ongoing extension of agriculture and deforestation, resulting in soil erosion and denudational processes apart from the coast (see also Caputo et al. 1991; Simeoni and Corbau 2009; Labuz 2015). At many places, human impact on the coasts can be traced back in historic times, and geoarchaeology is used to identify changes of shorelines, sedimentation, and sea level (Behre 2001; Anthony et al. 2014; Peeters and Momber 2014). Coastal protection against the increased frequency of storms is another problem, which leads to the design of continuous new and higher dikes. It can be stated that overall the shape of some thousands of kilometers of coastline in Europe is the result of anthropogenic structures. However, beside that, heavy pollution due to tourist settlements or rivers bringing pulp mill sewage to the sea and oil spillage from heavy road traffic additionally are causing severe problems in coastal areas.

Pleistocene and Holocene Sea Levels

In northern Europe, glaciation has eliminated all traces of sea levels from former interglacial highstands, and those are mostly missing in areas without significant uplift like around the North Sea basin and the English Channel. In contrast evidences of higher interglacial sea levels as sediments with marine fossils (e.g., the foraminifera *Hyalinea balthica* for colder events in the older Pleistocene, or the tropical gastropod *Strombus bubonius* for the younger interglacials in the Mediterranean) or even as geomorphic features like coastal terraces are widespread in the geologic young environments of the alpidic orogenesis. Côte d'Azur between France and Italy, northern Sicily, or Apulia and Calabria in southern Italy are world famous sites for the discussion of higher interglacial sea levels. At some places, such as in SW and NE Peloponnesos, Greece, or southern Italy, terrace sequences reach several



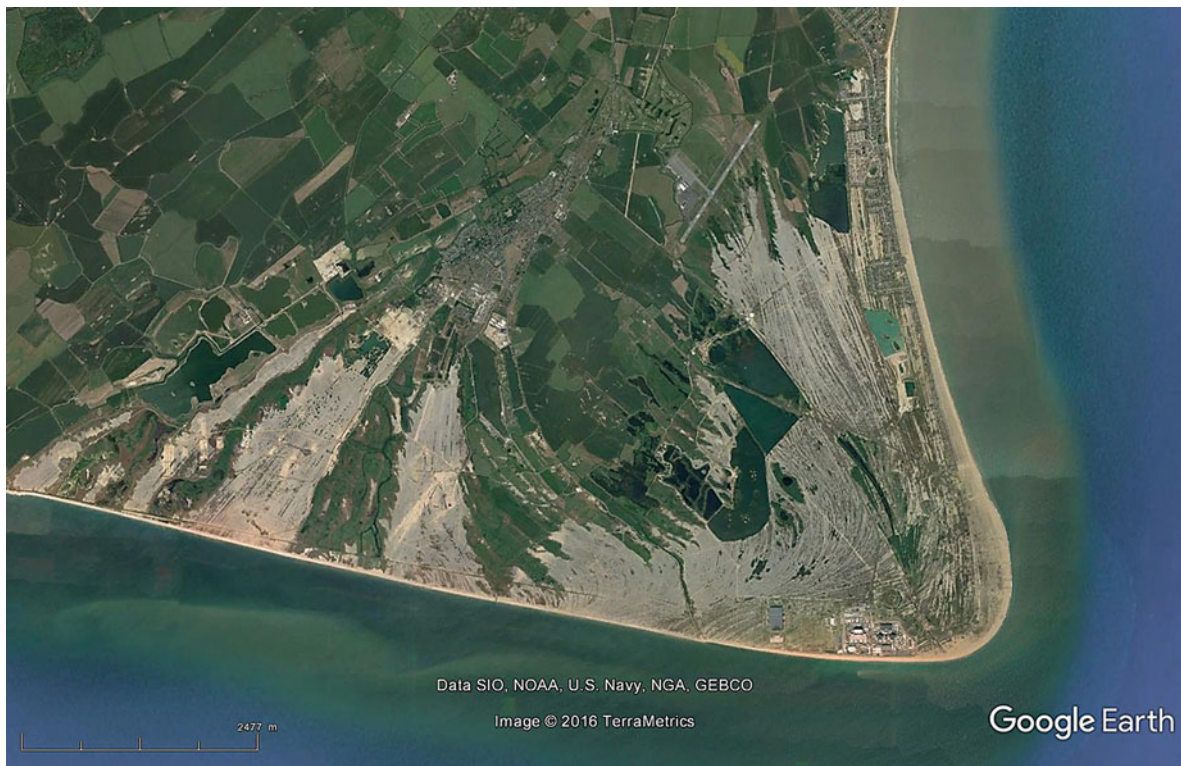
Europe, Coastal Geomorphology, Fig. 19 Growth steps of a pebble spit built from W at the German Baltic coastline (Credit: Google Earth)



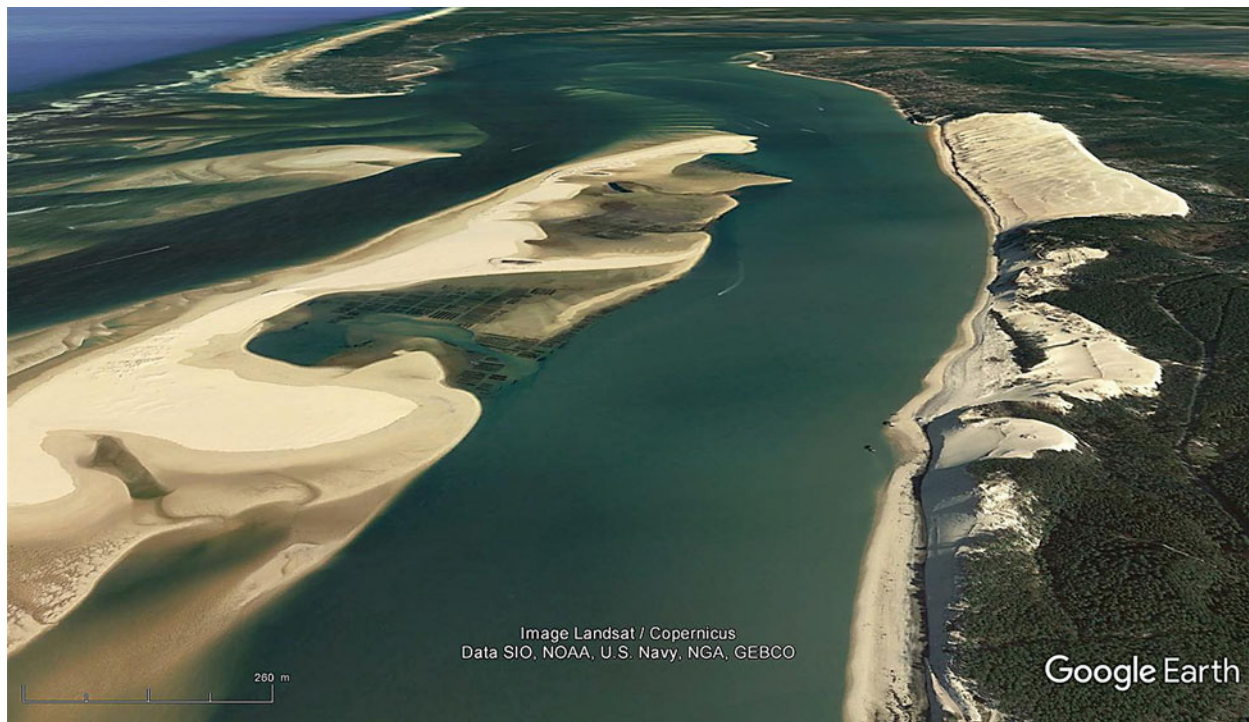
Europe, Coastal Geomorphology, Fig. 20 Line of seven barrier islands along the German North Sea Coast (Credit: Google Earth)



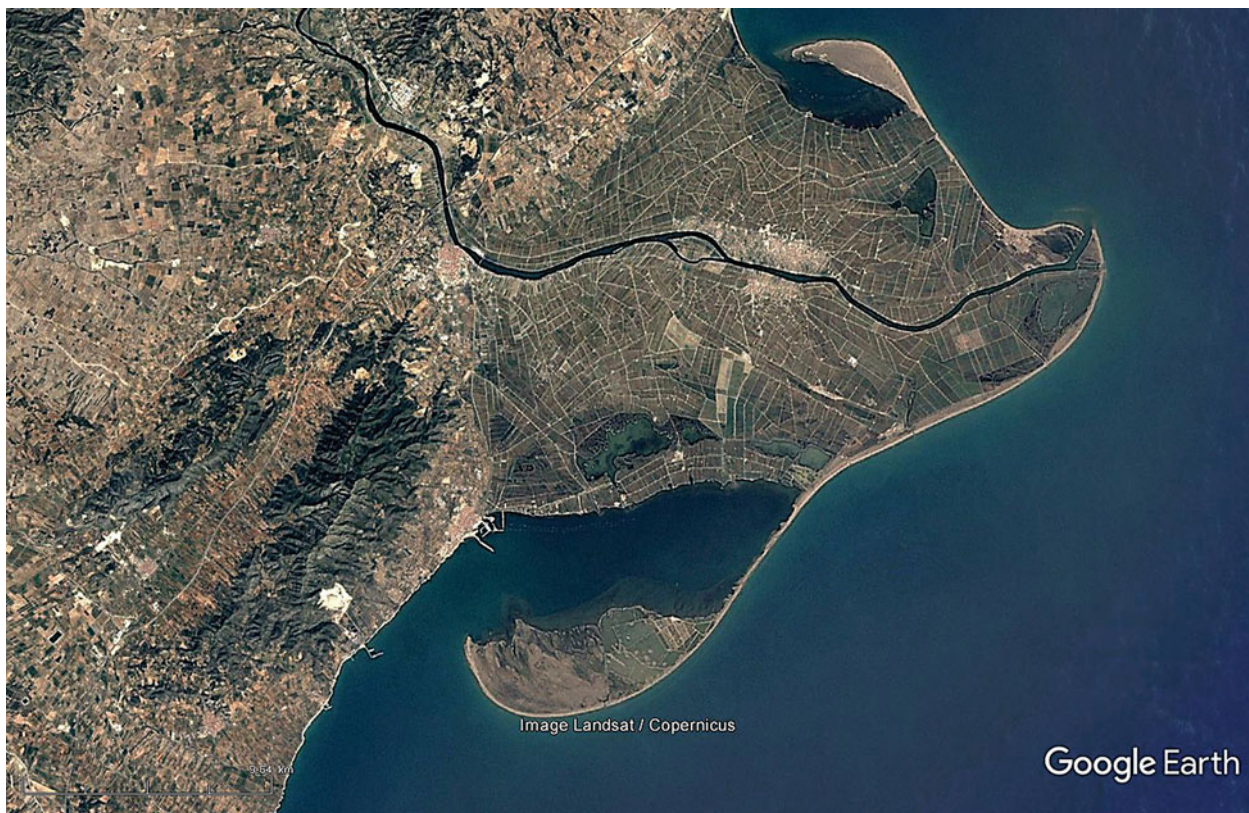
Europe, Coastal Geomorphology, Fig. 21 3.3 km long bar transforming into a barrier island ENE of Inverness, northern Scotland (Credit: Google Earth)



Europe, Coastal Geomorphology, Fig. 22 Beach ridge systems of Dungeness foreland, SE England, with >500 ridge segments (Credit: Google Earth)



Europe, Coastal Geomorphology, Fig. 23 Sand bars and active dunes along the SW coast of France (Bassin d’Arcachon near Bordeaux; (Credit: Google Earth)



Europe, Coastal Geomorphology, Fig. 24 Ebro Delta (Mediterranean Spain), with two winglike spits (Credit: Google Earth)



Europe, Coastal Geomorphology, Fig. 25 Satellite image of the delta of River Po, Italy. Progradation mostly is a young Holocene feature (© ESA, Eurimage 1989, GAF 1989)

100 m above sea level. Unfortunately, the absolute dating determination remains still rather difficult. The coasts and the islands of the eastern Mediterranean exhibit remnants of higher sea levels but mostly in short sections, bordered by faults and areas of significant subsidence. The differentiation of substages within the same interglacial (like oxygen-isotopic stages 5e, 5c, 5a) is still under research. Nevertheless, the study of high (or uplifted) interglacial sea-level evidence has contributed greatly to the understanding of Mediterranean neotectonics.

Holocene sea level is a topic of intensive discussion in the scientific community. In northern Europe, a very good

dating of glacio-isostatic rebound by the analysis of uplifted beach ridge sequences exists (Fig. 22), whereas along the marshy German, Danish, Netherlands, and English coastlines of the North Sea, geo-archaeology has much contributed to the identifying of former Holocene sea levels (Behre 2001). Here, the sea-level curve of the last 6,000 years is characterized by at least two transgressions and regressions of 1–3 m. Some archaeological remains of the Bronze Age below high water level in the high tide range area of western France point to a sea-level rise of more than 3 m in this geologic stable region. Along the Mediterranean coastlines, archaeology as well contributes to the identification of the Holocene sea-level curve: drowned Phoenician harbors and ruins and quarries from Greek antique times, the position of Roman fish tanks, and even relics of early Christian periods can be found below sea level. Beside these man-made structures, natural sea-level indicators are very suitable for relative dating approaches, among them bioerosive notches (Fig. 9) or bioconstructive elements like trottoirs or algal and vermetid rims (Fig. 15 and Kelletat 1997). They are more reliable than beach ridge sequences, beachrock, or terraces along deltas. In spite of literally hundreds of dated and measured sites, there is no general agreement about the eustatic sea-level movements in the younger Holocene in the Mediterranean. But, nevertheless, these studies have led to a much better understanding of the ongoing and vivid neotectonics in this part of Europe, reaching a maximum of 9 m on western Crete (Greece) since the year AD 365 (Pirazzoli et al. 1996).

Conclusion

The complex contours of the European continent, its long shorelines, and its position between high Arctic to subtropical latitudes as well as per-humid to arid climates result in the presence of nearly all coastal forms and types of the world except of mangroves and coral reefs. Marine forming processes as waves and tides also show large variations. Beside anorganic factors, organic ones contribute to coastal forming additionally and in particular influence coastal ecosystems as in the macro-algae belts in the north, or bioerosive and bioconstructive indicators dominantly in the Mediterranean. As coastlines normally occur along a horizontal level, dislocations like uplift out of Holocene and interglacial levels as well as tilting are perfect indicators for the analysis of tectonic movements or those from glacio-isostatic rebound. Additionally, the long and well-documented history of human occupation and numerous archaeological remains in the coastal realm contribute to the differentiation of even small environmental changes along the coasts during the younger Holocene well into historical and modern times.

Cross-References

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- [Changing Sea Levels](#)
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- [Tectonics and Neotectonics](#)
- [Tidal Flats](#)

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Eustasy

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Definition

Eustasy originally referred to a globally simultaneous and uniform change in sea level. However, sea-level changes are affected by many processes that operate on different temporal and spatial scales. These processes are described more fully

Eustasy, Table 1 Summary of processes affecting global sea level

Process	Time scale (year)	Magnitude (m)
Changes in volume of ocean water		
Expansion/contraction of ice sheets (glacial eustasy)	10^4 – 10^5	120–140
Expansion/contraction of mountain glaciers	10 – 10^4	0.5
Changes in ocean temperature/salinity (steric changes)	10 – 10^4	10
Changes in ocean basin volume		
Plate tectonics and mountain uplift	10^7 – 10^8	100–500
Sedimentation (continental shelves)	10 – 10^8	100
Glacio-isostasy	10^4 – 10^5	100–500
Hydro-isostasy	10^4 – 10^5	30–40
Local sea-level changes		
Sediment loading and compaction (river deltas)	10–1000	<1–20
Localized faulting and tectonic deformation	1–1000	<1–3
Land subsidence (subsurface fluid extraction)	10–100	<1–2
Oceanographic and atmospheric variability	<1–10	<1–10

After Gornitz (2013) and sources cited therein

below and summarized in Table 1. Therefore, the original concept of eustasy, as a globally uniform rise in sea level, has been challenged, and some authors suggest abandoning the term altogether (e.g., Carlson and Clark 2012). Nevertheless, eustatic sea level plays an important role within the framework of geophysical models that calculate glacial isostasy (see below; Peltier 2004; Mitrovica et al. 2010; Lambeck et al. 2010; Spada 2017). In current usage, the eustatic sea-level change is the global sea-level change averaged over the surface of the oceans, over a specified time interval (Spada 2017).

Suess (1888) originally attributed eustasy to crustal subsidence and sediment deposition. Removal or addition of water to oceans during glacial-interglacial cycles was another proposed cause. Fairbridge (1961) summarized these theories as follows: (1) *tectono-eustasy* (tectonic deformation of the ocean basins), (2) *sedimento-eustasy* (loading of crust by sediments), and (3) *glacio-eustasy* (removal/addition of water to the oceans by expansion or contraction of polar ice sheets). Another process is *hydro-isostatic* loading of the seafloor by glacial meltwater (Walcott 1972; Clark et al. 1978; Mitrovica and Milne 2002).

Mean global sea level can change in two fundamentally different ways – the first by altering the mass (or volume) of ocean water and the second by varying the volume of the ocean basins. In the first case, the mass (or volume) of the contained liquid (i.e., ocean water) varies; in the second, the shape and dimensions of the container (i.e., the ocean basins)

are modified. The land-to-sea transfer of ice mass, and vice versa, during glacial-interglacial cycles represents the primary means of changing the amount of water in the ocean and, hence, global sea level (i.e., glacio-eustasy) (Table 1). These changes operate on timescales ranging between 10,000 and 100,000 years. However, changes in ocean temperature or salinity (i.e., steric changes) also affect water density, ocean volume, and hence water level, over much shorter timescales. On the other hand, changes in ocean basin volume caused by plate tectonics and mountain building span periods of tens to hundreds of millions of years. Loading or unloading of ice mass also alters basin shape (i.e., glacio-isostasy).

Seismic reflection profiles on continental shelves, collected during petroleum exploration in the 1970s, showed similar patterns of “sequence boundaries,” or regional unconformities, in passive margin sedimentary deposits from many parts of the world (see “► Sequence Stratigraphy”). Sequence boundaries have been widely interpreted as marking major episodes of synchronous eustatic sea-level falls, which are summarized in global sea-level curves (e.g., Vail et al. 1977; Haq et al. 1987). Although widely used, these curves have been criticized on grounds that sequence boundaries are uncritically accepted as eustatic, that magnitudes of sea-level fluctuations are largely speculative, and that boundaries have not been accurately dated (for good overviews, see Christie-Blick et al. 1990; Miall 1997). However, curves from various sources show certain common features, suggesting that at least some of these proposed sea-level events could be worldwide in scope. More recently, Miller et al. (2005, 2011) have developed a eustatic sea-level curve that combines marine oxygen isotope data with continental margin and seismic stratigraphy, corrected for effects of basin subsidence, sediment compaction, and loading.

Causes of Eustasy

Sea level integrates the effects of diverse geological and meteorological phenomena, such as glacio-eustasy, glacio-isostasy, plate motions, etc., which operate on differing spatial and temporal scales (Table 1). The relative magnitudes of these processes vary from place to place, thus tending to obscure any globally coherent sea-level signal. As a result, the search for a universal “eustatic” sea-level curve has turned toward obtaining improved regional and local (“relative”) sea-level curves (Lambeck et al. 2010). Yet, as shown below, improved modeling of glacio-isostatic adjustments and detailed, accurately dated sea-level data from areas of known uplift enable reconstruction of global sea-level rise over the last 18,000 years and the past century. Processes affecting global sea level over various geologic timescales fall into three main classes (Table 1). These involve (1) changes in the volume or mass of ocean water, (2) changes

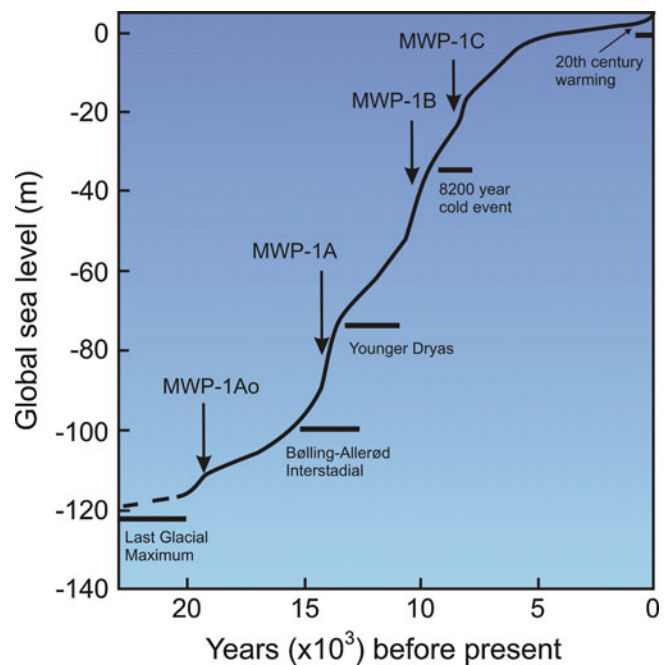
in the volume of the ocean basins, and (3) dynamic changes of the ocean surface.

Changes in the Volume or Mass of Ocean Water

Glacio-eustasy. Melting or accumulation of continental ice sheets during the last 2.7 million years has produced large changes in ocean water volume and global sea level. These oscillations of sea level over timescales of tens to hundreds of thousands of years are caused primarily by changes in solar radiation arising from periodic variations in the earth's orbit around the Sun (the “Milankovitch” cycles, Imbrie et al. 1984).

Changes in sea level during glacial-interglacial cycles have been estimated from fluctuations in the ratio of ^{18}O to ^{16}O in deep-sea microorganisms, such as foraminifera from ocean sediment cores, and also by dating emerged coral reefs on tectonically rising islands (see “► [Geochronology](#), ► [Coral Reefs](#), ► [Coral Reef Coasts](#), ► [Coral Reefs, Emerged](#)”). Variations in the oxygen isotope ratio serve as a proxy for alternations in global ocean volume, and hence in sea-level, linked to changing masses of continental ice. Sea-level curves for the past several hundred thousand years have been constructed from well-dated raised coral terraces on Barbados and Papua New Guinea (Lambeck et al. 2002; Siddall et al. 2007). These data show highstands of sea level at approximately 193,000, 125,000, 100,000, and 83,000 years ago.

The post-glacial marine transgression of the last 19,000 years has been plotted by careful radiometric dating (^{14}C and ^{230}Th - ^{234}U) of corals from three widely separated oceanic islands with known tectonic trends and minimal glacio-isostatic influence and various other localities. A compilation of data from these sources yields a generalized sea-level curve that shows a remarkably consistent and uniform rise of around 120–130 m, which therefore represents a eustatic curve for the last deglaciation (Fig. 1, after Gornitz 2013, Table 5.1, and sources cited therein; Peltier and Fairbanks 2006; Carlson and Clark 2012; Deschamps et al. 2012; Liu et al. 2016). However, several periods of accelerated sea-level rise (termed “meltwater pulses”) are detected: the oldest at around 19,000 years BP (labeled MWP-1A₀), followed by the largest upward sea-level jump, 16 m in ~350 years (meltwater pulse 1A, MWP-1A), during the relatively mild Bølling-Allerød period between 14,600 and 14,000 years ago (Deschamps et al. 2012; Fig. 1). This episode was later succeeded by two weaker meltwater pulses (MWP-1B and MWP-1C, respectively, Fig. 1). Meltwater pulse 1B – between 11,500 and 11,000 years ago – occurred after a return to warmth, following the Younger Dryas cold period between 12,800 and 11,700 years ago. Although fossil corals from Barbados record this sea-level jump, elsewhere this event appears only as a slight acceleration in the overall



Eustasy, Fig. 1 Generalized sea-level curve of the postglacial marine transgression (Based on data in Table 5.1, Gornitz 2013)

trend of rising sea level (Bard et al. 2010). Meltwater pulse (MWP1-C) may be related to an episode, around 8200 years ago, when colder temperatures gripped Greenland and Europe. The cold may have been triggered by the final catastrophic drainage of giant remnant glacial lakes along the southern edge of the Laurentide Ice Sheet and discharge into Hudson Bay (Carlson and Clark 2012). Thus, meltwater pulse events generally correspond to major periods of rapid breakup and melting of the ice sheets (Fig. 1).

The process of glacio-eustasy is now well established for the late Cenozoic. Antarctica has been glaciated for at least 14 million and probably as much as 34 million years (Galeotti et al. 2016). The late Paleozoic Gondwana glaciation led to a major marine regression. A late Ordovician glaciation was centered on the Sahara Desert in North Africa. Beyond these, other glacial episodes are less well established.

Evidence for Milankovitch cycles in the geologic record extends possibly as far back as early Mesozoic (Olsen and Whiteside 2009). Additional indications of climate (and sea-level) variability associated with orbitally driven cycles also appear in Eocene-Oligocene boundary offshore marine sediments from the western Ross Sea, Antarctica (Galeotti et al. 2016), and in Pliocene glaciomarine deposits from the Ross embayment (Naish et al. 2009). A transitional period, roughly 1.2–0.9 million years ago, marked the onset of at least eight major glacial-interglacial cycles that lasted roughly 100,000 years, each. Prior to the mid-Pleistocene transition, older orbital cycles were largely dominated by 41,000-year obliquity cycles. Reasons for the changeover from late

Pliocene-early Quaternary obliquity-dominated cycles to 100,000-year eccentricity-driven cycles within the last millennium years remain unexplained.

Hydro-eustasy: Redistribution of water among the major terrestrial hydrological reservoirs, in response to major climate shifts, can also affect the ocean's level. A potentially significant terrestrial reservoir on timescales of tens to hundreds of thousands of years resides in groundwater (Hay and Leslie 1990). Sediments may contain enough pore space to hold the equivalent of 76 m of sea level (or 50 m with isostatic adjustment).

Steric changes: Over timescales of decades to centuries, changes in ocean water density become important contributors to sea-level change. Warming, cooling, or changes in salinity of ocean water affect its volume (but not its mass content). Thermal expansion of the upper layers of the oceans, as the earth's climate warmed approximately 0.85 °C during the last hundred years, could account for 0.1–0.7 mm/year of the observed sea-level rise of 1.5–1.9 mm/year during this period (Church et al. 2013). The thermal expansion of ~1.1 mm/year between 1993 and 2010 represents about a third of the observed recent sea-level rise of 3.2 mm/year.

Changes in Volume of Ocean Basins

Tectono-eustasy: On geologic timescales of tens to hundreds of millions of years, major factors controlling sea level include changes in seafloor spreading rates, cooling of oceanic crust, and passive margin thermotectonic subsidence (Rona 1995; Miller et al. 2005, 2011). At times of rapid seafloor spreading, a comparatively larger volume of younger, warmer, more buoyant basaltic crust is created at mid-oceanic ridges. This displaces water, raising sea level on the continental margins. Slow seafloor spreading rates produce the opposite effect. Plate collision, thrusting, and terrane accretion can also cause changes in sea level on timescales of 10^7 years. Continental collision reduces land area due to compression, producing a corresponding increase in ocean area which can lead to a drop in sea level of up to 100 m (Table 1). Other tectono-eustatic mechanisms include hot-spot activity and deep mantle upwelling (Hallam 1992).

Sedimento-eustasy: Displacement of water by sedimentation on continental margins and ocean basins displaces water, causing an apparent rise in sea level.

Glacio-isostasy: Loading and unloading of ice masses during major glacial cycles induce nonuniform glacio-isostatic adjustments of the earth's crust (Table 1). Resulting changes in coastal elevation strongly affect the location of the earth's coastlines and local sea level. For example, much of Canada and Scandinavia, formerly covered by ice, are still rebounding isostatically, and local sea level is falling. Conversely, eastern United States and the low countries of

Europe, at the southern edge of the ice, are still subsiding, with locally rising sea level (Peltier 2004; Engelhart et al. 2011). GIA-related subsidence accounts for a third to half of the observed recent sea-level rise along the US east coast (Engelhart et al. 2011; Engelhart and Horton 2012).

Geophysical models (GIA or glacial isostatic adjustment models) calculate the redistribution of mass accompanying the last deglaciation, as well as that associated with recent global warming (Peltier 2004; Mitrovica et al. 2010; Lambeck et al. 2010; Peltier et al. 2015; Spada 2017). These models filter out residual glacio-isostatic movements which contaminate the twentieth century sea-level records based on tide gauges. This has allowed an estimation of the climate warming-induced “absolute” or eustatic sea-level change (e.g., Church et al. 2013; Jevrejeva et al. 2014). An alternative approach employs paleo sea-level proxies, such as tidal zone fossil microfauna, plants, and well-dated basal peat samples, to calibrate and constrain GIA models and to reconstruct paleo sea-level history (e.g., Engelhart et al. 2011; Engelhart and Horton 2012; Gehrels and Woodworth 2013). Recent GIA models also include rapid melting of large ice reservoirs which changes the gravitational pull between the shrinking ice mass and surrounding ocean, the Earth's rotation, as well as the instantaneous (elastic) component of the glacial isostatic response. The changing mass of each major ice sheet or large glacier would result in a geographically distinct pattern of sea-level change, referred to as a “sea-level fingerprint” (Mitrovica et al. 2010, 2011; Spada 2017). In turn, the geographically variable sea-level trend can help pinpoint the meltwater source.

Dynamic Changes

Atmospheric and oceanographic processes are responsible for fluctuations in sea level that last from 1 to 2 years to a decade or more and are coherent over long distances. These fluctuations arise from ocean currents and closely coupled oceanographic-atmospheric phenomena, such as the Northern Atlantic Oscillation (NAO) or the El Niño–Southern Oscillation (ENSO) in the Pacific Ocean (see “► [El Niño–Southern Oscillation \(ENSO\)](#)”; Table 1). During the El Niño phase of the ENSO cycle, anomalously warm water propagates eastward, splitting into north and south poleward-propagating Kelvin waves, causing sea level to rise (for several months) along the west coasts of the Americas (Tingstad and Smith 2007). During the positive phase of the NAO cycle, the pressure gradient between the Icelandic Low and the Azores High intensifies, strengthening prevailing westerly winds across the North Atlantic. Storm tracks move farther north-east than usual; Great Britain and Scandinavia experience milder but stormier than average winters. Higher storm surges generated by the strong westerlies combine with

reduced atmospheric pressure and above-average sea surface temperatures to raise regional sea level during the event (Woodworth et al. 2007). These fairly short-lived dynamic ocean fluctuations generally do not lead to permanent change in average ocean height. Yet, their effects on the coastal zone and its inhabitants, superimposed on a rising global sea level, can have profound consequences (see “► Natural Hazards”).

Conclusions

Sea-level change is a complex phenomenon, varying considerably over space and time. Nonetheless, considerable progress has been made in modeling of glacio-isostasy and in accurate dating of Holocene sea-level proxies. This information will enable a clearer separation of the present-day eustatic (global) sea-level signal from regional to local geophysical and oceanographic factors.

Cross-References

- [Changing Sea Levels](#)
- [Coral Reefs](#)
- [Coral Reef Coasts](#)
- [Coral Reefs, Emerged](#)
- [Geochronology](#)
- [Geodesy](#)
- [Isostasy](#)
- [Natural Hazards](#)
- [Sequence Stratigraphy](#)

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Eustasy and Sea Level

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Original and Modern Definitions of Eustasy

The concept of eustasy has been defined as the global (or “absolute”) sea level and/or synchronous uniform worldwide changes in mean sea level (Neuendorf et al. 2005). Considering marine regressions during glaciations and interglacial transgressions, Suess (1888) introduced this Greek term for “well fixed,” or “static,” referring to the assumed globally synchronous water levels in the world ocean with implied permanent and rigid basin walls and floor. This concept emerged decades before the very existence, let alone knowledge of the viscous-elastic nature of the mantle-crust system became known. Solid Earth deformation and associated plate tectonic processes are the essential features of eustatic phenomena (Rovere et al. 2016b). The complex interrelationship between the viscous-elastic crust-mantle complex, fluctuating ocean levels, basin dimensions, and the gravitational and rotational influence of the variable ice and water loads on the crust are key aspects of eustasy.

Eustatic (ESL) and Relative Sea Level (RSL)

Eustasy in the current definition essentially involves fluctuations that take place in water volumes located within and on

the crust of the Earth. The term eustasy in contemporary thinking corresponds to mean global sea level without the impact of local factors. The concept refers to the position of the sea surface as related to a fixed datum, such as the center of the earth or a satellite in fixed orbit around the earth. However, because this position varies due to various “eustatic changes,” a somewhat inconsistent term, it is impossible to determine the size of past variations in the eustatic position of sea level (Burton et al. 1987). Constant changes in the uneven topographic and bathymetric configurations of the ocean basins lead to continuous changes in the shapes and dimensions of these basins. Any observed and measured sea level alteration related to a land-based reference frame, i.e., uplift and subsidence of land masses, is defined as a relative RSL change that modifies eustatic, global sea level positions (Rovere et al. 2016b). Changes are due to vertical displacement of the global ocean surface and of the surrounding land masses as well as additional factors. Movements occur either in the same or opposite direction. In the general literature, however, the term “eustatic change” often is limited to ice-equivalency-related glacial and deglacial fluctuations in the level of the World Ocean.

Global and Local Sea-Level Changes

As opposed to the theoretically conceived eustatic global mean sea level (ESL), “eustatic fluctuations” thus are not worldwide and uniform in extent. The mechanisms of changes that operate over local and spatial scales during specific time intervals are not considered eustatic. They result from temporal and local changes in water volumes in the world ocean and in mean global sea level (Gornitz 2005, Miller et al. 2005). A purely eustatic signal can not be observed in isolation-eustatic sea level therefore is not a “tangible” physical entity with a spatial scale. Satellite technology will establish quantitative global changes in the future. At present, the eustatic signal still provides a reference value regarding the unscaled global eustatic mean sea level. Eustatic sea level thus may be regarded as a general climate change index in terms of melting and freezing ice cover, as well as oceanic temperature and salinity variations. It is also modified by changes that occur during other temporal processes related to variations in ocean level, water volume, water mass, as well as other factors. The following changes, often described as “eustatic,” are prime modifiers of unscaled global mean sea levels (Rovere et al. 2016b).

“Eustatic fluctuations” are caused primarily by water volume changes in the world ocean and in mean sea level (Gornitz 2005; Miller et al. 2005), as well as by changes that occur during other temporal processes related to variations in ocean level, water volume, and water mass. The following

factors contribute to major sea level changes, often designated as “eustatic”:

1. *Glacio-eustatic changes* occur by the accumulation or disintegration of global ice sheets. They denote melting and growth of continental ice volumes over time (Daly 1934; Peltier 1999; Lambeck et al. 2012). Periodic massive meltwater flow from thawing continental ice sheets and glaciers played a singular role during the numerous Quaternary eustatic sea-level rise events.
2. *Hydro-eustatic changes* (Hay and Leslie 1990; Gornitz 2005; Khan et al. 2015) refer to water redistribution between different global reservoirs in the form of snow, surface water, soil moisture, and ground water. The changes do not include the continental glacial meltwater effects.
3. *Thermosteric and halosteric density changes*; water volume contraction or expansion result from changes in temperature and salinity, respectively, due to ocean water cooling or warming and salinity increases or decreases (Gornitz 2005; Rovere et al. 2016).
4. *Tectonic eustatic changes* refer to long-term, mostly tectonics-related ocean basin volume reduction by ocean water displacement during magma and lava intrusion or an increase in basin capacity due to newly created accommodation space. These processes involved slow and fast rates of sea-floor spreading, mid-ocean ridge formation, and plate subduction, including ocean trench development (Rona 1995; Leonard et al 2004; Miller et al. 2005).
5. *Sediment eustasy*. Displacement of water by accumulation of large sediment volumes on the seafloor along continental margins and in ocean basins also raises the ocean level (Gornitz 2005).
6. *Geoidal eustasy, continental ice cover, and gravity anomaly changes*. Geoidal eustasy refers to ocean level changes caused by altered global distribution of ocean water due to geodetic changes in ocean level (Mörner 1976; Mitrovica and Peltier 1991). Planet Earth is broadly flattened at the poles, bulging at the equator, and the ocean surface does not parallel the Earth surface. The elevation of this “geodetic sea level,” with respect to the distance to the center of the Earth, varies by as much as 180 m. While the continental crust underwent viscoelastic rebound due to deglaciation, the low-latitude gravity anomaly that evolved during the preceding peak glacial stage had vanished. Its disappearance drove water masses of the oceanic geoid to migrate from lower to higher Earth latitudes. Consequent sea level changes in the last four millennia involved the north Atlantic region during the current Late Holocene highstand. Low-latitude oceanic islands, such as the Cocos (Keeling) Islands, experienced concurrent RSL decline, the emergence of marginal new land areas from the eastern Indian Ocean (Woodroffe and McLean 1990; Table 1).

Eustasy and Sea Level, Table 1 Summary of processes affecting global sea level (Gornitz 2005, p. 440)

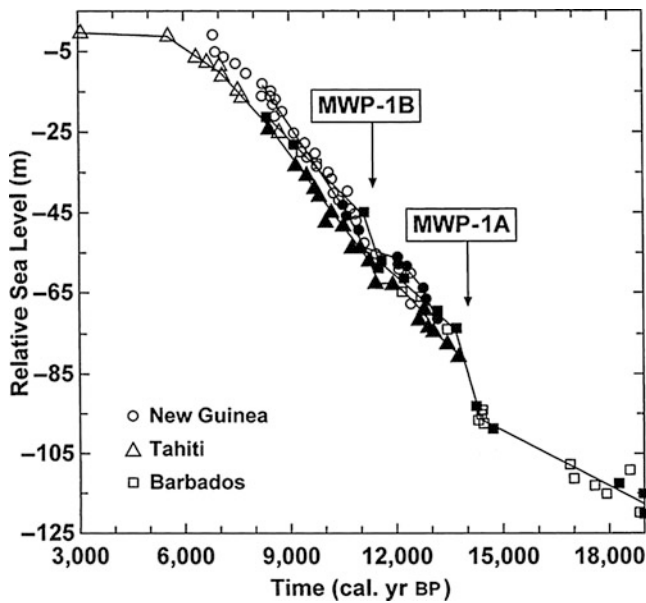
Process	Rate (m/ka)	Duration (year)	Amplitude (m)	References
Change in ocean water quantity				
Glacio-eustasy	10	10^4 – 10^5	120–150	Bard et al. (1996); Fairbanks (1989)
Hydro-eustasy	0.5–5	10^4 – 10^5	~50	Hay and Leslie (1990)
Steric changes	I	10 – 10^4	10	Rona (1995)
Change in ocean basin volume				
Tectono-eustasy				
Ocean ridges	0.01	108	500	Rona (1995)
Plate collisions	0.002	107	100	Rona (1995)
Hot spots	0.001	108	100	Rona (1995)
Sedimento-eustasy	0.003	108	80	Rona (1995)
Glacio-isostasy	1–10	104	100	Peltier (1999)
Dynamic changes				
Geostrophic currents	–	1–10	0.01–1	Gornitz (1995)
Atmospheric forcing	–	1–10	0.01	Gornitz (1995)
ENSO	–	3–8	0.1–0.5	Gornitz (1995)

Relative Sea-Level Change Factors

The interplay between eustatic sea level and factors that determine RSL triggers relative sea-level changes (Figure 1). While uplift or subsidence do take place in the adjacent land areas, the volume and mass of ocean water bodies remain unchanged. Interaction between eustatic and relative sea-level changes create a variety of geomorphic markers that are critical in the evaluation of relative sea-level changes, as well as the local transgression and regression history in a given coastal region (Baker et al. 2001; Miller et al. 2005). Direct and indirect markers include marine notches carved in bedrocks, erosional and aggradational marine coastal terraces, submerged and emerged coral reefs and other shallow subtidal fauna and flora elements, interface between sand units with subtidal-intertidal and supratidal (eolian) sedimentary signatures, and others (*q.v.*, ► [Gulf Shorelines, Last Eustatic Cycle](#) – Encyl. Chapter by Otvos).

Glacio-Isostatic Adjustment (GIA) and Hydro-Isostatic (Added/Reduced Water Load) Effects

These processes occur before, during, and following ice sheet establishment and melting. GIA is defined as the “viscoelastic response of the Earth to the redistribution of ice and ocean loads” (Dragert et al. 1994; Peltier 1999; Woodroffe and Horton 2005; Lambeck et al. 2012). Depending on the different stages of the process, these factors are related to ice



Eustasy and Sea Level, Fig. 1 Sea-level curve of the post-glacial marine transgression. Based on radiocarbon and U-Th dating of corals from Tahiti, Barbados, and New-Guinea (Gornitz 2005, modified from Bard et al. 1996)

loading, as well as melt water loading and unloading. Ice sheet growth results in subsidence beneath continental ice masses. Responding to loading, viscous mantle matter flows away from ice-loaded regions and creates an uplifting peripheral bulge in the “near-field” of the ice margin (Daly 1934). Mantle flow starts under loaded crustal areas and creates an uplifting peripheral forebulge in the “near field,” close to formerly glaciated regions. Even in remote “far-field” regions, at great distances from glaciation centers, water removal from the ocean results in crustal uplift due to reduced hydrostatic pressure by overlying waters. During and following crust relaxation, “unloading,” as melting disposes of the continental ice sheets, the flow direction of the viscous mantle matter reverses toward the formerly ice-loaded site. The peripheral forebulge decays. Forebulge subsidence controls sea-level changes even in “far-field” equatorial areas (Khan et al. 2015). Crustal unloading results in isostatic uplift at rates that may reach 5–10 cm/year. Ocean waters fill bathymetric lows as they gradually develop in the process of forebulge collapse. This “equatorial ocean siphoning” effect (Mitrovica and Peltier 1991) led to RSL decline and the consequent Mid-to-Late Holocene highstands of Pacific islands (Woodroffe and Horton 2005; Rovere et al. 2016a, b).

Differences in the width of continental shelf and in its impact on hydro-isostatic adjustment during the Late Holocene caused a differential crustal flexure under Australia’s eastern coastal region. This has also defined the Mid-Holocene time of the marine highstand in the area (Flood and Frankel 1989; Baker et al. 2001). “Continental levering” occurs in the final stage of post-glacial sea-level rise along the

continental margins in far-field areas. Upward tilt of continental coastal sectors takes place in response to increasing water load in ocean basins. The seafloor subsides and the viscous upper mantle matter is pushed toward the continents. At the same time, due to lithosphere flexing, coastal areas may experience local uplift with consequent RSL decline (Woodroffe and Horton 2005).

Rapid ice sheet melting that leads to sea-level decline within a few thousand kilometers of the ice sheet and a progressively increasing rise in sea level at greater distance in tandem effects of crustal rebound, involves changes in gravitation attraction as well. The largely elastic deformation involves diminution of the gravitational pull that melting ice sheets exercise on the ocean waters. Rotational effects of Earth do also contribute to the process (Milne and Mitrovica 1998; Creveling et al. 2015). When an ice sheet grows, the mutual gravitational pull attracts water masses. When the ice sheet melts, the pulling forces on the water decrease, with a net effect of RSL fall in the near field and an RSL rise in the far field (Woodroffe and Horton 2005; Rovere et al. 2016a and other sources).

Additional RSL changes and their causes (Lambeck and Nakada 1990; Gornitz 2005; Woodroffe and Horton 2005; Rovere et al. 2016a, b; Rona 1995) are as follows:

1. **Interseismic land uplift and coseismic land subsidence** along subduction zones due to divergent and convergent mantle current flow, respectively.
2. **Volcanic isostasy** caused by increased volcanic loading on the crust. Current flow within mantle displaces matter away from loaded sites.
3. **Sediment compaction**-related seafloor subsidence increases accommodation space to be occupied by ocean water. This occurs particularly in marginal ocean areas underlain by a thick sequence of unconsolidated deposits (e.g., Mississippi and Nile deltas). Sediment redistribution, erosion from land, and redeposition in ocean do not change the ESL either.
4. **Karst solution-unloading phenomena.** Loss by karst terrain reduction and lowering, cavity formation and collapse influence RSL by dissolution and consequent compaction processes (Rovere et al. 2016b).

Interaction between eustasy and local RSL forcing factors – examples (q.v., Encyclopedia Articles, ► Modes and Patterns of Shoreline Change, ► Gulf Shorelines, Last Eustatic Cycle Chapters)

Specific Examples of Sea Level Fluctuations

Combined eustatic and RSL factors were responsible for the multimillennial time lag between the start of the Holocene

highstand in the Indo-Pacific realm and the North Atlantic region. By 5–6 ka B.P. when the RSL of the Gulf of Mexico stood at ~ -4 m, Pacific mean sea level reached +2–3 m at various Australian and Pacific island sites (Pan et al. 2018; Otvos, Last Interglacial chapter, present volume). Recorded low sea levels during the Glacial Maximum, 18 ka B.P., appear at various depths below present sea level. These depth figures differed between various locations by as much as 30 m (Nakada and Lambeck 1989; Lambeck and Nakada 1990; Woodroffe and Horton 2005). Isostatic and other RSL-related factors have resulted in the uplift of Late Pleistocene (MIS 5c Substage) US Atlantic coast terraces from intermediate to Last Interglacial (LIG) record elevations (e.g., Wehmiller et al. 2004; Rovere et al. 2016a). Anomalously high (+6 to +9 m) Late Pleistocene LIG sea-level markers in areas generally considered tectonically stable in South Florida, South Africa, the granitic Seychelles Islands “mini-continent” in the western Indian Ocean, and parts of Australia (e.g., Flood and Frankel 1989; Baker et al. 2001; Muhs et al. 2011; Creveling et al. 2015) are yet to be adequately explained by eustatic RSL models. These sites may have been affected by local mantle current upwelling.

Dynamic changes (Table EB) are causing short-term fluctuations in sea level due to coupled ocean-atmosphere variations. They include El Niño, La Niña, ENSO, SOI, AMO (Atlantic Multidecadal Oscillation), NAO (North Atlantic Oscillation), and several other global and regional oceanographic-weather cycles (Gornitz 2005, and others). An indirect impact by increased or decreased precipitation over ocean basins also influences ESL (Rovere et al. 2016). (q.v., Encyclopedia Articles: *Last Interglacial*; *Barrier Island Formation and Development*).

Summary

Eustasy relates to global water volume and mass changes in the world's ocean driven by a variety of processes, primarily glacio- and hydroeustatic in nature. Isostatic effects, manifested among others, by peripheral forebulges created by continental ice load, as well as their decay effect near, intermediate, and far-field settings. They are linked to regional uplift and subsidence processes. Isostasy, combined with other factors, helps to explain major differences between Quaternary sea-level curves established in various regions on the planet. Glacial isostatic adjustment, a major RSL component makes the observed local sea level change different from a theoretically established global mean sea level (eustatic) value. The record observed at any given field site is related to RSL change. Its value differs from the mean global eustatic value. At the same time, somewhat inconsistently, the change itself is attributed to the mechanism of “glacio eustasy”. In addition to isostatic processes equatorial ocean siphoning,

changes in ice-sheet related gravitational anomalies, and releveling do also influence relative sea level positions and their changes. Lithospheric plate subduction, oceanic seafloor spreading, mid-oceanic ridge, and deep oceanic trench formation, potentially also mantle current up- and downwelling are among the most important, long- and very long-term tectonic influences that impact ocean basins and global eustatic changes. “Eustatic” changes, most validly defined, are already monitored by the new global satellite altimetry.

Cross-References

- ▶ Continental Shelves
- ▶ Eustasy
- ▶ Glaciated Coasts
- ▶ Isostasy
- ▶ Seismic Displacement
- ▶ Tectonics and Neotectonics

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Faulted Coasts

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Faulted coasts are chiefly associated with active continental margins where tectonic plates are colliding. They often correspond to bold coasts, characterized by continuously steep and straight scarps. However, terraced coastal areas dislocated by faults as well as fiord or estuary coasts controlled by faults have also to be included into the broad category of faulted coasts.

Sea Cliffs and Fault Scarps

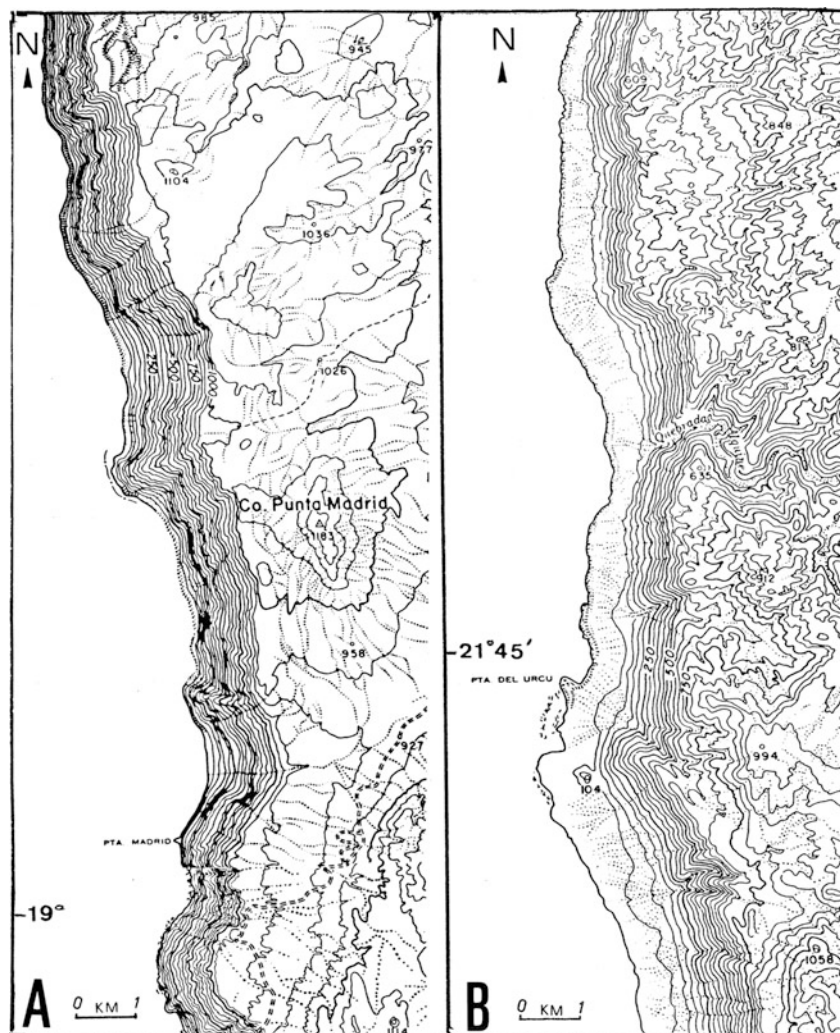
In some instances, coastal cliffs may be identified as fault scarps of direct tectonic origin which separate raised blocks of land from dropped ones that have been depressed below sea level. Plunging cliffs, cliffs which are not fronted by shore platforms or beaches and plunge abruptly into deep water, can be the product of recent faulting if down-thrown blocks have subsided to appreciable depths. In that case, and not only in sheltered waters and even if the resistance of rocks is not considerable, plunging conditions could persist long after faulting and the initial profile of the coast may long remain practically unaltered (Johnson 1919). Cliff recession is very slow since erosional processes are inhibited by the reflection of incoming nonbreaking waves against vertical or nearly vertical walls rising out of deep water. Furthermore, deep waters at the cliff base prevent accumulation of sediment derived from alongshore transport or subaerial scarp weathering, resulting in the lack of abrasive materials which could increase the erosive power of the waves. However, in the course of time, hydrostatic pressure and hydraulic

quarrying caused by the rise and fall of standing waves, and the compression of air in rock fractures eventually are able to produce notches and caves. Narrow erosional platforms may subsequently develop, allowing breaking waves to attack the cliff and destroy the plunging condition (Trenhaile 1987).

Several coastal cliffs in New Zealand are thought to be actual fault scarps (Cotton 1916). For instance, the cliffs along the Wellington fault, on the northwestern shore of Port Nicholson have been explained in this way. They slope at about 55°, show little evidence of marine modification at the shore level, and descend to a depth of about 20 m, close and parallel to the shoreline. In Russia, on the Murmansk coast, to the west of the White Sea, Archean granites and gneisses are faulted along a line which almost coincide with the present coastline. The foot of the cliffs extends to a depth of several tens of meters without alteration of slope. During storms, waves are reflected from the scarp wall and there are no wave-cut notches or fragments of benches (Zenkovich 1967). In the United States, some steep cliffs along the coast of California are attributable to fault scarps. For example, a down-thrown block has dropped below sea level on the northeastern side of San Clemente Island, forming a straight cliff with no intervening shelving bottom (Shepard and Wanless 1971). The Gulf of California, Mexico, gives nice examples of recent faulted coastlines, especially in the islands next to its western side where straight, continuous, and precipitous escarpments display faceted spurs and truncated structures, are not bordered by shelves, have deep water at their base, and show angular contact with the adjacent submarine basin (Shepard 1950). All these features are evidence of no marine erosion after faulting that is probably recent. For instance, the sea cliff on the southwestern side of Angel de la Guarda Island corresponds to a fault scarp which steeply plunges to the bottom of a trench located at a depth of about 1,000 m. In volcanic areas, cliffs can be directly produced by faulting induced by vulcanicity. Sea cliffs devoid of a bordering shelf on the southern flank of Kilauea Volcano, in Hawaii, can be identified as

Roland P. Paskoff: deceased.

Faulted Coasts, Fig. 1 The very high cliff of northernmost Chile corresponding to a major fault-line scarp. (From topographic maps of the *Instituto Geográfico Militar* of Chile). (a) active cliff, (b) abandoned cliff with emerged wave-cut terraces at its base



Faulted Coasts, Fig. 2 The very high cliff of northernmost Chile near 21°25'S lat. The toe of the scarp is covered by thick screens which are the result of salt-weathering and seismic shocks. Wave-cut platforms extent at its base. (Photo: R. P. Paskoff)



direct normal fault scarps although, at least in some cases, they may correspond to gigantic slump scarps.

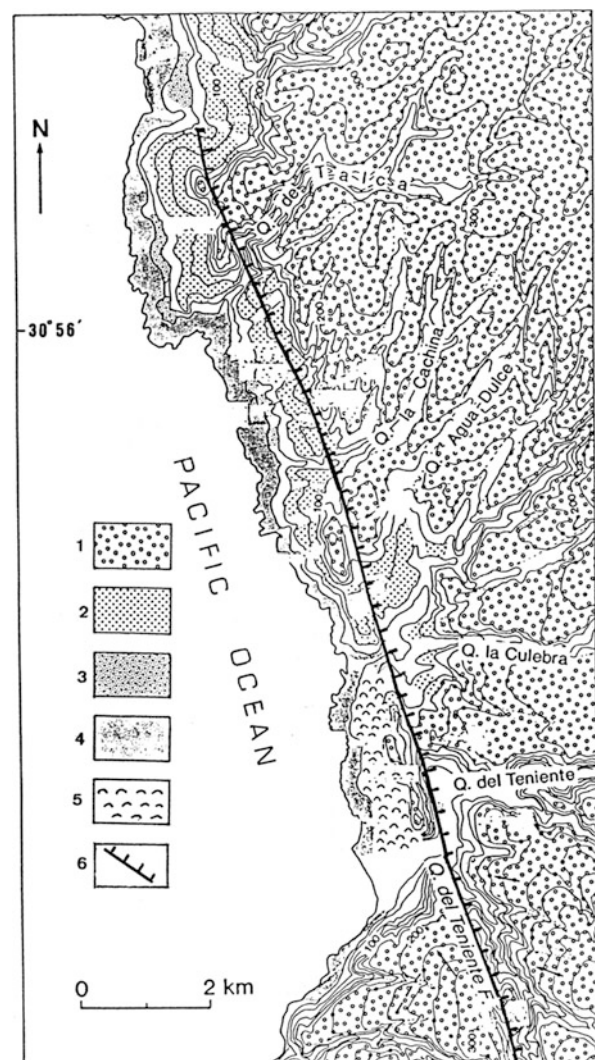
Sea Cliffs and Fault-Line Scarps

Most tectonically controlled coastal cliffs correspond to fault-line scarps that are scarps produced by structurally induced erosion at ancient faults. As a matter of fact, on the one hand there has been little time for important faulting since the post-glacial rise of sea level, on the other hand older original tectonic features are not retained for long because of the erosional effects of subaerial geomorphic processes during periods of low sea level. For example, along the steep Santa Lucia Mountain coast of central California, the continental shelf is in some parts less than 0.6 km wide, suggesting that this is a fault-line coast since a retreat of a relatively short distance, due to marine erosion, has followed faulting. In Japan, the straight mountainous coastline which characterizes the northwestern part of Shikoku stretches along the so-called Median Tectonic Line. In the Malta archipelago, the coastline pattern of the main island is mainly controlled by faulting. Here, in the northern part, bays correspond to downthrown blocks that were partially submerged. In the southwestern part, spectacular cliffs which can be more than 200 m high are erosionally derived from a major northwest trending fault, known as the Maghlaq fault that has been displaced vertically by at least 230 m and probably moved during the Quaternary (Paskoff and Sanlaville 1978).

In northernmost Chile, where the oceanic Nazca plate is subducted under the continental South American plate, the extremely arid coast is characterized, over more than 800 km, by a striking, precipitous, northward trending cliff which is 700 m high on average. Because of its length and height, this scarp represents a major geomorphic feature at a worldwide scale. From Arica (18°S lat.) to Iquique (20°S lat.) the cliff is almost vertical and its top can reach 1,000 m above sea-level. At present, it is an active feature, dropping into the sea and receding under marine action. As a result, ephemeral streams of wadi type are cut by the scarp and hanging above sea level (Fig. 1a). Landslides, triggered by earthquake shocks and affecting rock masses previously loosened by intense tectonic fracturation, are also contributing to the cliff retreat. From Iquique down to 27°S lat., the cliff remains very high but not so steep. As a matter of fact, it is no longer an active one since a belt of emerged wave-cut platforms, several hundred meters wide, extends at its base, indicating a relative fall in sea level (Figs. 1b and 2). The cliff of northernmost Chile is a typical fault-line scarp (Paskoff 1978). Detailed geological mapping failed to reveal faulting at the foot of the cliff whose lack of straightness is also inconsistent with a direct tectonic origin. Actually, the cliff results from an important faulting phase which generated major, north-south trending, *en*

échelons, normal faults at the end of the Oligocene Epoch and uplifted the continental block which constitutes the present Coastal Cordillera. Subsequently, these faults retreated under marine erosion, over about 10 km which is here the width of the continental shelf, mainly during a major middle to upper Miocene transgression. Later differential crustal movements occurred.

This explains why the cliff is an active one in its northern part where subsidence may be related to a collapse of the continental margin toward the Chile-Peru trench, and an abandoned one in its southern part where, on the contrary, an uplift motion has been taking place.



Faulted Coasts, Fig. 3 Geomorphological sketch of the southern part of the Talinay Heights coastal area in northcentral Chile (modified from Ota et al. 1995). (1) Talinay I wave-cut terrace (Pliocene?), (2) Talinay II wave-cut terrace, (3) Talinay III wave-cut terrace (oxygen isotopic stage 9?), (4) Talinay IV wave-cut terrace (oxygen isotopic stage 5), (5) dunes, (6) Quaternary fault

Coastal Terraces and Faulting

Former wave-cut platforms that have become emerged because of a fall of sea level due to some combination of eustatic and crustal processes are not unusual on coastal belts. When located on active margins, these elevated marine terraces generally show evidence of active faulting which has occurred in Quaternary time (Pirazzoli 1994).

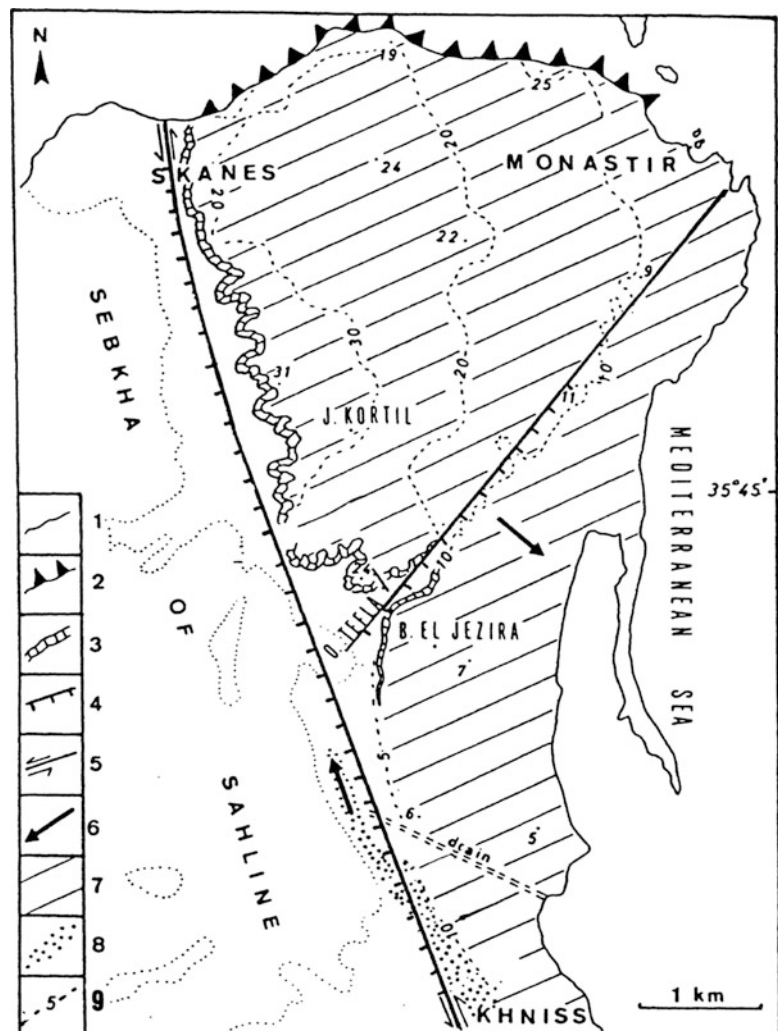
In north-central Chile, the 110-km-long coast of the Talinay Heights area, between 30°S and 31°S lat., is characterized by a flight of several emerged Plio-Quaternary marine terraces which has been densely and strongly faulted, uplifted, and tilted (Ota et al. 1995). Dip-slip normal faults, which are early Pleistocene in age, have a clear topographic expression, forming conspicuous scarps. The most significant one is the Quebrada del Teniente fault (Fig. 3). Distinctively straight, NNW–SSE trending, it runs parallel to the present shoreline and crosses the entire area over a length of about 60 km, dislocating the terraces and generating a striking

east-facing scarp. The maximum amount of vertical displacement reaches 52 m.

In eastern Tunisia, the Monastir surroundings (36°N lat.) are wellknown for their Tyrrhenian (name used in the Mediterranean for the last interglacial marine episode, i.e., oxygen isotopic stage 5) deposits and for the term Monastirian which was incorporated to the standard Mediterranean terminology for Quaternary marine shorelines early in the twentieth century. For a long time Monastir was the type locality for the 18–20 m terrace. This classic, widely quoted term was finally and definitely dropped when it was established that crustal deformation had occurred near Monastir (Paskoff and Sanlaville 1981). The area is characterized by a triangular wave-cut platform (Fig. 4) which dates back to the late Quaternary (oxygen isotopic substage 5e) and has been affected by two main faults. The Wadi Tefla fault which trends north-east is a normal fault which dislocates the platform in two levels and has a maximum vertical throw of 15 m. It was erroneously considered to be a former sea cliff separating two

Faulted Coasts,

Fig. 4 Geomorphological sketch of the Monastir area in eastern Tunisia (modified from Paskoff and Sanlaville 1981). (1) low rocky coastline, (2) sea cliff, (3) erosional scarp of tectonic origin, (4) normal fault, (5) strike-slip fault, (6) tilting, (7) wave-cut platform dating back to the last interglacial age (oxygen isotopic substage 5e), (8) former coastal barrier dating back to the last interglacial age (oxygen isotopic substage 5e), (9) contour line in meters



distinct marine terraces. The NNW–SSE Skanes–Khniiss fault is a major fault of regional significance which displays a vertical displacement that reaches 50 m. It also shows a sinistral strike-slip component characterized by an horizontal movement of at least 500 m. The fault has been active in historical times since it cuts a Roman mosaic on a floor of a house which was built in the second century AD

Fiord or Estuary Coasts and Fault Control

In plan view, coasts developed on intrusive and metamorphic terrains may show intersecting linear patterns of fiords or estuaries, depending on the latitude, marking former glacial or fluvial erosion exploiting the structural weakness of faults. For instance, in southernmost Chile, especially between 44°S and 55°S lat., fiords regionally named *canales*, penetrate inland. They correspond to glacial troughs which were eroded and overdeepened during the last Pleistocene glaciation by active Andean glaciers, mainly in Jurassic granitic rocks. The fiord pattern is strikingly controlled by fault lines which are mainly oriented N–S, NE–SW, and NW–SE. Fault lines were selectively exploited by eroding glaciers. Fiords form a dense network of long marine arms, characterized by straight stretches and right angles (Paskoff 1996).

As already stated by Mclean (1982), the concept of faulted coast has now to be used in a broader sense than initially defined by the pioneers of coastal geomorphology. It encompasses not only cliffs controlled in some way or other by faulting, but also landforms found in the coastal belt which are directly or indirectly related to lines along which vertical or horizontal movements of tectonic origin have taken place.

Cross-References

- [Changing Sea Levels](#)
- [Cliffed Coasts](#)
- [Cliffs, Erosion Rates](#)
- [Geochronology](#)
- [Shore Platforms](#)

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Geochronology

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Definitions

In coastal research, the terms “geochronology” or “geochronostratigraphy” are used to determine time scales for coastal processes and coastal evolution utilizing relative stratigraphic techniques (morpho-, pedo-, bio-, lithostratigraphy) and absolute dating methods, whereas relative and absolute approaches complement each other.

Dating Methods

In the last 30 years, applications of the numerical (“absolute”) dating of coastal structures, sediments, and processes were improved, new dating methods were established, and the precision and accuracy of existing age determination methods were considerably increased. Numerous papers and some books describe dating techniques in Quaternary sciences in detail (e.g., Rink and Thompson 2015; Malainey 2011; Walker 2005).

Mass spectrometric radiocarbon, thorium-230/uranium-234 isotope (Th/U), electron spin resonance (ESR), optically stimulated luminescence (OSL), and terrestrial cosmogenic nuclide (TCN) dating are the most commonly used absolute dating techniques in coastal research (Table 1). In addition, paleomagnetic dating has proved useful for a differentiation of marine and littoral sequences of the Middle and Early Quaternary (e.g., Garces 2015; Jovane et al. 2013), and the

amino acid racemization (AAR) method is a relative chronostratigraphic method that needs calibration with other numerical approaches like radiocarbon or Th/U dating (e.g., Demarchi and Collins 2015; Wehmiller 2013).

Radiocarbon (¹⁴C) Dating

The importance of the radiocarbon method for dating Holocene and Late Pleistocene marine terraces, beaches, and other coastal features cannot be overestimated. It is the most common and widely applied dating method in geochronology.

Carbon-14 (¹⁴C) is a radioactive carbon isotope with a half-life of 5730 ± 40 years. However, ¹⁴C ages are calculated using the original Libby half-life of 5568 ± 30 years. The ¹⁴C isotope is produced in the upper atmosphere by interactions between nitrogen atoms (¹⁴N) and cosmic rays. As part of CO₂ molecules, ¹⁴C mixes with atmospheric carbon in the form of the stable isotopes ¹²C and ¹³C. The natural concentration of radioactive ¹⁴C in the atmosphere is approximately 1.176×10^{-12} atoms per ¹²C atom, and the ¹⁴C content in living organisms is in equilibrium with the atmospheric ratio or their specific environment (e.g., freshwater, saltwater). The CO₂ exchange between the organism and its environment discontinues with death, and ¹⁴C decays to ¹⁴N. Radiocarbon ages are expressed in years ¹⁴C before present (BP), with BP referring to the time before AD 1950, since human-induced nuclear detonations in the 1950s and 1960s significantly changed the atmospheric ¹⁴C composition.

There are two laboratory methods of radiocarbon dating: the conventional decay counting method, which measures beta rays from the ¹⁴C decay process, and the accelerator mass spectrometry (AMS) technique, which determines the amount of ¹⁴C particles in a sample directly. The AMS technique is more sensitive than the decay counting method, and because of this, counting times (minutes to hours, instead of days to months) and sample weights (less than 1 mg C) are greatly

Geochronology, Table 1 Some important dating methods in coastal research (⁺frequently used, but little suited for this dating method)

Method			Dating material	Time scales to be dated	Major problems influencing dating results
<i>Absolute dating methods</i>	<i>Isotope dating methods</i>	<i>Radiocarbon (¹⁴C)</i>	All carbon-containing material, e.g., wood, charcoal, peat, mollusk shells*, coral*, foraminifera*, ostracods*	Holocene up to Late Pleistocene conventional ¹⁴ C method: 20 to 30 ka ¹⁴ C BP; AMS ¹⁴ C method: approx. up to 50 ka ¹⁴ C BP	Variability in atmospheric ¹⁴ C (cosmic, Suess and nuclear bomb effects); contamination of samples by younger C-bearing material; * marine reservoir effect
		<i>Thorium-230/Uranium-234 (²³⁰Th/²³⁴U)</i>	Coral, mollusk shells ⁺ , bones ⁺ , teeth ⁺ , speleothems ⁺ , travertines ⁺	Holocene up to Late Middle Pleistocene coral: approx. up to 200 ka BP. Mollusk shells: approx. up to 130 ka BP (less reliable)	⁺ Isotope mobilization, esp. uranium migration (radiometric system may not be closed); variability of ²³⁴ U/ ²³⁸ U ratio in seawater
		<i>Terrestrial cosmogenic nuclides (TCN)</i>	Quartz (¹⁰ Be, ²⁶ Al) and carbonate (³⁶ Cl)-bearing sediments and surfaces	Holocene to Pliocene	Inheritance of nuclides from pre exposition, eroding surfaces, large background levels in rocks
	<i>Radiation induced dosimetric dating methods</i>	Optically stimulated luminescence (OSL)	Dune sands, littoral sands (and cobbles), storm and tsunami deposits	Holocene up to Late to Middle Pleistocene at least up to 300 ka BP	Incomplete bleaching; isotope mobilization, esp. uranium migration; variations in water contents
		Electron spin resonance (ESR)	Coral, mollusk shells, foraminifera, some gastropods	Holocene up to middle Pleistocene coral: up to 400–600 ka BP. Mollusk shells, foraminifera, and gastropods: up to 200–300 ka BP (upper dating limit still unknown)	Isotope mobilization, esp. uranium migration; recrystallization of sample
<i>Relative dating methods</i>		<i>Paleomagnetic dating</i>	All material with stable magnetic remanence, e.g., fine-grained coastal sediments	Middle and Early Quaternary, limited to paleomagnetic reversals and events	Destruction of primary magnetic signal and acquisition of secondary magnetization
		<i>Amino acid racemization (AAR)</i>	Mollusk shells	From several 100 a to several 100 ka BP	Results require calibration through absolute dating techniques; racemization rates may vary between genera; diagenesis temperature histories may differ between geographic regions

reduced. Dating is theoretically limited to ca. 50,000 years, however, due to sample or laboratory contamination rather to 30,000 years.

Since the 1960s, the radiocarbon method has been used successfully for dating Holocene and Late Pleistocene samples. The ¹⁴C method delivers reliable results for wood, charcoal, and plant matter (e.g., peat, seeds, leaves), while the dating of humus, speleothems, calcretes, bones, and teeth is frequently inaccurate. In coastal research, the ¹⁴C dating method is widely applied to the dating of marine macrofauna (e.g., mollusk shells, coral) and microfauna (e.g., foraminifera, ostracoda). Radiocarbon dating is also used to date the formation of coastal marshes and peat layers that are used, e.g., as archives for sea-level evolution (Kemp et al. 2017; Gehrels 1999), and it is an excellent tool for geoarchaeological research in coastal settings where it can be cross-checked with relative chronologies based on artifacts (Brückner 1997, 2003).

Variations in the production of atmospheric ¹⁴C may be caused by sunspot activity, changes in the intensity of the Earth's magnetic field, or variability in cosmic radiation. These sources of error can be reduced by calibrating ¹⁴C dating results (at present the calibration table covers the last 50,000 years) with a dendrochronologic tree ring calendar (dendrochronology: an absolute dating method based on the counting of tree rings); with Th/U-dated corals, speleothems, and foraminifera; and with varved sediments (Reimer et al. 2013). In addition, marine samples may be influenced by the so-called marine reservoir effect, which is caused by the variability in the ¹⁴C concentration in seawater. Deep ocean water has lower ¹⁴C contents than the atmosphere. The ¹⁴C dating of marine fauna (mollusks, coral, microfauna) from shorelines influenced by upwelling deep ocean water results in ¹⁴C ages that are too old. The global mean reservoir effect for marine carbonates is approximately 400 years but varies spatially and temporally.

Case Study: ^{14}C Dating of Beach Ridges in Spitsbergen

The littoral zone of northern Andréelund (northern Spitsbergen) is characterized by more than 100 beach ridges ranging in elevation from present sea level to 70 m above. They represent uplifted former shorelines (Brückner and Schellmann 2003; Brückner et al. 2002; Forman et al. 2004). Two conditions led to their formation: (a) The West Spitsbergen Current, the northernmost continuation of the Gulf Stream extending up to 80°N , created seasonally ice-free conditions in fjords. This allowed wave action, which caused the formation of beach ridges. (b) The current elevation of these beach ridges is a result of coastal uplift due to the glacio-isostatic rebound following the last deglaciation.

The radiocarbon dating results of mollusk shells (Fig. 1) illustrate (a) the onset of deglaciation and the beginning of beach ridge formation in the Allerød Interstadial ($11,970 \pm 60$ years ^{14}C BP, minus the reservoir effect of approximately 400 years) due to the arrival of the West Spitsbergen Current; (b) that fjords were blocked by sea ice during the Younger Dryas cold phase (this period of no beach ridge formation ranged approximately from 11,100 to 10,600 ^{14}C BP (minus the reservoir effect; see above)); and (c) that the major glacio-isostatic uplift occurred between 12,000 and 9200 years ^{14}C BP (minus the reservoir effect; see above) and amounted to more than 140 m. At approximately 12,000 ^{14}C BP, eustatic sea level was at least 70 m lower than today. Since then, the oldest beach ridges have been uplifted from less than 70 m below up to 70 m above present sea level.

Thorium-230/Uranium-234 Isotope ($^{230}\text{Th}/^{234}\text{U}$) Dating

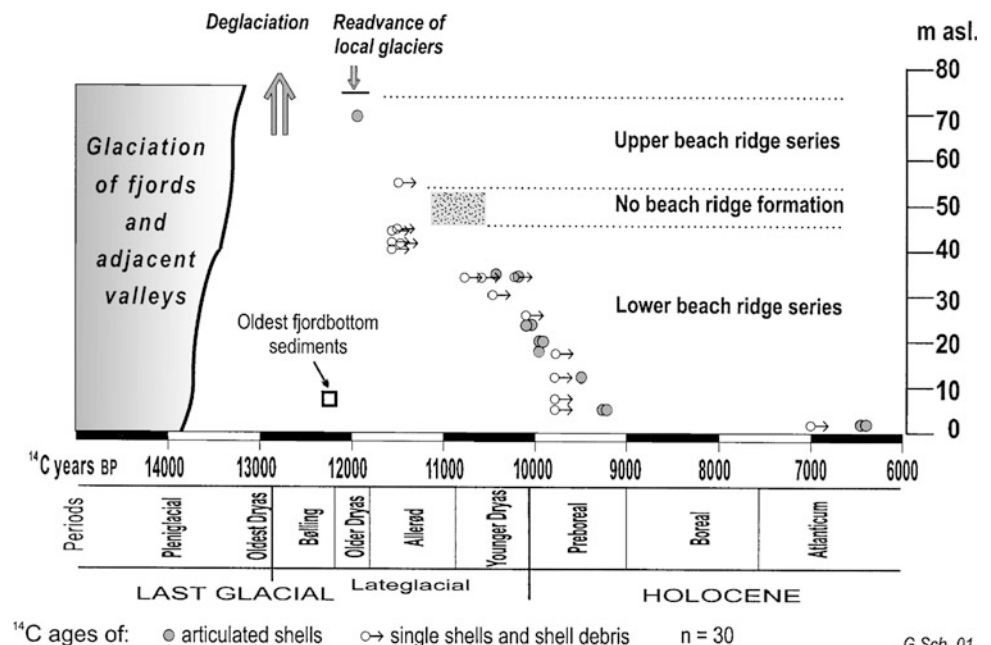
The thorium-230/uranium-234 ($^{230}\text{Th}/^{234}\text{U}$) dating method is the most frequently used uranium isotope age determination technique. In coastal research, this method is utilized for dating marine and lacustrine carbonates (Edwards et al. 2003). It is used, first of all, for dating Middle and Late Quaternary corals (Zhao and Collins 2011), sometimes for dating submerged coastal speleothems (Stirling 2015), and to a lesser extent for dating mollusk shells.

The method is based on the fact that coral and mollusk shells during formation take up the mother nuclide ^{234}U from seawater without its daughter nuclide ^{230}Th . Modern corals contain uranium with an initial $^{234}\text{U}/^{238}\text{U}$ ratio of 1.15 ± 0.03 (Edwards et al. 2003), and after the organism's death, ^{234}U decays to ^{230}Th . The ^{230}Th half-life of 75.4 ka determines the upper dating limit of this method, which is reached at 500 to 600 ka (approximately seven times the half-life of ^{230}Th). However, reliable data have been provided mainly for the past 130 to 200 ka.

The uranium-234 and thorium-230 isotope content can be determined by α -counting (alpha-spectrometry), by thermal ionization mass spectrometry (TIMS), as well as by inductively coupled plasma mass spectrometry (ICP-MS) or by laser-ablation multi-collector ICP-MS (Spooner et al. 2016). The mass spectrometric methods offer a higher precision (the analytical error is between 0.5 and 7%), require smaller sample sizes (0.2 to 0.05 g), and need shorter measurement times (MC-ICP-MS only some minutes).

Geochronology, Fig. 1

Deglaciation and evolution of beach ridges in Northern Spitsbergen (Woodfjord, Wijdefjord) based on ^{14}C dating. Beach ridges (in m above sea level) are represented by circles. All ages are uncalibrated ^{14}C years BP. The average reservoir effect for marine carbonates (approximately 400 years) needs to be considered for the correlation of the ^{14}C ages to geologic time periods (e.g., Younger Dryas). Articulated shells give reliable ^{14}C ages for beach ridge formation. The ^{14}C ages of coastal features based on single shells and shell debris are frequently overestimated since this material may have been relocated (modified after Brückner and Schellmann 2003)

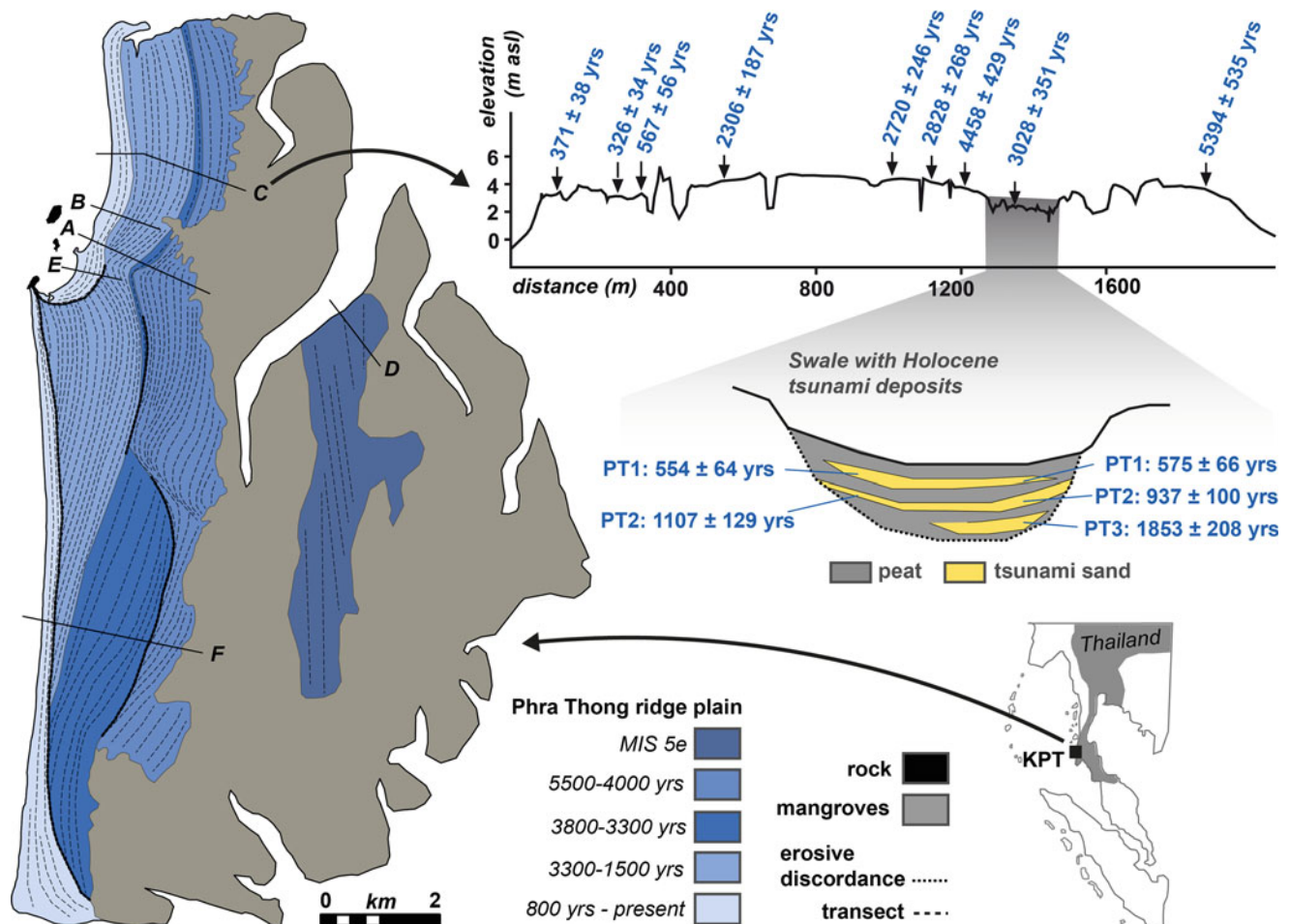


Although the precision of Th/U dating has significantly improved due to advances in analytical techniques, numerous sources of error still limit the accuracy of the approach. One key assumption in Th/U dating is that neither the mother nor the daughter nuclides are relocated after uranium accumulation in the coral structure or mollusk shell (a closed system). However, uranium is easily soluble in an oxidation environment and may be added or removed. The resulting error in age determination can amount to tens of thousands of years and more. Partially recrystallized coral samples are neither suited for Th/U dating, since uranium and thorium may be mobilized in the process of calcite formation through the recrystallization of the aragonitic coral structure. Th/U and ESR dating results of paired coral samples from the MIS 5a-1 coral reef near the south point of Barbados (Fig. 2) illustrate very well that the apparently minor analytical errors (precision) of TIMS Th/U ages (less than 1%, ESR between 5% and 8%)

do not necessarily guarantee a better temporal resolution (Schellmann et al. 2004).

Mollusk shells are more prone to uranium and thorium migration than coral. Therefore, they are less suited for Th/U dating (Kaufman et al. 1971). While corals absorb approximately 3 ppm uranium from seawater, mollusk shells take up most of their uranium (approximately 0.2–11 ppm) after death from ground and seepage water. The majority of this uranium is taken up within the first few centuries and millennia after their death (Schellmann et al. 2008). However, later isotope exchange with the surrounding environment is possible. The variability of the $^{234}\text{U}/^{238}\text{U}$ ratio in seawater can also cause errors in the Th/U dating of Pleistocene coral and mollusk shells (Bard et al. 1992).

Despite all potential sources of error, mass spectrometric Th/U dating of corals was utilized for the extension of the ^{14}C calibration curve to up to 50 ka BP (Reimer et al.



Geochronology, Fig. 2 Evolution of the beach ridge plain on Phra Thong Island (KPT), SW Thailand, on the basis of quartz OSL data. While the eastern part of the island formed during MIS 5e, the western ridge plain section was accumulated – with varying progradation velocities – during the last ~6000 years. Quartz OSL ages of tsunami-laid

sands archived in peat of the swales indicate predecessors of the 2004 Indian Ocean Tsunami that inundated the island 500–600 years ago (PT1), 900–1100 years ago (PT2), and ~1800 years ago (PT3). Details in Brill et al. (2015, 2012)

2013). The current knowledge of Last Interglacial sea-level changes (75 ka to 130 ka BP) is largely based on α - and mass spectrometric Th/U dating of fossil coral. Even the Th/U dating results of Penultimate Interglacial coral reefs (around 200 ka BP) can reach considerable accuracy, while the Th/U dating of Middle Pleistocene and older coral samples is not reliable due to their increased weathering and the still unknown $^{234}\text{U}/^{238}\text{U}$ ratio in seawater during those times. To summarize, Th/U dating of fossil coral produces accurate results for up to the last 200 ka BP, while the Th/U dating of mollusk shells is less reliable.

Optically Stimulated Luminescence (OSL) Dating

Luminescence dating techniques are typically used to determine the time period that has elapsed since sediment grains have last been exposed to daylight during transportation. The fundamental concept of this method (Preusser et al. 2008; Aitken 1998) is the use of naturally occurring semiconducting crystalline minerals, such as quartz and feldspar, as natural dosimeters. These minerals record the time span they have been exposed to ionizing radiation generated by the decay of uranium, thorium, potassium, and rubidium in the surrounding sediment, as well as by the influence of cosmic rays while their sediments were morphodynamically stable. Ionizing radiation transfers charge within the crystal lattice by moving electrons from their stable state to so-called electron traps. In the traps, electrons are stable until released by exposure to heat or sunlight, whereas the number of trapped electrons increases with the time and dose of natural radiation. The released electrons recombine with so-called luminescence centers, a process accompanied by the emission of photons. This light emission is the luminescence signal that is used for dating. For coastal sediments, such as foredunes or littoral sands, the resetting event (called “bleaching”) is the exposure of the sand grains to daylight during transport by wind or waves.

To determine the luminescence signal accumulated in a mineral, different modes of laboratory stimulation can be used. While luminescence emissions evoked by light (optically stimulated luminescence, OSL) are sensitive enough to sunlight resetting in nature for being applicable to a large variety of coastal sediments, luminescence signals stimulated by heat (thermoluminescence, TL) are less light sensitive and, therefore, lately not used anymore for dating coastal sediments. The most commonly applied OSL techniques are infrared stimulated luminescence (IRSL) for feldspar and blue light stimulated luminescence (BSL) for quartz.

Luminescence techniques measure the radiation dose (equivalent dose, D_E) that has accumulated in the minerals

since the last resetting of charge, either using the additive dose method or the regeneration method. These can be conducted on many aliquots of a sample (multiple aliquot approach), on one aliquot of a sample (single aliquot approach), or on one grain of a sample (single grain approach). However, since the late 1990s, in the large majority of all studies, the single aliquot regenerative dose (SAR) approach of Murray and Wintle (2000) is applied. While quartz is usually the preferred dosimeter for Holocene deposits, feldspar may be advantageous for older sediments due to a more extended dating range. However, feldspars are often affected by signal loss over time that is not related to heat or light exposure (anomalous fading). Alternative approaches for dating Middle to Late Pleistocene deposits have come up during the last 10 years in the form of post-infrared infrared (pIRIR) stimulated luminescence dating (Thomsen et al. 2008) or thermally transferred OSL (TT OSL; Wang et al. 2006).

In addition to dose determination, measurements of the environmental ionizing radiation rate (dose rate or annual dose, D') are required for the age calculation of a sample. D' is estimated in the laboratory using, e.g., gamma spectrometry or directly in the field using a portable gamma spectrometer. The sample age is then calculated by dividing D_E (in Gy) by the annual dose D' (in Gy/ka). Luminescence methods have been successfully used for generating deposition chronologies ranging from a few years to at least 100–150 ka using quartz and up to at least 300 ka using feldspar. The precision reached in luminescence dating is typically 5–15%.

Luminescence methods are applied to a wide range of depositional environments. While aeolian sediments, including coastal dune sands, are best suited for luminescence dating due to prolonged exposure to light during wind transport, water-laid coastal sediments may be suited as well (Lamothe 2016; Jacobs 2008) depending on water depth and the concentration of suspended sediment. Complete bleaching of coastal sediments prior to their deposition is particularly likely in shallow marine environments that are located in regions with intense insolation. Sand- to silt-sized sediments have been dated from a large range of coastal settings, including littoral deposits (Brill et al. 2015; Reimann et al. 2010), intertidal deposits (Zander et al. 2007; Madsen et al. 2005), tsunami deposits (Brill et al. 2012; Huntley and Clague 1996), and storm deposits (Brill et al. 2017; May et al. 2015). Lately, also cobble-sized beach sediments have been dated using OSL (Simms et al. 2011). While signal resetting is, thus, usually sufficient to allow OSL dating of coastal sediments, significant inaccuracies may be introduced when radioactive disequilibria (e.g., due to uranium uptake or excess in coastal carbonates) complicate dose rate determination (Zander et al. 2007).

Case Study: OSL Dating of Beach Ridge Formation and Tsunami Inundation in Thailand

The beach ridge plain on Phra Thong Island, SW Thailand, provides a good example for the applicability of OSL dating in coastal settings (Fig. 2). Quartz OSL ages of the beach ridges reveal that island formation started already during the Last Interglacial sea-level highstand (MIS 5e) and continued – with varying velocities – during the last 6000 years following the mid-Holocene sea-level maximum (Brill et al. 2015). At the same time, quartz OSL dating of tsunami-laid sand sheets archived in the peat of the swales indicates ages of 490–550, 925–1035, and 1740–2000 years for the last three predecessors of the 2004 Indian Ocean tsunami (Brill et al. 2012).

Electron Spin Resonance (ESR) Dating

The first systematic studies and applications of electron spin resonance (ESR) dating of corals and mollusk shells were conducted in the 1980s. Since then, a large number of Pleistocene coastal features and sediments have been dated using ESR. Like luminescence dating, ESR dating is a radiation-induced dosimetric dating approach. It is based on the radiation damage in minerals that increases with age and is caused by ionizing radiation from natural radioactivity and cosmic rays. These damages generate paramagnetic centers and radicals that can be measured using ESR (Schellmann and Radtke 2015).

An ESR age derives from the ratio of the accumulated radiation dose D_E to the dose rate D' as already described for OSL. However, unlike OSL, the D_E in ESR dating usually is still determined by an additive dose method, for which 20 or more aliquots of the same sample are stepwise irradiated with increasing γ -doses. Trapped electrons increase in number with increasing gamma doses and, thus, cause the amplitude of the ESR dating signal to increase. The D_E is determined by extrapolating the resulting dose response curve to zero ESR intensity. D' can be determined by the same approaches as used for estimating the dosimetry of OSL samples (in the laboratory or in the field). The correct determination of D' and D_E is essential for any ESR age estimate.

Schellmann et al. (2008) and Blackwell et al. (2016) describe ESR dating of Quaternary materials and discuss their accuracy and precision. In the last years, both precision and accuracy of ESR dating of fossil corals and mollusk shells have improved due to the higher resolution of ESR measurements and due to new techniques for D_E determination (Schellmann and Radtke 1999, 2001). However, ESR ages can deviate by several thousand years, if the natural radiation dose of the sample was affected by uranium or thorium mobilization or by changing moisture conditions in the past. Further problems may arise from the heterogeneity of the

surrounding sediments and from the open system behavior of shells with respect to uranium migration.

ESR dating results of mollusk shells that were embedded in heterogeneous beach sediments are most affected by these sources of error. Dating uncertainties can reach more than 15%. Therefore, ESR dating results for mollusk shells from littoral deposits are frequently limited to the mere differentiation between Holocene, Last Interglacial, Penultimate Interglacial, and older sediments (Schellmann and Radtke 2000).

On the other hand, dating of coral originating from relatively homogeneous reef limestone can generally be accomplished with errors of 5 to 8%. Substages of sea-level highstands (e.g., Last Interglacial Oxygen Isotope Substages 5a, 5c, and 5e) can be discriminated due to improvements in the determination of the D_E . ESR ages of corals from the Late Pleistocene are now as accurate as those based on Th/U dating (Fig. 3). Therefore, the application of both methods simultaneously is recommended for the dating of coral up to 200 ka BP.

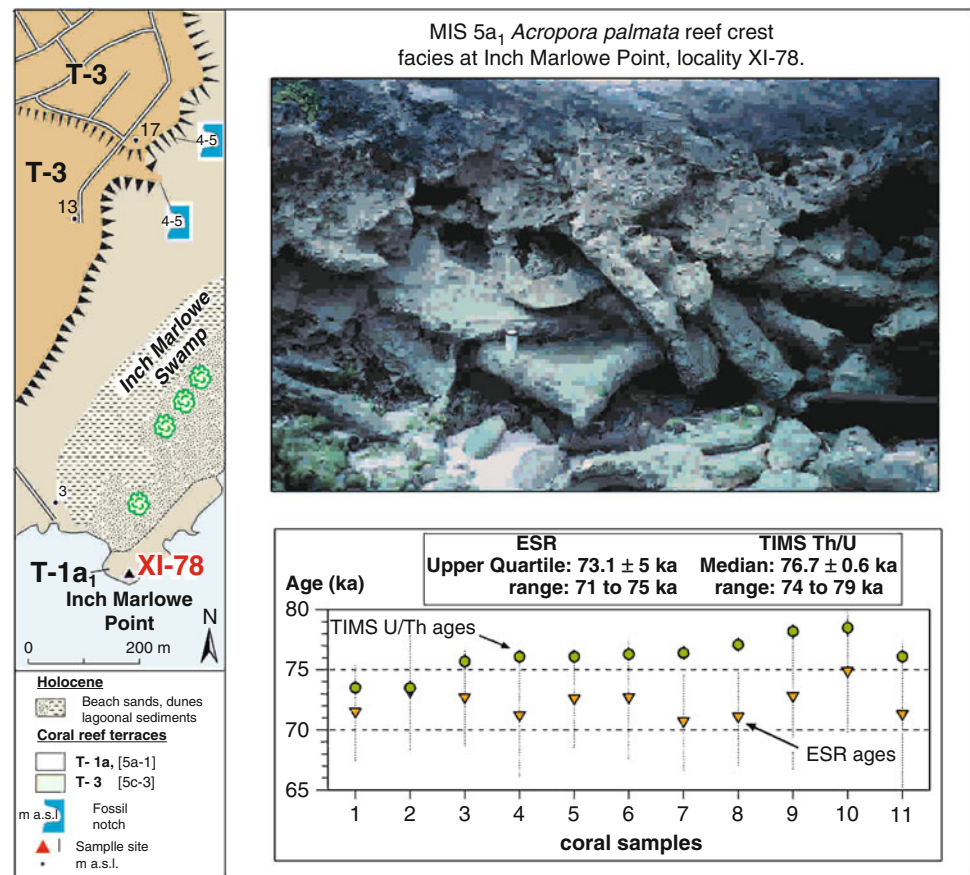
The ESR method allows for the dating of corals and mollusk shells for a larger time span than any other Quaternary dating method. The lower dating limit lies at a few thousand years BP, and the ESR method should theoretically allow for the dating of corals and mollusk shells that are a few million years old. However, numerous ESR dating results for mollusk shells and coral illustrate that the upper dating limit lies at 200 to 300 ka and 400 to 600 ka, respectively. This limit is largely dependent on changing conditions due to diagenesis and weathering. For example, the recrystallization of the aragonite coral structure or mollusk shell can cause considerable underestimates of ages, since it eliminates previously stored radiation.

ESR has been widely applied to sequences of raised coral reef terraces with ages of up to 500–600 ka, e.g., terraces on Sumba island (Pirazzoli et al. 1991) and on Barbados (Radtke and Schellmann 2006; Schellmann and Radtke 2004a). Mollusk shells and gastropods with ages ranging up to 200 to 300 ka were also dated using ESR, e.g., mollusk shells at the Patagonian Atlantic coast (Schellmann and Radtke 2000) or Late Pleistocene gastropods in eolianites on Cyprus (Schellmann and Kelletat 2001).

Case Study: ESR Dating of Pleistocene Coral Reef Terraces in Southern Barbados

ESR dating of the fossil coral reef terraces on Barbados delivers an impressive example for the application of this method (Fig. 4). In southern Barbados, Late and Middle Pleistocene coral reef terraces rise up to 63 m above present sea level. The ESR and similarly the TIMS $^{230}\text{Th}/^{234}\text{U}$ ages increase with height up to 224 ka (MIS 7). The age-height relationships allow for the calculation of neotectonic uplift rates in Barbados and of paleosea-level changes (Schellmann and Radtke 2004a; Schellmann et al. 2004; Potter et al. 2004).

Geochronology, Fig. 3 ESR and TIMS Th/U ages ($\delta^{234}\text{U} > 141$ and $< 157\text{‰}$) of MIS 5a-1 coral reef terrace near Inch Marlowe Point, southern coast of Barbados (modified after Schellmann and Radtke 2004a and Schellmann et al. 2004)



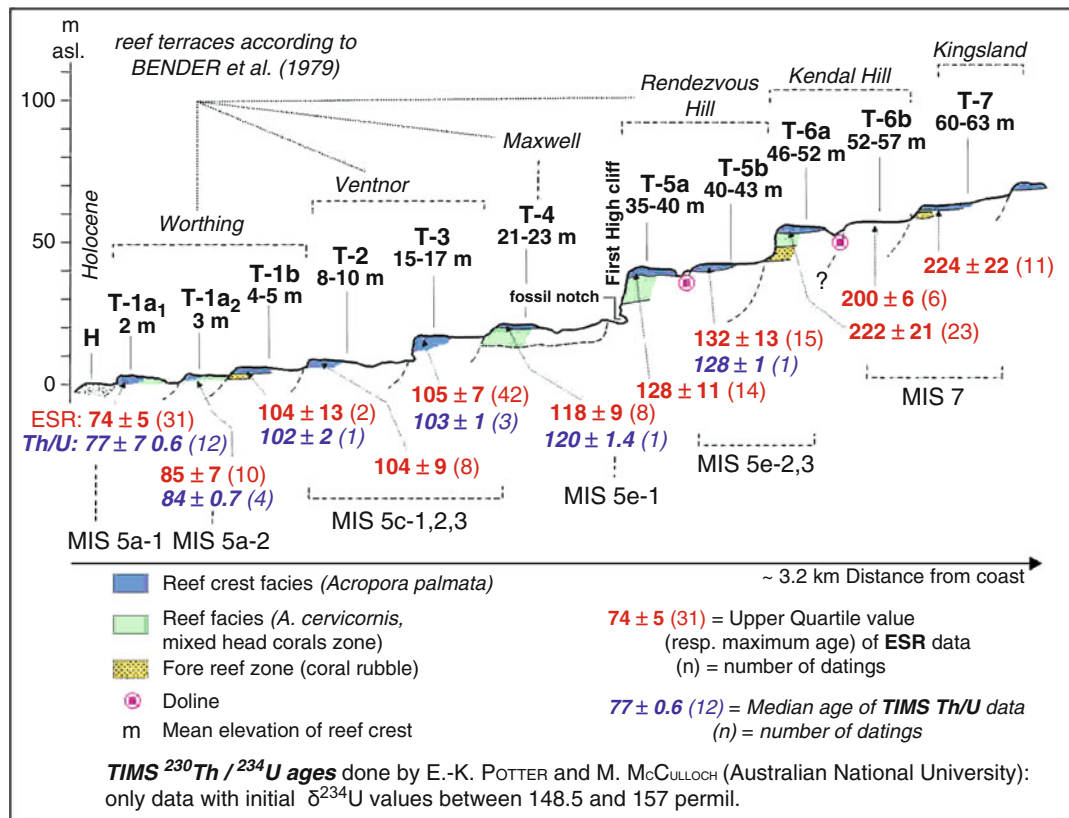
Terrestrial Cosmogenic Nuclide (TCN) Dating

Dating techniques based on the measurement of terrestrial cosmogenic nuclides (TCN) determine either exposure ages or burial ages for rock surfaces and sediments over time scales of thousands to millions of years (Dunai 2010; Gosse and Phillips 2001). TCN are produced in the uppermost rock formations where these are exposed to cosmogenic radiation. At the Earth's surface, cosmic radiation is dominated by the secondary cosmic ray flux that results from interactions of primary cosmic radiation (i.e., mostly protons) with the atmosphere. These secondary particles, mainly high-energy nucleons and muons, interact with the mineral atoms through different mechanisms (i.e., spallation, thermal neutron capture, and muon capture) to generate new *in situ* nuclides in the exposed rocks or sediments. The TCN production rates vary with geographical site characteristics such as latitude, altitude, and local shielding (e.g., surrounding topography) and decrease exponentially with depth below surface. While nucleons dominate production rates in the uppermost 2–3 m, their contribution becomes insignificant below this depth where production is essentially controlled by muons.

TCN that (i) are stable or long-lived radioactive (i.e., half-life > dating range), (ii) have low background levels in the

rocks that shall be dated, (iii) are not significantly affected by meteoric input, (iv) can be produced in common rock types, and (v) are well understood in terms of their production pathways have been increasingly used for dating applications in Earth sciences since the development of AMS facilities in the late 1980s which allows their precise measurement (Dunai 2010). Until now, mainly the stable noble gases ^3He and ^{21}Ne , as well as the radionuclides ^{10}Be , ^{26}Al , and ^{36}Cl , are routinely applied. While the stable TCN can provide exposure ages for pre-Quaternary surfaces, ^{36}Cl , ^{26}Al , and ^{10}Be , with half-lives of ~ 0.3 Ma, ~ 0.7 Ma, and ~ 1.4 Ma, respectively, allow to establish exposure and burial histories that theoretically cover the entire Quaternary. However, inheritance (i.e., nuclide accumulation prior the dated event), large background levels, and erosion of unstable surfaces may constitute significant challenges for accurate dating.

In coastal contexts, TCN applications have focused on dating the exposure time of quartz cobbles in depositional marine terraces (e.g., Binnie et al. 2016; Perg et al. 2001), or the exposure time of abrasive platforms (Regard et al. 2012), using either ^{10}Be or a combination of ^{10}Be and ^{26}Al . Lately, Ramalho et al. (2015) and Rixhon et al. (2017) used ^3He and ^{36}Cl to determine exposure ages for wave-emplaced basaltic and coral boulders, respectively.



Geochronology, Fig. 4 ESR and TIMS Th/U ages of uplifted Middle and Late Pleistocene coral reef terraces in a transect on southern Barbados (slightly modified after Schellmann and Radtke, 2015). Details in Schellmann and Radtke (2004a, b)

Amino Acid Racemization (AAR) Dating

The amino acid racemization (AAR) dating technique is a relative age determination method for various organisms like marine mollusks, coral, or foraminifera (e.g., Demarchi and Collins 2015; Wehmiller 2013). AAR is applied to time scales ranging from hundreds to hundreds of thousands of years. The method is based on the rate of racemization and epimerization of indigenous amino acids preserved in mollusk shells or corals from coastal deposits. This protein degradation processes are crucially influenced by the temperature history during diagenesis and the rate of hydrolysis of residual proteins.

Several uncertainties in the applicability and reliability of AAR remain, since (a) racemization rates may vary between genera and (b) diagenesis temperature histories may vary between different geographic settings. The AAR method seems to be reliable for local and regional aminostratigraphic units only. Since AAR is a relative dating technique, its results need to be calibrated using numeric dating methods like radiocarbon or U-series dating.

Paleomagnetic Dating

Paleomagnetic dating determines the natural remanent magnetization of sediments and correlates it to the known paleomagnetic polarity stratigraphy. Permanent magnetism is carried by ferromagnetic minerals that are minor constituents of almost all rocks. Magnetic minerals align to the Earth's magnetic field and keep that orientation when they are deposited or precipitated in the sedimentary matrix during early diagenesis. Garces (2015), Calvo-Rathert (2015), and Hambach et al. (2008) describe the physical properties and some applications of this method.

In the Quaternary several instabilities of the Earth's magnetic field occurred in terms of variations in field intensities and field directions, forming geomagnetic reversals and excursions (Singer 2014; Hambach et al. 2008). They can be used for relative dating of sedimentary and volcanic rocks. The paleomagnetic field reversals at the base of the so-called normal Brunhes and reverse Matuyama chrons at 0.78 Ma and 2.6 Ma, respectively, are the most important polarity changes of the Earth's magnetic field in recent geological history. The Jaramillo normal (c. 0.9 to 1.1 Ma) and Olduvai normal (c. 1.7

to 1.9 Ma) subchrons are two important reversals within the Matuyama chron. At least 27 short-lived field excursions occur within the Brunhes (14) and Matuyama (13) chrons, even though their global evidence is still under research. The most prominent short-lived excursion is the “Blake Event” (approximately 120 ka), which seems to be a worldwide reversed interval (Singer 2014).

Sample applications of the paleomagnetic method include the dating of marine terraces in southern Italy (Brückner 1980) and the dating of a sequence of beach ridges and coastal dunes at the Moroccan Atlantic coast (Brückner and Hambach, unpublished data). The paleomagnetic method requires a continuous sequence of coastal formations and a narrow sampling strategy. Paleomagnetic dating allows to differentiate within the Middle and Older Quaternary, where other dating methods fail. However, lately paleomagnetic dating has also been applied to tentatively determine the relocation of wave-emplaced coastal boulders during the Holocene (Sato et al. 2014).

Summary

In the context of dating Quaternary coastal structures, sediments, and processes, new dating methods were established, and the precision of traditional methods was considerably increased. Besides age determinations by relative methods like the amino acid racemization (AAR) and the paleomagnetic method, the use of numerical dating techniques has significantly increased. Fossils like corals, mollusk shells, or foraminifera can be dated by radiocarbon (^{14}C), thorium-230/uranium-234 ($^{230}\text{Th}/^{234}\text{U}$), or electron spin resonance (ESR) and aeolian or beach sediments by optically stimulated luminescence (OSL). Each dating technique has its own time range, precision, and accuracy, which has to be taken into account.

Cross-References

- Coral Reefs
- Coral Reef Coasts
- Coral Reefs, Emerged
- Isostasy
- Uplift Coasts

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Geodesy

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According to the traditional classical definition, attributed to F. R. Helmert in 1880, geodesy is *the science of the measurement and mapping of the earth's surface*. Nowadays, this definition is deemed to encompass determination of the topography of the earth's landmasses, the bathymetry of the seafloor, the shape of the surface of the ocean, and form and nature of the earth's internal and external gravity field.

History

The origins of modern geodesy can be traced to antiquity. The ancient Greeks speculated on the shape of the earth with Pythagoras, and later Aristotle, introducing the concept of a spherical earth. The first scholar to make a scientific estimate of the shape of the earth was Eratosthenes of Alexandria (276–195 BC) who obtained a value for the circumference of earth to within 2% of modern estimates. A later estimate by Posidonius (135–51 BC) was used by Ptolemy (AD 90–168) to produce the monumental eight volume *Guide to Geography*, which contained maps of the world destined to be the definitive work for the next fifteenth centuries. Columbus, based on such maps, was to underestimate the distances from Europe to India and Asia and, in turn, discover the Americas. Advances in astronomy, instrumentation and physics in the sixteenth and seventeenth centuries led to Isaac Newton (1643–1727) and Christian Huygens (1629–95) proposing that the earth was not spherical, but flattened at the poles. This proposal led to some controversy and, it was not until the late eighteenth century, following increasingly accurate measurements on the earth's surface, that an ellipsoidal model for the shape of the earth was universally adopted.

Essential Geodetic Concepts

Definition of the size and shape of the earth does not usually involve surface topography or the bathymetry of the oceans, but the surface of mean sea level and its theoretical continuation under the continents. Topographic heights and bathymetry are measured relative to this hypothetical surface which is known as the geoid. By definition, the geoid represents an equipotential surface and is therefore a direct representation of the earth's gravity field. The geoid is a mathematically complex surface, its shape being strongly influenced by irregular mass distributions below the earth's surface. It is thus unsuitable as a reference surface for a geometric figure of the earth. For further details on the earth's gravity field and the geoid, the reader is referred to Heiskanen and Moritz (1967).

For geodetic calculations, such as computing geodetic latitude and longitude and defining map projections (e.g., Bugayevskiy and Snyder 1995), an ellipsoid of revolution (a three-dimensional surface generated by rotating an ellipse 360° about its minor axis) is defined. This simple figure is known as a reference ellipsoid and is defined by a semimajor axis (equatorial radius) and semiminor axis (polar radius). The earth's semimajor axis is about 21 km greater than the semiminor axis, with the flattening being approximately 1/300. A commonly used reference ellipsoid is that associated with the World Geodetic System 1984 (WGS84) (NIMA 1997).

For geodetic measurements, the concept of geodetic datum is introduced. A geodetic datum is a three-dimensional representation of the surface of the earth through a framework of discrete points whose coordinates and elevation (either above mean sea level or a reference ellipsoid) are accurately known. In individual regions, datum coordinates and computations are based on a local reference ellipsoid, which corresponds closely to the shape of the geoid in that area. Modern satellite techniques, such as the Global Positioning System (GPS) (*q.v.*), rely on geocentric reference ellipsoids which have the center of the earth as the origin. To ensure compatibility between GPS coordinates and coordinates in the framework of the local geodetic datum, a mathematical function defining the transformation of the coordinates between the two systems must often be computed. A geodetic datum forms the basic structure for densification surveys which, in turn, can be incorporated into maps or Geographic Information System (GIS) (*q.v.*) databases. The concepts of geodetic datums and reference systems are covered in detail by Torge (1991).

Applications of Geodesy

Space geodetic measurement techniques (descriptions of very long baseline interferometry (VLBI), satellite laser

ranging (SLR), and GPS are given in Seeber (1993)) are used to define international geodetic datums, such as the International Terrestrial Reference Frame (ITRF), which provide the basic infrastructure for positioning and navigation on the surface of the earth. These reference frames are used for global and regional plate tectonic studies, measurement of isostatic rebound, and connection of *tide gauges* (*q.v.*) for monitoring the levels of the world's oceans (see also “► [Tidal Datums](#)” (*q.v.*)). They are also critical for satellite orbit tracking required for such applications as satellite *altimetry* (*q.v.*) and *SAR* (*q.v.*) studies. Global geodetic measurements are also used to derive parameters that can describe variations in the rotation of the earth (e.g., length of day, polar precession, and nutation), and the dynamics of the atmosphere, oceans, and solid earth (Lambeck 1988).

Local geodetic measurements rely on survey instrumentation such as electronic distance meters (EDM), precise levels and *GPS* (*q.v.*) receivers. Such measurements are primarily used for mapping purposes, providing local control for *photogrammetry* (*q.v.*) and *remote sensing* (*q.v.*), and for engineering applications such as monitoring physical changes in man-made structures and ground surfaces. For a complete appraisal of geodesy and terrestrial surveying, the classic text is Bomford (1980).

Cross-References

- [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
- [Geographic Information Systems \(GIS\)](#)
- [Global Positioning Systems \(GPS\)](#)
- [Photogrammetry](#)
- [Remote Sensing: Wetlands Classification](#)
- [Tidal Datums](#)
- [Tide gauges](#)

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Geographic Information Systems (GIS)

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Definition

Geographic Information Systems (GIS) are computerized spatial decision-making tools that work with place-based information and allow users to display, manage, and analyze geographic data. Since coastal features typically show distinct spatial patterning, coastal managers and scientists have incorporated this technology from its beginnings. However, coastal data often represent dynamic, multidimensional systems that present problems to this historically land-based technology.

History

The concept of a GIS originated from noncomputerized decision-making techniques. One classic example is a decision-making technique developed by Ian McHarg, a landscape architect and author of *Design with Nature*. McHarg's method for identifying suitable locations for development was based on defining the most favorable characteristics of various sets of information such as slope, soil type, geology, land use, and accessibility and then combining the layers into a composite map (McHarg 1969). Although McHarg's technique may be carried out in an analog fashion, it is a popular example of the power of applying digital computer technology.

The beginning of the modern, digital, GIS began in the late 1960s. Several efforts in developing digital spatial systems began concurrently and involved applications for urban planning, transportation, natural resources, and the US Census. The first use of the term GIS was the Canada Geographic Information System, implemented in 1964 to identify marginal agricultural lands for rehabilitation (Foresman 1998; Wright and Bartlett 2000).

Data Models

A data model represents the spatial entity and its relationship to the real world. There are two primary data models: raster models and vector models. Raster models generally depict continuous data by representing data points in a regular pattern of cells. The size of the cells, for example, 10 m by 10 m, defines the geometric resolution and a series of row and column coordinates define the spatial extent. Continuous data commonly depicted with a raster model include surface

elevation, precipitation, and satellite image data such as land cover.

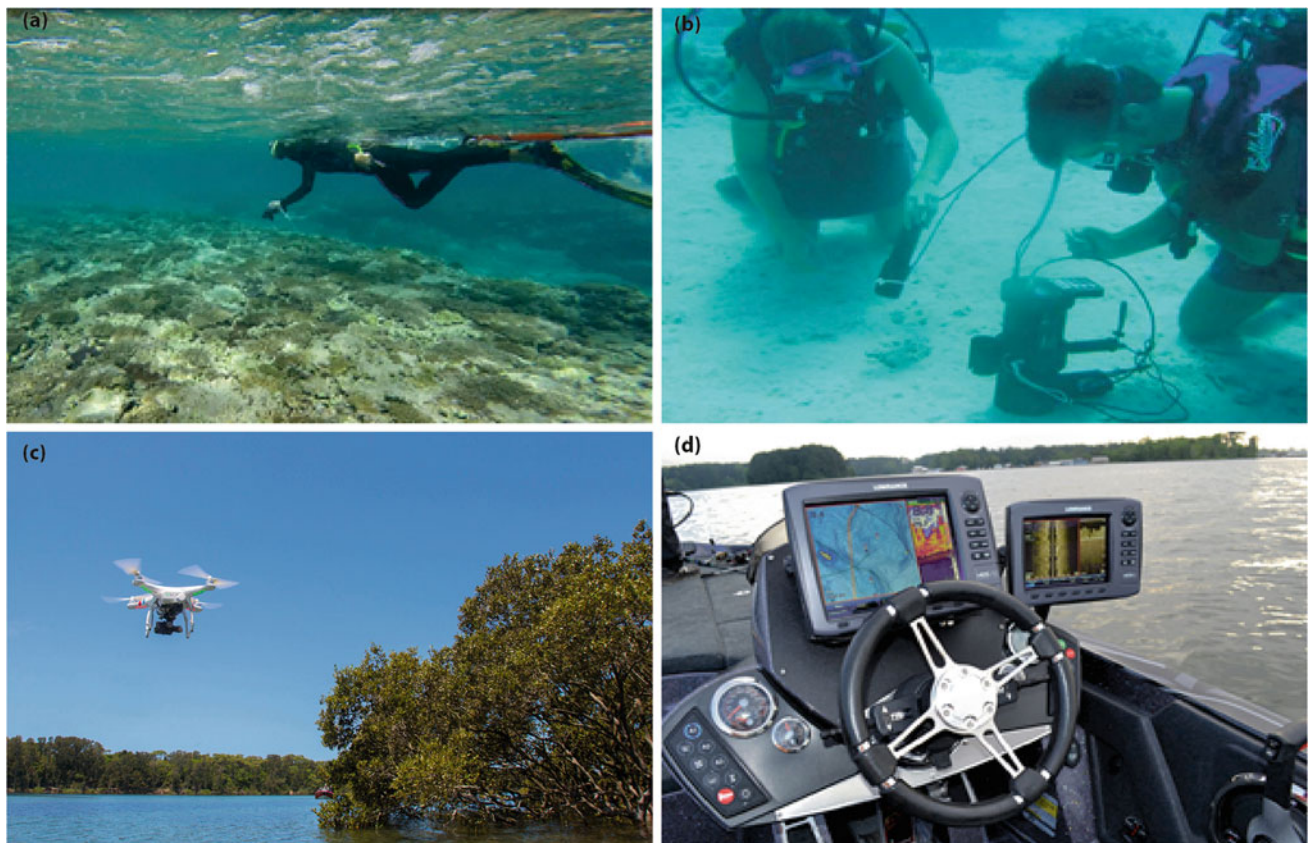
Vector models are topological, representing discrete data with nodes and lines that can be represented as points, lines, or polygons (Bernhardsen 1992). The node may define a starting point, an intersection, an end point, or any point along a line. The topological model retains the relationship and connectivity between and among points, lines, and polygons. For example, a series of nodes and lines may define two adjacent counties; the shared border is captured and stored only once in a vector model and that relationship remains intact even when the geometry is changed.

Functionality

A GIS typically uses hardware and software that allows the following functionality:

- Data input/capture
- Data storage/management
- Data manipulation
- Data analysis
- Data output and display

If data are not already in a digital form, data capture may be a time-consuming component of building a GIS. Coastal scientists use a combination of primary and secondary data. Examples of primary data capture techniques include hand-digitizing maps, electronic scanning that converts points and lines, Global Positioning Systems (GPS) that digitally record surveyed locations, manual entry of geographic coordinates, and the use of instruments with in-built GPS technology that add geographic referencing to information collected, such as bathymetric survey points from sonar, above-water aerial images from drones, and below-water images from autonomous underwater vehicles (AUVs) (see Fig. 1). Attribute characteristics are assigned once geographic entities are digitally captured. For example, a river (the entity) may be hand digitized and its name would be stored separately as an attribute. At minimum every attribute will have a unique identification code and there may be any number of additional



Geographic Information Systems (GIS), Fig. 1 Collecting primary spatial data in the coastal zone: (a) A snorkeler towing a GPS along in a floating waterproof bag while taking ground referencing photographs above a coral reef, (b) scuba divers collecting spectral signatures of the seafloor to support broader scale mapping using remote sensing

techniques, (c) a drone collecting imagery from a GoPro above mangroves in Minnamurra River, New South Wales (Photo credit: Mark Newsham). (d) A coupled GPS monitor and side-scan sonar instrument (Photo credit: Frank Sargeant)

attributes allocated to each entity. Secondary data, both base data and thematic data, may be acquired from an ever-growing multitude of sources. Spatial base data, such as roads, rivers, and administrative boundaries, are often easily downloaded from government sources and other public sharing domains. In using secondary data, one must ensure data quality such as completeness, accuracy, currency, and lineage. Acquiring secondary data may also require a conversion process from one proprietary format to another or conversion from one coordinate system or map projection to another.

Data storage and management is conducted using the database management system (DBMS) component of a GIS. Users create, update, and store the geographic entity files and their associated attributes in a DBMS. Metadata, information about the history and extent of the spatial data, should similarly be stored and managed. Once collected and stored in a GIS database, spatial data may be manipulated (queried, classified, and edited) using tools inherent in GIS software packages.

The ability to analyze spatial data is the component that differentiates GIS from desktop or web-based mapping software. GIS software provides tools to analyze distance, direction, connectivity, and proximity through the use of spatial and attribute queries. Furthermore, data modeling tools allow the depiction of two- and three-dimensional characteristics of the earth's surface, subsurface, and atmosphere. The movement of humans or natural features through linear networks such as roads and streams may be modeled. Map overlay analysis involves combining two or more layers of information through logical operations. McHarg's suitability analysis is a perfect example of the use of map overlay analysis.

Finally, data output and display capabilities allow GIS users to prepare graphic images of their results. The images may be printed as paper maps or stored as image files for electronic display or for use in web-based applications.

Coastal GIS

Since coastal data are spatial, GIS technology was recognized early in the 1970s as a useful tool for coastal science applications and for coastal zone management decision-making. The first GIS coastal applications were developed for tasks such as development or tourism planning along coasts (Bartlett 2000). Early efforts to apply GIS were thwarted by the limited availability of software, hardware, trained personnel and data models. Furthermore, the built-in functionality necessary for coastal applications that require three-dimensional and temporal analysis was not available.

The structure of the data model had also limited coastal GIS applications. While raster-based models allow simple programming and high volume storage, their spatial

resolution is compromised. Conversely, vector systems portray finer resolution but inappropriately portray boundaries that are "fuzzy." Dynamic segmentation (network models), object models, and others "hold considerable potential for advancing the utility of GIS within the coastal and marine environment" (Bartlett 2000).

Many coastal environments are controlled by physical and biological processes which can easily be upset by natural or human-induced disturbances. GIS therefore offers an opportunity to work with spatially referenced data to explore, infer, and predict coastal trends in both space and time. These often occur at broad geographical scales that enable scientists or managers to make meaningful observations and statements about the behavior of landforms and biological communities. Table 1 provides some examples of management issues across different types of coastal environment, with corresponding spatial analyses conducted to address these issues.

Unique Challenges for Spatial Analysis in Coastal Environments

Because of the presence of the sea, coastal environments present some unique challenges to the conduct of spatial analysis. These include the variable range of tidal datums particularly for referencing vertical heights (e.g., tidal mean sea level versus chart datum) and the uncertainty that this variability introduces to the definition of coastal boundaries, the rapid speed of change in shoreline features (Ellis 1978), and the practical limitations imposed on conducting fieldwork underwater (Hamylton 2017).

Tidal datums are base elevations for vertical measurements, such as under and above water elevation that are defined by phases of the tide. They are used for establishing coastal boundaries, limits of the territorial sea, water depths, critical shorelines on charts, and the limits of various coastal management responsibilities (Ellis 1978). Hirst and Todd (2003) categorize the problems associated with defining coastal boundaries into describing, visualizing, and realizing problems. Descriptive confusion arises from vague or ambiguous descriptions in legislation and policy documents, references to a datum. In some cases, coordinates are omitted entirely. To visualize or map a coastal boundary, it is necessary to represent it as a digital feature within a GIS. This raises questions of how a coastline should be defined and how accurately it can be mapped. The *Tidal Interface Working Group* of the *Intergovernmental Committee on Surveying and Mapping* have defined some commonly used tidal interface boundaries, including highest astronomical tide, mean high water springs, mean high water, mean sea level, mean low water springs, mean low water, and lowest astronomical tide (McCutcheon and McCutcheon 2003). The need to arrive

Geographic Information Systems (GIS), Table 1 Some examples of spatial analysis supported by GIS across different coastal environments

Coastal environment	Spatial analysis examples
Sandy coastlines	Simulating sea-level rise impacts, including coastal erosion and plant succession across sand dunes (Bruun 1988; Feagin et al. 2005)
	Predicting spatial and temporal variation in long shore drift sand volumes (Cooper and Pilkey 2004)
	Assessing the downstream effects of large dams in American Rivers (Graf 2006)
Rocky coasts (cliffs and platforms)	Mapping factors driving cliff failure along the Oregon coastline, including tectonic uplift, subsidence, wave attack, groundwater seepage, beach and cliff geology, El Nino events, storms, and high waters (Shih and Komar 1994)
	Monitor hard rock coastal cliff erosion from a laser scanner in North Yorkshire, UK including undercutting, detachment of blocks, debris, soils, falls, slides, and flows (Rosser et al. 2005)
Coastal wetlands	Quantifying area and width reductions to coastal barriers in a delta wetland (South Louisiana) using historical maps and airborne imagery (Williams et al. 1992)
	Tracking the removal of marshes, mangroves, and tidal flats to develop fish and shrimp farms and housing in Hong Kong (Nelson 1993)
	Modeling estuarine surface elevation dynamics (Rogers et al. 2012)
Coral Reefs	Identifying sources of sediment to coral reefs (McKergow et al. 2005)
	Modeling the route of lionfish invasion in the western Atlantic and Caribbean (Johnston and Purkis 2011)
	Modeling the risk of cyclone damage to coral reefs on the GBR from tracks and wind conditions (Puotinen 2007)
Cold coasts: Permafrost, sea ice, glaciers and fjords	Simulating rock falls from paths of steepest descent, flow paths, and changes to direction of permafrost progression (Noetzli et al. 2006)
	Simulating ice-sheet response to climate change (Hanna et al. 2013)
	Modeling riverine nitrogen delivery to Vejle Fjord (Denmark) from leaching, point source emissions, soils, terrain, and ground water head elevation and nitrate decay (Skop and Sørensen 1998)
Urbanised coasts	Bathymetric and volumetric change analysis of dredged pits in Lower Bay complex, New York Harbour (Wilber and Iocco 2003)

at an accurate and unambiguous definition of a coastline is also difficult because height datums. Often there are offsets between national reference grids (e.g., the Australian Height Datum) and datums defined by gauge measurements of tidal fluctuations apparent (Hirst and Todd 2003). Given that mapping often forms a foundation for much geospatial analysis, it is important for GIS analysts to be aware of the complexities associated with describing, visualizing, and creating digital representations of the coastline.

Spatial referencing not only refers to the longitude and latitude (i.e., x and y) measurements for a given coordinate, but it also incorporates the height above a common datum (the z coordinate). There are several tidal and land datums that could potentially serve as a reference point for the definition of height measurements. When information is collected from a survey vessel on the water surface, tidal fluctuations will influence the height of that survey vessel above the seabed. This is the case when conducting a bathymetric survey of water depths using an echosounder attached to the hull of a boat. Tidal corrections must therefore be applied to any data collected from boats to account for any offsets due to the state of the tide. This is particularly important in locations where tidal ranges are large, in some locations they can be as wide as 11 m.

Most spatial analysis involves the collection of some form of in situ information, such as community census surveys or sediment samples. On shorelines with low topographic

gradients, the difference between low and high tide can represent the submergence or exposure of substantial areas of intertidal shoreline. Moreover, the accessibility of estuarine environments depends on the ability of boats to access field sites through channels. The navigability of channels depends on the depth of water. Planning and implementing fieldwork therefore requires an accurate knowledge of how tidal fluctuations interact with local coastal topography. Because weather in coastal zones is often highly changeable, the feasibility of visiting field sites further depends on the coincidence of appropriate weather conditions with a favorable tidal window. This limits the amount of field sampling that can be done, restricting the spatio-temporal range of samples collected to seasons and geographical areas that coincide with favorable fieldwork conditions. Consequently, areas that coincide with difficult conditions (i.e., the breaker zone) are notoriously under-represented in coastal fieldwork campaigns. In some situations, a feature of interest (e.g., a kelp bed or coral reef) may require a detailed population survey. The analyst may need to employ self-contained underwater breathing apparatus (SCUBA) to achieve the necessary proximity to the field site for an adequate length of time to collect data for analysis. At a regional scale, this may result in a clustered distribution of samples, governed largely by the number of dives a fieldworker can do at a given site and how far they can swim underwater before surfacing because of an empty air cylinder.

Conclusion

Despite the practical challenges of the coastal zone, GIS remains an important tool for coastal applications. The use of GIS in coastal applications continues to grow, supported by new technologies for collecting spatially referenced data, such as drones and AUVs. Nevertheless, Bartlett's emphasis on the importance of establishing the user-base for whom the coastal zone GIS is intended, and the view of the coast and its component elements that is relevant to this constituency remains useful today. This is because this operational context will have a direct bearing on the more technologically oriented questions of hardware and software selection and the overall architecture of the system. It will also define the information products expected from the system, and hence the types of data, and the processes performed on these, that are required in order to produce the desired output (Bartlett 2000, p 18).

One measure of the importance of GIS to coastal scientists is the increasing number of user-groups, web pages, spatial data infrastructures and portals (e.g., Google Oceans and NOAA's Digital Coast), online web atlases, and conferences organized around the theme. Examples of coastal GIS conferences include the International Geographic Union's Commission on Coastal Systems, the Marine Biological and Habitat Mapping Group (GeoHab), and the Working Group on Marine Cartography of the International Cartographic Association collaborative international symposium known as CoastGIS which began in 1995 and continues to hold biennial conferences. It is clear that changes in users, data, and technology have increased the sophistication and volume of meaningful spatial analysis that can be carried out in coastal environments. It is an exciting time for harnessing the benefits of GIS technology; the challenge for managing and understanding coastal environments is to think about space and spatial processes in new and creative ways.

Cross-References

- [Beach and Nearshore Instrumentation](#)
- [Erosion: Historical Analysis and Forecasting](#)
- [Global Positioning Systems \(GPS\)](#)
- [Monitoring Coastal Geomorphology](#)
- [Nearshore Geomorphological Mapping](#)
- [Photogrammetry](#)
- [RADARSAT-2](#)
- [Remote Sensing of Coastal Environments](#)
- [Shoreline and Coastal Terrain Mapping](#)
- [Synthetic Aperture Radar Systems](#)

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Geographical Coastal Zonality

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Definition

The term “coastal zone” is used in scientific papers to describe the coastal environment or the region between the foreshore and the upper supratidal along a coastline. “Zones” in geoscience, however, are latitude-parallel belts between the equator and the poles, established in the first order by the different radiation/insolation on our globe.

Nearly all macro-patterns on Earth, which depend on climatic influences, like vegetation, soil formation, or forms of relief and inherent processes, show a zonal distribution. For the latter, including both relic and contemporary systems, this has been accepted for decades (Murphy 1968), and the term “climatic geomorphology” has therefore been established. Regarding coastal features and processes, however, these are looked upon for a long time mostly as a zonal, because cliffs, beaches, lagoons and barriers, deltas, etc. can be found in all latitudes; and rock-depending forms are regarded as a zonal, as well.

Davies (1980) established a classification of zonal coastal patterns based on prevailing wave and surf types and introduced his pioneering book on *Geographical Variation in Coastal Development*, giving an overview of the zonal coastal forms and processes, all inductively identified. Other authors then studied the global distribution of single features and their differentiation according to latitude (Kelletat et al. 2014; Martini and Wanless 2014; Perillo et al. 2009; Spalding et al. 2007; Stutz and Pilkey 2011; Wolanski et al. 2009), or the influence of humid or arid tropical climates on coastal landforms.

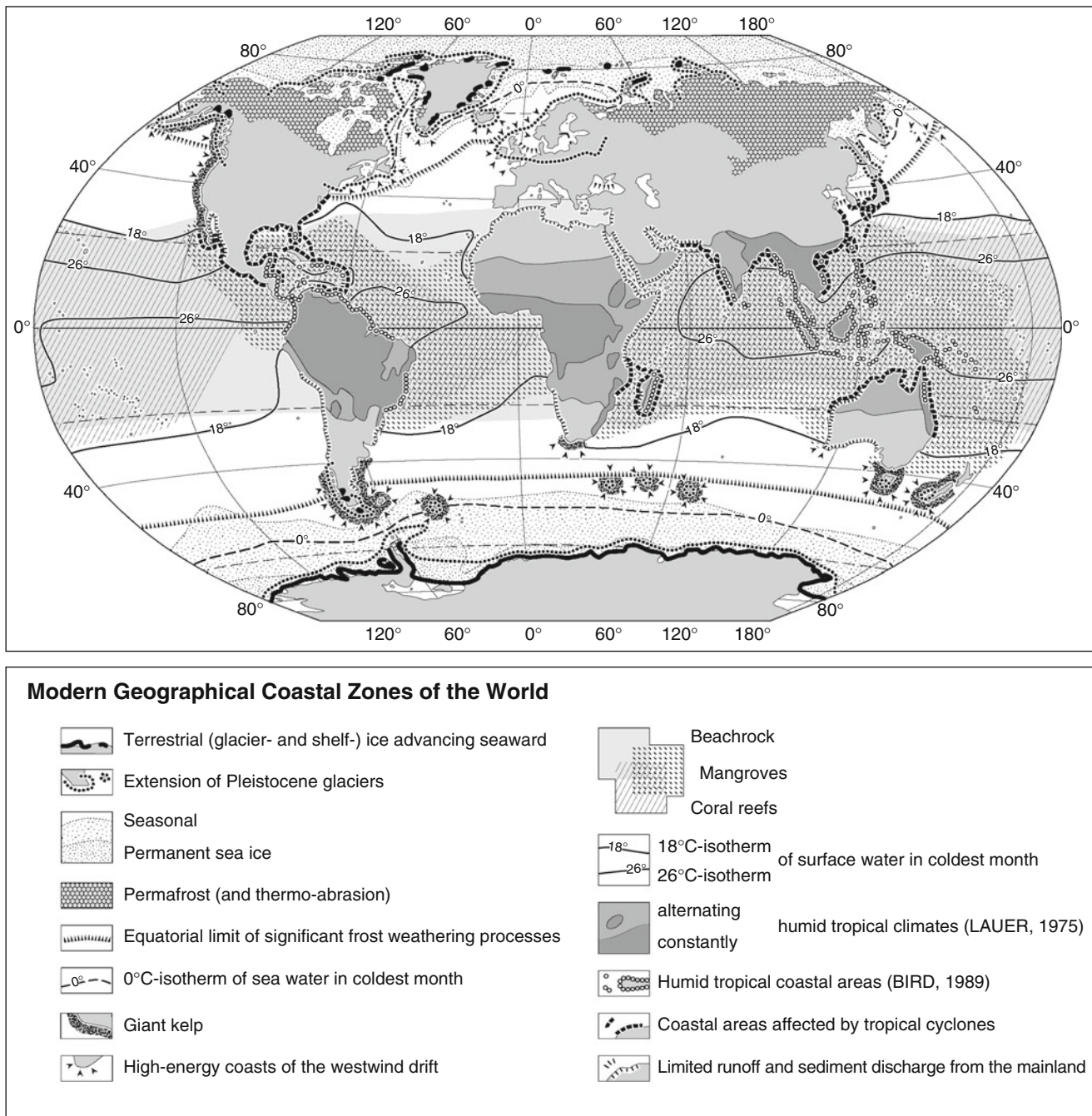
Almost two decades ago, a “system of zonal coastal geomorphology” was developed by Valentin (1979). This system is limited to the definition of 51 types of non-coastal landforms which were characteristic of the Wisconsin (Wurm)

glacial climato-morphogenic zones. Many of these areas were in part drowned during the postglacial sea-level rise. The glacio-eustatic drowning is in fact the only “littoral” factor in this system, but it is not a zonal process since it can be observed worldwide. In his “Atlas of Coastal Geomorphology and Zonality,” Kelletat (1995), however, applied the question of zonality in a strictly inductive sense, that is, interpreted from field evidence and areal distribution patterns, in contrast to previous deductive models. This is only one of the first steps, and a complete system of geographical coastal zonality will remain an unsolved problem, so far as the general features of coastal evolution have been firmly established. We are still far from this aim today. Some examples of zones and subzones on selected meridional transects may elucidate the problem (see also Fig. 1).

The zonality of the coasts of Europe and the Middle East is particularly influenced by two special conditions: the warming effect of the Gulf Stream in the north and the semi-enclosed oceanic basin of the Mediterranean Sea. Thus, it is only north of the Arctic Circle in the fresh or brackish waters of fiords that reach far inland and into the Baltic Sea, where the coastal influence of sea ice becomes recognizable. Between coastal freezing of the Baltic and permanent sea ice around the North Pole, another zone can be found without shore and sea-ice contacts because of the warming of the Gulf Stream around northernmost Norway (Martini et al. 2009). In hard rocks, there are only fiords and skerries affected by contemporary coastal processes. Further to the equator, the temperate coastal zone follows (Pratolongo et al. 2009).

The next important morphodynamic boundary marks the equatorial limit of frost weathering which reaches 54°N and comprises parts of the German coasts and eastern coasts of the British Isles. Thick medium-sized algae are found in the rocky littoral (*Fucus*, *Laminaria*, and others) that extends to about 46°N in France. South of 44°N latitude in the Mediterranean, calcareous algae and encrusting vermetids create youthful organic rock formations, which sometimes grow together to form trottoirs and even microatolls, particularly south of 36°N latitude. South of 38°N latitude, the occurrence of beachrock increases as do coral reefs south of 30°N in the Red Sea; while south of 28°N, the northernmost outposts of mangroves can be found at the Sinai Peninsula.

The Mediterranean may be used to illustrate problems inherent in the field methods that are commonly applied. Davies (1980) classified coasts within the zone of the “lower latitudes” and so did Valentin (1979), specifying them as “subtropical.” Davies puts the Mediterranean as well as the whole of Australia into the same zone; whereas in Valentin’s system, Australia is divided into a tropical and a subtropical zone. Both authors use the same major elements for classification, that is, the distribution of corals,



Geographical Coastal Zonality, Fig. 1 Modern geographical coastal zones of the world (Kelletat 1995)

mangroves, and beachrock. But it was only the existence of beachrock that caused the inclusion of the Mediterranean into the subtropical zone or into the zone of the lower latitudes, respectively. In the southern parts of the Australian coast, beachrock does not exist, and the subtropical zone would be erroneously shown to extend beyond the beachrock distribution area. On the other hand, the boundary of the coral reef distribution area is practically the same as that of beachrock at the east coast, whereas mangroves are found in some areas of

the south coast far beyond the usual areas of the subtropics in the Northern Hemisphere. This is an important anomaly.

Along the east coast of North America, the zonality of presently active coastal processes is more closely spaced (Kelletat 1989). The effects of sea ice on rock and beach coasts extend as far south as 46°N, and to around 40°N significant frost weathering processes can be observed. South of 35°N there are compound oyster reefs, that is, organic rock formations, together with an increasing number

of beaches with massive shell deposits. From 34°N to 30°N, hypersaline tidal flats occur in places almost completely free of vegetation (or only with microscopic plant material present), especially around the highest spring tide level. Latitude 28°N marks the poleward boundary of mangroves, 26°N that of the coral reefs. In total, there is only a separation of 20 latitudinal degrees between the coral reefs of the tropical zone and the sea-ice effects of the subpolar zone, in comparison to at least 38° on the European side of the Atlantic Ocean. These findings are in contrast with those on the coasts of eastern Australia and New Zealand. Here, the morphological impact of both sea ice and frost on the coastal environment is missing. Giant kelp (*Durvillea*, *Macrocystis*, and others) can be found as far north as 41°S, thus effectively protecting the rocky coasts from heavy surf. Serpulids and sabellariae form solid organic substrates as far north as 38°S (eastern Australia) and 37°S (New Zealand), respectively, whereas oyster reefs reach as far south as 33°S (Australia) and 37°S (New Zealand).

Within tidal limits, hypersaline tidal flats almost free of vegetation are found as far south as 42°S in New Zealand and 28°S in eastern Australia. Latitude 38°S is the southern boundary of mangroves. Just south of the limit of massive coral reefs, that is, 25°S, is the change from the tropical to the extratropical species in the rocky littoral.

The zones of giant subpolar algae and mangrove communities commonly reach within 3° latitude and sometimes even overlap. In contrast, the distance between them being 14° on the North American east coast and at least 18° in Europe to Middle East. In Florida, the northern boundary of the mangrove zone is 2° north of that of the coral zone, whereas in Australia/New Zealand it is 13°, but in North Africa/Middle East (e.g., Sinai Peninsula), the coral reefs extend 2° farther north than the northernmost mangrove, *Avicennia marina* being the species in all areas. These differences cannot properly be explained yet, mainly because the extreme and mean values of both air and water climates differ significantly at these boundary lines.

Thus, both zonal and azonal features at the coasts are manifold, zonal aspects are mainly given by climatic and biogenic parameters (at least the latter partly depends on the first), and geodynamic, structural, petrographic, or sedimentological properties dominate the azonal aspects.

Most authors have tried to define a rather simple system of coastal zonality, either by separating tropical (including subtropical) and extratropical areas or by dividing the Earth's coasts into high, mid, and low latitudes. In his synthesis of 1980, Davies characterizes these zones as follows:

High-latitude coasts: They are frozen part of the year and seasonal ice is active on the beach. Characteristics are low wave energy because of freezing and limited fetch, many pebbles at the beaches due to former glaciation and frost weathering, barriers developed well and widespread, dunes

rather limited, and tidal flats dominated by grassy vegetation. Cliffs are mainly formed by frost action; mass wasting is common, often due to permafrost decay. Further categories within high-latitude coasts may be determined by the duration of the sea-ice cover.

Mid-latitude coasts: They show high wave energy and frontal storm maxima, frequent cliffs in hard rock with quarrying and abrasion, sloping intertidal rock platforms, and weaker developed barrier systems but with partly very extensive dune areas. Biogenic effects are believed to be insignificant, except for dunes and tidal flats. A differentiation of this broad zone may be made by the limit of Ice Age glaciers deeply influencing the types of sedimentation along the coastline.

Low-latitude coasts: They are exposed to a relatively permanent approach of wind, waves, and swell, and only reduced wave energy is active. Algae and corals contribute to coastal forming, and lithification of dunes and beaches may occur. The widespread coastal sediments are of fine-grain deposits. Dunes exist in limited areas only, and rocky coasts with horizontal rock platforms are restricted. Perpendicular cliffs are very rare (except in arid areas); salt weathering is typical as well as extensive tidal woodlands (mangroves). A differentiation should be made regarding the more humid and the more arid environments.

Although the spatial distribution of some important coastal phenomena is generally known, it is difficult or nearly impossible to define the zonal limits by means of climatic data (although radiation and climate are the main sources of the zonal patterns of the Earth).

The 18 °C-isotherm of surface water in the coldest month of the year is the thermal factor which sets limits for the coral reef zone. For other coastal features, a definition is much more complicated; mangroves occur in very humid to extremely arid regions (e.g., precipitation nearly zero to thousands of millimeters per year) and within limits of low to high evaporation. Mean annual air temperatures at mangrove sites vary from about 28 °C to nearly 15 °C, those of the coldest month from more than 25 °C to only 7 °C, and coldest water temperatures from more than 25 °C to nearly 6 °C. As thermal limit for mangroves (*Avicennia* sp. or *Kandelia* sp. in Japan), several frost days with air temperatures of −4°/−5 °C may be defined, whereas the number of months with mean temperatures below 20 °C may reach from twelve to zero (as well as the number of arid months).

The climatic data for beachrock are at least as heterogeneous as for the polar limits of mangroves. They are annual precipitation from nearly zero to several thousands of millimeters, low to high evaporation, zero to twelve arid months, mean annual air temperature between 15 °C and more than 27 °C, mean temperature of the coldest month down to 7 °C, as well as water temperature, which may never reach 20 °C during the year as in Namibia. When comparing the

beachrock distribution on the islands of Oahu (Hawaii) and Madagascar, a clear relationship to the arid areas cannot be detected; microclimatic aridity may be more important than high water temperature. Both islands are situated in the core zone of beachrock distribution. In its marginal fringe like the Mediterranean Sea, beachrock in higher frequency and quality is clearly restricted to the warmer and more arid southern and eastern parts. It is only sporadic and poorly developed in its more humid and colder north and west.

In higher latitudes, the equatorial limits of giant kelp at the coasts show a similar and rather high differentiation (i.e., annual precipitation from around 300 mm to more than 1300 mm, mean annual temperatures from 7.5 °C to at least 17 °C, and the coldest month even from 13 °C down to –5 °C). The limiting factor seems to be air temperature, not exceeding a monthly average of 20°/21 °C. The average water temperature may vary from below zero to more than 15 °C. Twenty degrees centigrade seems to be lethal.

All features discussed here belong to the younger Holocene or even to contemporary decades, that is, an extremely short geologic period under a stable climatic regime. It is, however, difficult and till now partly impossible to commit to a satisfactory definition of limiting climatic conditions for most of the coastal features. In this respect, the related sciences are far from a consensus.

For example, it is difficult to define zones and subzones by coastal forms because of their multi-temporal origin, that is, from very different climates and, therefore, inherent with aspects of other zones. It may be prudent to restrict a consensus to the distribution of actual coastal processes (or those of the younger Holocene, when sea level was roughly at its present position). This will show a more homogeneous picture, comparable for all zones. A difficult, time-consuming problem that remains is detecting and evaluating processes. Due to the present state of research, we usually attempt to infer the process itself from its result (i.e., we infer cementation from the occurrence of beachrock and horizontal rock platforms from the process of salt/frost weathering or water-layer leveling, etc.). A fundamental question is, whether we should define a zone or subzone by the dominant process (if one exists) or by a combination of more than one process. Does a zone or subzone need to show only adequate processes or their forms? Should they dominate spatially, or should a zone be characterized by one process even if its geomorphological effect and importance is minimal? Additional problems include insufficient or inadequate knowledge of coastal forms and processes worldwide as well as regionally and thematically.

Conclusions

In reference to coastal features, one is faced with problems that are dominated by azonal geodynamic, geological,

structural, tectonic, petrographic, or sedimentological factors and by seawater or surf conditions, by climatic parameters, or by biospheric processes. Integration of all these categories into only one model of coastal zonality seems not to be applicable.

Knowing that contemporary coasts are formed during the last 7000 years, an extremely short timespan in geological and geomorphological times, the question arises as to whether the morphodynamic patterns now detected are representative of the last interglacial. Has the Holocene reached its climax conditions or is this still in process? An ancillary question regards zone shifts when climatic conditions changed. If it is true that coral reefs have their lethal winter temperature limit at 18 °C and giant kelp at about 20 °C in summer, a general change of 0.5 °C to 1.0 °C (which may occur within decades) would transfer their zonal limits for hundreds of kilometers. Singular and azonal events like changing ocean currents or El Niño effects may reinforce this tendency.

Being aware of the delicate balance of geomorphological zones and subzones during a given geological or climatic condition, it may be interesting to reconstruct those zones and subzones valid for former geological periods (which might have been longer lasting once they had reached their climax with certainty) and to learn from these results. On the other hand, by detecting coastal zones of former periods, it sometimes may be possible to deduce their environmental (i. e., paleoclimatological) parameters. As a synthesis, a complex world map (Fig. 1) may show the state of the art in inductive coastal zonality.

Cross-References

- [Arctic, Coastal Geomorphology](#)
- [Beachrock](#)
- [Beach Sediment Characteristics](#)
- [Climate Patterns in the Coastal Zone](#)
- [Coral Reefs](#)
- [Desert Coasts](#)
- [Glaciated Coasts](#)
- [Holocene Coastal Geomorphology](#)
- [Ice-Bordered Coasts](#)
- [Mangroves, Geomorphology](#)
- [Weathering in the Coastal Zone](#)

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Geohydraulic Research Centers

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Definition

Geohydraulic research centers are research organizations that specialize in science and engineering research relating to oceans, coastal regions, and estuaries. In particular, they have laboratory facilities including flumes and basins where waves and currents can be generated and interact with coastal structures and geomorphology.

Introduction

Several geohydraulic research centers specialize in coastal science and engineering research. Some are current or former government research facilities that have been partially or completely privatized. The centers have laboratory facilities used for both research and scale-model studies to aid in project design. Some centers have developed numerical models for project design. The following summarizes capabilities of the major geohydraulic research centers.

Canadian Hydraulics Centre

The Canadian Hydraulics Centre (CHC), located in Ottawa, Canada, is a government agency and a unit of the National Research Council, Canada’s leading scientific research organization (National Research Council Canada 2017). The CHC has extensive laboratory facilities in an 8,400 m² building.

Facilities include a 14.2-by-63 m coastal wave basin with a wave generator that produces long-crested, irregular waves with significant heights up to 0.4 m, and the basin has high-efficiency wave absorbers. The basin can be configured to generate bidirectional currents and combined wave and current interactions. There is a multidirectional wave basin that is 30-by-36 m and 3 m deep with a 60-segment directional wave generator that can produce irregular waves with heights up to 0.5 m and regular waves up to 0.75 m, and the wave generators have active wave absorption. NRC has a 30-by-50 m large area basin for large 3D studies with wave generators capable of producing irregular long-crested waves with significant heights up to 0.15 m and regular waves up to 0.25 m. A second directional wave generator can produce irregular waves with significant heights up to 0.4 m. The large area basin has pumps that can generate discharges up to 1600 L/s. There is an ice tank that is 7-by-21 m and 1.1 m deep that has a towing carriage and can accommodate a wave machine. The temperature inside the chamber can be varied down to –20 °C and can produce ice of up to 0.6 m.

The NRC has a large wave flume that is 2-by-97 m and 2.75 m deep. It has a wave generator that can produce irregular waves with significant heights up to 0.75 m and regular waves up to 1.25 m; the wave generator has active wave absorption. The NRC has a steel wave flume that is 1.2-by-64 m and 1.2 m deep. It has a wave generator that can produce irregular waves with significant heights up to 0.25 m, and it has active wave absorption. There is a high discharge flume that provides a steady flow in an open channel with continuously adjustable discharge from 0–1700 L/s.

Catalonia University of Technology, Spain

The Maritime Experimental and Research Flume was built in 1992 at the Maritime Engineering Laboratory of the Catalonia University of Technology for experimentation in maritime harbors and coastal engineering (Catalonia University of Technology 2017). The flume is 3-by-100 m and 7 m deep and has a wave generator that produces irregular waves and regular waves with maximum heights of 1.6 m. There is a bidirectional pumping system that can pump 2000 L/s.

Centre for Marine and Renewable Energy, Ireland

The Centre for Marine and Renewable Energy (MaREI), Ireland, is the marine and renewable energy research and development organization supported by the Science Foundation Ireland (Centre for Marine and Renewable Energy 2017). The Environmental Research Institute of the University College of Cork coordinates MaREI efforts with 130 researchers working across six academic institutions and collaborating with over 45 industry partners. MaREI has a facility for small- to medium-scale laboratory testing of ocean and maritime systems and a 2600 m² facility that houses four wave tanks.

MaREI's ocean basin is 18-by-25 m and 1 m deep with a wave generator that produces irregular waves with a peak significant wave height of 0.16 m, peak period of 1.4 s, and maximum height of 0.32 m. Its deep ocean basin is 12-by-35 m and 3 m deep with a wave generator that can produce significant wave heights of 0.6 m, peak period of 2.7 s, and a maximum wave height of 1.1 m. Its wave and current flume is 3-by-28 m with a depth of 1.2 m that can study unidirectional waves and wave-current interaction. Its wave generator can produce a significant wave height of 0.16 m, a peak wave period of 1.5 s, and a maximum wave height of 0.35 m, and the facility can generate currents of 1500 L/s. There is a wave flume for students that is 0.75-by-15 m and 1 m deep.

Coastal and Hydraulics Laboratory, United States

The Coastal and Hydraulics Laboratory (CHL) is in Vicksburg, Mississippi, USA, and was formed in 1996 through a merger of the Hydraulics Laboratory founded in 1929 (originally the Waterways Experiment Station) and the Coastal Engineering Research Center founded in 1930 (predecessor organization was the Beach Erosion Board) (Coastal and Hydraulics Laboratory 2017). CHL is part of the seven-laboratory complex of the Engineer Research and Development Center (ERDC) of the Corps of Engineers that includes seven research laboratories in four states. The ERDC has 2,200 employees, including 250 in CHL.

CHL has the largest coastal and hydraulic laboratory complex in the world with about 140,000 m² of laboratory facilities under roof. About a quarter of these facilities are used for coastal engineering research and scale modeling with the remainder used for hydraulic engineering research and scale modeling. CHL has several wave flumes, all with irregular wave generators including the following: 2-by-105 m, 1.4 m water depth, and 0.45 m wave height; 3.5-by-80 m, 1.4 m water depth, and 0.45 m wave height; 3-by-65 m, 1 m water depth, and 0.5 m wave height; 1.5-by-65 m, 1 m water depth, and 0.5 m wave height; 0.6-by-60 m, 1 m water depth, and 0.6 m wave height; 1-by-45 m, 0.6 m water depth, and 0.2 m wave height; and 1-by-45 m, 0.6 m water depth, and 0.2 m wave height. CHL's L-shaped facility tests 3D stability of structures such as breakwaters and is 15-by-75 m at the wave generator and 25 m wide at the testing end. It operates with water depths up to 1.5 m and uses a wave generator with a wave board that is 4 m high and 15 m wide and produces 0.6 m waves.

CHL has about 30,000 m² of covered laboratory space for 3D wave basins used for studying waves in harbors, movable-bed modeling, coastal processes, wave interaction with ships and other floating structures, and other coastal engineering applications. Large remote-controlled scale-model ships (e.g., container ships) measuring about 7 m in length can be employed to test ship maneuverability in irregular directional seas with currents and wind. Long-wave harbor resonance is tested in a facility with a wetted area of over 4,000 m² and 14 irregular wave generators, each approximately 5 m long, which generate waves with periods from 1 to 10 s.

Its Long Shore Transport Facility is a wave basin that is 30-by-50 m (expandable to 80 m) and 1.4 m deep that is used to study longshore currents and sediment transport. Wave generators produce regular or irregular waves with heights up to 0.4 m, periods of 1.0–3.5 s, and angles up to 20° from shore normal. A longshore current recirculation system has 20 independent upstream and 20 downstream flow channels each with a turbine pump that discharges up to 75 L/s. The system controls cross-shore distribution of wave-driven longshore currents and recirculates currents from downstream to upstream. Downstream flow channels have sediment traps that record sand accumulation and a dredging system that recirculates sand collected in the traps to the upstream end.

CHL's ESTEX Hyperflume facility is used for conducting unsteady, nonuniform flow and sediment transport research and is 15-by-18 m, 3 m deep and connected to an 18-by-107 m and 1.2 m deep basin and a parallel 3-by-122 m and 3 m deep basin. There is a 15 million liter full-scale test facility to study levees and develop new methodologies to assist in flood-fighting efforts. CHL's has 3D ship and tow simulators for design of navigation channels that include a complete full-scale ship bridge and real-time visualization and can simulate ships, tows, and small crafts.

The Coastal Modeling System developed by CHL is a suite of coupled 2D numerical models for simulating waves, hydrodynamics, salinity and sediment transport, and morphology change. CHL's Beach-fx model evaluates the physical performance and economic benefits and costs of shore protection projects. CHL's 2D and 3D Surface-Water Modeling System and Groundwater Modeling System are now offered through the private sector company Aquaveo.

Danish Hydraulics Institute, Denmark

The Danish Hydraulics Institute (DHI) is an engineering consultant firm headquartered in Denmark with 30 offices throughout the world, largely involved in hydraulic and hydrological modeling software (Danish Hydraulics Institute 2017a, b). It was founded in 1964 by the Technical University of Denmark and, although independent, is associated with the Danish Academy of Technical Sciences. It has an offshore wave basin in Hørsholm, Denmark, that is 20-by-30 m with a depth of 3 m. The basin has a wave generator with 60 individually-controlled flaps that can produce short-crested irregular 3D waves and swell. Wind can be produced by fans and pumps can be produced currents. There is a shallow basin that is 25-by-35 m with a depth of 0.8 m and a multidirectional segmented irregular wave generator, currents that be generated by external recirculation of water, and wind generated by fans. The facility has a flume that is 0.74-by-28 m with a depth of 1.2 m.

DHI has a suite of 1D, 2D, and 3D numerical models under its MIKE umbrella. For example, MIKE 3 is a 3D modeling system that simulates flows, cohesive sediments, water quality, and ecology in rivers, lakes, estuaries, bays, coastal areas, and seas in three dimensions. MIKE SHE is an integrated hydrological modeling system for surface and groundwater flows.

Deltares, Netherlands

Deltares is a major technological institute in the Netherlands that focuses on river basins, coastal regions, and river deltas (Deltares 2017a, b). Delft Hydraulics, a famous hydraulic laboratory founded in 1927, became a part of Deltares in 2008. Deltares has over 800 employees located in Delft Utrecht in the Netherlands.

Hydraulic/coastal facilities at Deltares include its Atlantic Basin, built in 2009. This is a multifunctional facility with a total size of about 650 m², in which waves and tidal currents can be simulated. The basin is 8.7-by-75 m with a height of 1.3 m, allowing a water depth of 1 m. In the test section, it has a closable pit of 130 m² and a depth of 0.5 m to allow modeling of a mobile bed. A sediment sieve is constructed

at the downstream side for the collection of lightweight material in morphological tests. The wave generator consists of 20 independently-controlled wave paddles that make it possible to produce oblique waves with directional spreading and with a maximum significant wave height of 0.25 m at a period of 2 s and a maximum wave height of 0.45 m. The wave generator has automatic reflection compensation. A current can be generated via a pumping system either following or opposing the waves with a maximum pumping capacity of 3000 L/s. Depending on the project-specific needs, the basin layout can be modified, for example, by building wave-damping beaches or sluice gates.

The Pacific Basin is a multifunctional wave-current basin that is used for coastal- and offshore-related projects in which 3D effects are important, such as testing breakwater heads and foundations of offshore structures. The wave basin is 22.5-by-30 m and has a basin height of 1.25 m, allowing a maximum water depth of 1 m with a minimum depth of 0.25 m. The wave generator has a width of 14 m and a wave board height of 1.2 m. It can generate waves with a frequency range of 0–2 Hz, a maximum significant wave height of 0.21 m at a wave period of 1.9 s, and a maximum wave height of 0.4 m. It has a pumping system with one-directional flow and a capacity of 1800 L/s.

The Scheldt Flume is a 1-by-110 m wave flume with a depth of 1.2 m that is used for studies such as breakwater armor stability and wave impact loading. The flume has wave generators at both sides that enables one side to be used for wave generation and the other for wave dissipation. The wave generators can produce significant and maximum wave heights of 0.25 m and 0.4 m, respectively. At one side of the flume, a pumping system is installed that has a maximum pumping capacity of 600 L/s.

The Delta Basin is a 50-by-50 m wave basin with a water depth of 1 m that is equipped with two multidirectional irregular wave generators with a total length of 66.4 m, placed at a right angle to each other. The wave generators can produce both periodic and irregular long-crested or short-crested waves. Both wave generators are equipped with active reflection compensation. Maximum wave heights and significant wave heights are 0.6 m and 0.25 m, respectively.

The Delta Flume is 5-by-291 m and 9.5 m deep. Its wave generator produces maximum wave heights and significant wave heights of 4.5 m and 2.2 m, respectively, with wave periods of 1–20 s. The facility makes it possible to test at full scale the effect of extreme waves on dikes, dunes, breakwaters, and offshore structures.

The Deltares Intake and Outfall Structure (IOS) basins 1 and 2 can be used for model testing of most hydraulic structures. The large IOS-1 basin is 12.5-by-20 m and allows for modeling cross flows of channels or rivers. The smaller IOS-2 basin is 6-by-16 m and allows higher water levels, so that smaller designs can be studied on a larger scale. There are

three pumping stations with a capacity of 1000 L/s to pump water between the reservoir and the flume, making it possible to produce high waves and at the same time to simulate variations in the water level caused by storm surges and tides.

The AlphaLoop is a facility for investigating a wide range of multiphase flows in pipeline systems. The AlphaLoop allows for single-phase (water), two-phase (water and air), and three-phase (water, air, and sediment) experiments.

Deltares has its Delft3D software that provides 3D computations for coastal, river, lake, and estuarine areas. It calculates flows, sediment transport, waves, water quality, morphological developments, and ecology. In addition, Deltares has the software, XBeach, a 2D model for wave propagation, long waves and mean flow, sediment transport, and morphological changes of the nearshore area, beaches, dunes and back barrier during storms.

Flanders Hydraulics Research, Belgium

The Flanders Hydraulics Research is a division of the department of the Flemish Ministry of Mobility and Public Works, Belgium (Flanders Hydraulic Research 2017). It was founded in 1933 as a part of the Antwerp Maritime Services, became part of the Ministry of Public Works and the Waterways and Marine Affairs Administration in 1945, and in 2006 was integrated into the Flemish Ministry of Mobility and Public Works.

Facilities include a wave flume that is 4-by-70 m with a depth of 1.4 m. The wave generator can produce waves of height 0.65 m at a depth of 0.9 m. There is a small flume of length 0.7-by-3.17 m and depth of 0.86 m that has a wave generator that can produce 0.4 m height waves in a water depth of 0.6 m.

A 3D facility is 12.2-by-17.5 m with a depth of 0.45 m and a testing area that is 10-by-12 m. The facility has an adjustable water level and can be used with or without a moveable bed to test wave overtopping, wave intrusion into harbors, stability of beaches, and wave-current interaction. The facility has a moveable multidirectional irregular wave generator that is 12 m wide and can produce short- and long-crested waves. A pump with a maximum flow of 175 L/s can create current velocities up to 0.3 m/s both in the direction of the wave movement and against this direction.

A sloping bed current flume is 0.5-by-34.8 m with a depth of 0.76 m and can be used to calibrate measurement devices such as velocimeters. A multifunctional test basin is used for research on locks (design of filling and emptying systems, hawser forces), controlled inundation areas with reduced tide (design of inlet and outlet sluices, submersible dikes), and hydraulic characteristics of structures. It is 9.8-by-19 m with a depth of 1 m and can be subdivided in half. One half is operated as current flume with a pump capacity of 500 L/s

and maximum upstream water depth of 1.45 m. The other half is for lock modeling with a pump capacity of 120 L/s and water depths of 0.5–1.2 m. There is a 52 m² test area. Another current flume that can study flow patterns and head losses associated with weirs, fish passage, and other phenomena is 2.4-by-56.2 m with a depth of 1.15 m. Pumps can generate a maximum flow of 600 L/s.

A towing tank to study ship behavior in confined areas is cooperated with Ghent University. It is 7-by-88 m (67 m of the 88 m can be used for experiments) and maximum water depth of 0.5 m. Ship models are usually about 4 m (scale of 1:50–1:85). A wave generator can generate both regular and irregular waves to study the vertical motion of a ship under the influence of waves. An auxiliary carriage can move a second ship model on a straight trajectory to determine forces exerted on a ship by ships passing or overtaking it.

For studies of the Port of Zeebrugge, there is a scale model with a total area of 2000 m², horizontal scale of 1:300, vertical scale of 1:100, velocity scale of 1:10, time scale of 1:30, and discharge scale of 1:300,000. The model simulates a coastline of 15 km and extends a simulated 10 km into the ocean. The model has been used to study large cross currents at high water that limit inbound container and liquefied natural gas carriers to a window of passage of 4–6 h. In addition, it has been used to study methods to reduce dredging costs.

HR Wallingford, United Kingdom

HR Wallingford is an independent civil engineering and environmental organization in the United Kingdom with offices around the world (HR Wallingford 2017). It has the UK designation of a Scientific Research Association with company members that include many UK organizations including the British Ports Association, Association of Consultants and Engineers, Institution of Civil Engineers, and Royal Academy of Engineering.

HR Wallingford facilities are housed in buildings of 14,400 m². It has six wave basins ranging in size from 25-by-32 m to 32-by-75 m. Its largest wave basin is one of the largest unobstructed hydraulic test tanks in the world. Wave generators have active wave absorption and produce waves up to 0.25 m in height in water depths up to 0.8 m. A 27-by-55 m wave-current basin can generate uni- or bidirectional currents of up to 1200 L/s. There are two wave flumes that are 1.2-by-45 m and with operating depths of up to 1.5 m. There is a flume that is 1.8-by-100 m with an operating depth of up to 1.8 m. The flumes can produce regular and irregular waves with significant wave heights up to 0.3 m.

HR Wallingford is a member of a consortium of organizations including Artelia (formerly Sogreah), France; Bundesanstalt für Wasserbau, Germany; and Centre d'Etudes Techniques Maritimes et Fluviales, France that manages

TELEMAC-MASCARET, an integrated suite of software for free-surface flow (Telemac-Mascaret 2017).

Kajima Research Institute, Japan

The Kajima Research Institute (KRI) was established in 1949 and is in Tobitakyu, Chofu, Tokyo, Japan (Kajima Research Institute 2017). It is a research institute of Kajima Corporation, a private sector construction company.

KRI has a laboratory facility built in 1975, that is, in a 2,870 m² building. It houses a wave basin that is 20-by-58 m and 1.6 m deep, and its directional irregular wave generator can produce waves up to 0.6 m in height with wave periods from 0.5 to 5 s. The wave generator has active wave absorption. The facility also has a tsunami wave generator, and the basin is equipped to generate tides. KRI has a large wave flume that is 2-by-62 m and 2 m deep that is used to test wave forces on structures and other 2D phenomena. Its wave generator can produce waves up to 0.6 m in height, absorb reflected waves, and generate low-frequency, arbitrarily shaped wave forms to study tsunamis. There is a medium-size wave flume that is 0.7-by-60 m and 1.5 m deep, and its wave generator can produce irregular waves with maximum heights of 0.5 m and absorb reflected waves. Currents with a maximum discharge of 200 L/s can be generated in the flume.

Ludwig-Franzius-Institute, Germany

The Ludwig-Franzius-Institute for Hydraulic, Estuarine, and Coastal Engineering was founded in 1916 at the Leibniz University of Hanover, Germany (Ludwig-Franzius-Institute 2017). It has a 16,800 m² covered laboratory facility housing the second largest wave flume in the world that is 5-by-307 m with a depth of 7 m and a wave generator with active wave absorption that produces regular and irregular waves with a maximum wave height of 2 m. It has a 3D wave basin that is 25-by-40 m with a depth of 1 m and a directional irregular wave generator with 72 individual wave boards that can produce maximum wave heights of 0.47 m and has an active wave absorption system. In addition, the facility can generate currents of 300 L/s. It has a second wave flume with a length of 2.2-by-110 m and depth of 2 m and an irregular wave generator that can produce maximum wave heights of 0.35 m.

Manly Hydraulics Laboratory, Australia

The Manly Hydraulics Laboratory (MHL) is in Manly Vale, a suburb of Sydney, Australia (Manly Hydraulics Laboratory 2017). MHL was established in 1944 to provide specialist services relating to water, coastal, and environmental

challenges. It is now a business unit within the government and corporate service division of the Department of Finance, Services and Innovation.

MHL has 2,230 m² of covered experimental space and 3,530 m² of open-air experimental space. It has a 1-by-30 m flume with a depth of 1.8 m and adjustable floor and an irregular wave generator, an 0.6-by-11 m tilting flume with glass panels that is 0.8 m deep, and a 20-by-30 m wave basin that is 1 m deep wave basin with an irregular multidirectional wave generator.

Norwegian Marine Technology Research Institute, Norway

The Norwegian Marine Technology Research Institute (MARINTEK) performs research and development for companies in the field of marine technology (SINTEK 2017). It is part of SINTEK, the largest independent research organization in Scandinavia that is headquartered in Trondheim, Norway. The Norwegian University of Science and Technology is a public research university in Norway. It has a hydraulic laboratory at its campus at Trondheim University that has an ocean basin that is 50-by-80 m with a depth up to 10 m and maximum current velocities of 0.2 m/s at a 5 m water depth. It has a wave generator that can generate regular waves with a maximum height of 0.9 m and wave periods greater than 0.8 s and irregular waves with a maximum significant wave height of 0.5 m. A second wave generator has 144 individual wave boards and can generate regular waves with a maximum height of 0.4 m with wave periods greater than 0.6 s and multidirectional irregular waves with a significant height of 0.2 m and wave period greater than 0.7 s.

Oregon State University, United States

Oregon State University has the largest wave flume in North America (Oregon State University 2017). It is 3.7-by-104 m and has a depth of 4.6 m. The flume has a maximum water depth of 2 m for studies of tsunamis and 2.7 m for wind and storm waves. Its wave generator can produce regular, irregular, or user-defined waves such as tsunamis and has a wave period range of 0.8–12 s. It can produce a maximum wave height of 1.7 m at a wave period of 5 s or a 1.4 m tsunami.

Queen's University, Canada

The Coastal Engineering Research Laboratory of Queen's University was built in the late 1960s in Kingston, Ontario, Canada, and has a covered area of 1750 m² (Queen's University 2017). The facility is operated by the Department of Civil

Engineering and is the largest hydraulic laboratory in Canada. The Laboratory has a coastal model basin that is 21-by-21 m and equipped with an irregular wave generator. Tides and currents can be reproduced in the basin and a sand bed can be used in experiments. The laboratory has an oscillating water tunnel for research on boundary layers, forces on structures, and other problems. It has three 45 m wave flumes, two of which are equipped with irregular wave generators and one with a regular wave generator.

Scripps Institution of Oceanography, United States

The Scripps Institution of Oceanography located in La Jolla, California, USA, has a hydraulic laboratory with facilities to address coastal and oceanographic research (Scripps Institute of Technology 2017). The facilities are in a building with 31-by-46 m.

A flume has interior dimensions of 2.39-by-44.5 m and is 2.44 m deep. Windows, the longest being 5.5 m, allow optical access from the flume side. The wave generator can produce a maximum wave height of 0.45 m at 5 s. An open-circuit wind tunnel produces variable wind velocities ranging from 0 to 16 m/s with a 1.5 m water depth. The wave generator can produce a maximum wave height of 0.6 m. An open-circuit wind tunnel produces variable wind velocities ranging from 0 to 16 m/s. There is a flume that is 0.5-by-33 m with a 0.5 m water depth, a wave generator that can generate 0.25 m waves, and a reversible flow system that can generate a maximum flow of 41 L/s.

Sogreah, France

Sogreah is a company founded in 1923 with hydraulic laboratory facilities in Grenoble, France (Artelia 2017). It merged in 2009 with the construction management company, Coteba, to form Artelia. Its hydraulic laboratory facilities include 11,000 m² of physical model space under roof and 3200 m² of outside space on the outskirts of Grenoble. It has five wave tanks including a 40-by-40 m multidirectional irregular wave generator, three wave flumes, and 1600 m² for scale-model testing. Artelia is a member of the consortium including HR Wallingford that manages TELEMAC-MASCARET, an integrated suite of software for free-surface flow.

Tianjin Research Institute, China

The Tianjin Research Institute of Water Transport Engineering in Tianjin, China, is part of the Ministry of Transport (Zhang and Geng 2015). It built the largest wave flume in

the world in 2014. The flume is 5-by-450 m and 8–12 m deep. Its wave generator can produce a regular wave with a maximum height of 3.5 m. Currents can be produced with a maximum velocity of 1 m/s and a flow rate of 2000 L/s in the forward direction and 1400 L/s in the reverse direction. A sand bed is in the test area of the flume and is 5-by-100 m and 4 m deep. The facility will initially be used for research on scale effects in models, stability of breakwaters and sea walls, breakwater overtopping, structure foundations, and sediment issues.

University of Delaware, United States

The University of Delaware's Center for Applied Coastal Research located in Newark, Delaware, USA, has an Ocean Engineering Laboratory built in 1981 that is among the largest university laboratories in the United States (University of Delaware 2017). It has a two-story 30-by-38 m facility for laboratory experiments in coastal and ocean processes. Its wave basin is 20-by-20 m and 1.1 m deep and has a directional irregular wave generator that has 34-flap individually programmed wave paddles. It has a wave tank that is 0.6-by-33 m and 0.76 m deep. Its wave generator can produce regular and irregular waves, and water can be recirculated in the tank and produce a current of 0.3 m/s. A recirculating Armstrong flume is 0.4-by-0.6 m deep and used to generate hydraulic jumps. The center also has a sand beach wave tank that is a towing and wave tank with a cross section of 1-by-2 m.

Summary

There are major geohydraulic research centers in 14 countries in North America, Europe, Asia, and Australia. The number of scale-model hydraulic facilities has decreased somewhat over the past decade as numerical models have been replacing some of the facilities. However, new facilities have been built such as the largest flume in the world that was completed in 2014 in China. The facilities aid in understanding ocean, coastal, and estuarine processes and evaluate the efficacy plans to expand harbors, restore beaches, increase coastal resiliency, and promote sustainable development.

Cross-References

- [Beach and Nearshore Instrumentation](#)
- [Coastal Modeling](#)
- [Cross-Shore Sediment Transport](#)
- [Engineering Applications of Coastal Geomorphology](#)
- [Longshore Sediment Transport](#)
- [Navigation Structures](#)

- Physical Models
- Shore Protection Structures
- Surf Modeling
- Wave Climate
- Waves

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Geomorphology and Sea Level

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Definition

Geomorphological sea-level indicators can be defined as geomorphological features, erosional or depositional, which may be used to obtain accurate information regarding the position of relative sea level. In order to increase both their accuracy and their significance, it is best to be used together with other sea-level indicators, i.e., to determine their chronology.

Introduction

The geomorphology and sedimentology of the contemporary coasts are owed to the sea level changes during the Holocene and to a lesser degree during the Pleistocene. Environmental conditions, e.g., the prevailing hydrodynamics, coastal topography, and sediment availability, influence the nature of the coast and its response to sea-level rise. Environmental change may alter geomorphic process rates and process regimes in coastal landscapes. The reconstruction of past coastal landscapes, as well the reconstruction of past sea levels, ultimately allows us to thoroughly understand the palaeoenvironment.

There are a number of coastal geomorphological features used as sea-level indicators, with variable accuracy. In general, they may be the result of either erosional or depositional processes. Depositional features include beachrocks and marine terraces, while erosional include notches, sea caves, potholes, and tafoni, among others. One of the advantages of depositional sea level indicators is that they may provide datable material, while erosional features are generally inadequate for dating former sea levels.

Marine Terraces and Benches

A marine terrace is a coastal landform, a relatively flat, horizontal, or gently inclined surface of marine origin, most often an old abrasion or denudation platform, which has been uplifted generally by tectonics or eustasy (Fig. 1). It is usually bounded by a steep ascending slope on one side and by a steeper descending slope on the opposite side and it lies above or below the current sea level. A terrace may be several hundred meters wide and may reach as much as 600 m above present sea level, like at Huon Peninsula, Papua New Guinea.

Geomorphology and Sea Level, Fig. 1 Marine terraces owe their formation to the combined effect of eustatic and tectonic processes



Marine terraces owe their formation to the combined effect of eustatic and tectonic processes (Pirazzoli 2007). The study of such landforms contributes significantly to Quaternary stratigraphy, the reconstruction of eustatic sea level changes, and tectonic movements (e.g., Santoro et al. 2013).

There are various methods of dating marine terraces: from correlation or by direct dating of related materials including ^{14}C radiocarbon (for Holocene and Postglacial terraces); $^{230}\text{Th}/^{234}\text{U}$, ESR, OSL (for Pleistocene terraces); or by identifying guide fossils, such as *Strombus bubonius* in the Mediterranean for the Tyrrhenian stage. Recently, the cosmogenic nuclides method has also been used through ^{10}Be and ^{26}Al cosmogenic isotopes produced in situ. Terraces of the Marine Isotope Stage (MIS) 5 to MIS 13 may be identified, mapped, and dated with this method.

From a geomorphological point of view, the best sea-level indicator of a terrace is the limit between the upper part of the terrace and the adjacent cliff base, especially if it is marked by a notch, because it can be correlated with the corresponding eustatic transgression (Pirazzoli 2007).

Marine Notches

Tidal Notches

Recently, the term “tidal notch” is used as a sea-level indicator for marks of coastal erosion in the scientific literature. Particular profiles of bioerosion that undercut sheltered vertical

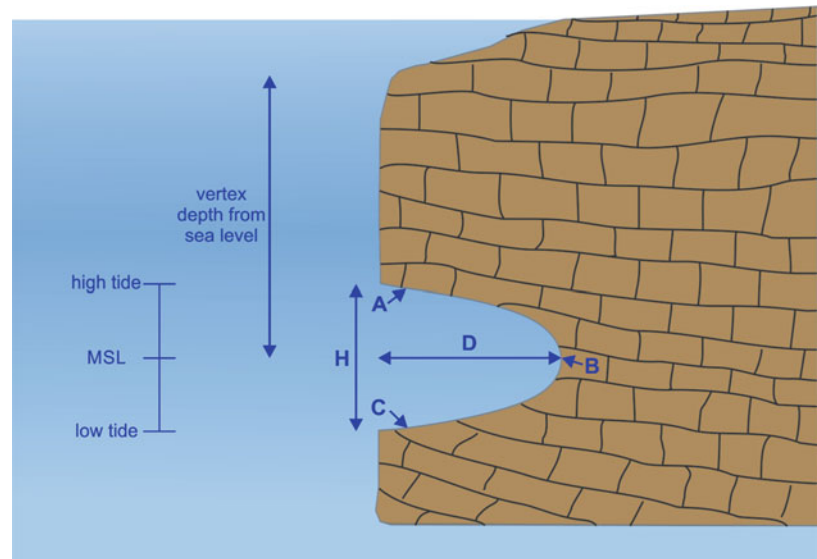
cliffs of carbonate rocks in the intertidal zone are called tidal notches, with a vertex at the biological mean sea level (BMSL), the limit between the midlittoral and infralittoral zones (see Laborel and Laborel-Deguen 2005). In the Mediterranean, a microtidal area, the BMSL corresponds, with an accuracy of ± 10 cm (Evelpidou and Pirazzoli 2016), to the local mean sea level (MSL). The most common profiles of bioerosion in tidal notches are recumbent V-shaped or U-shaped forms. The inward depth, D , of the maximum convexity, B , (vertex) is located near MSL. Above the base of the notch the notch floor, C , is flat if the tidal range is small (Fig. 2).

In a moderately exposed site, continuous wave action may splash sea water onto the roof (A). Thus the top of the notch will be shifted upward, even above the highest tide level. In this case the height becomes larger, while the base remains at the lowest tide level or slightly below it. The height (H) of the notch, therefore, increases with exposure; however, in very exposed sites, tidal notches may be lacking completely. If, after some deepening of the notch, the roof collapses, a new notch similar to the preceding one will form further inside the cliff.

Other marks of coastal erosion, due to variable rock resistance and nature or due to changes in erosional efficacy may be superimposed on the theoretical tidal notch profile. They can even modify the genetic identification of the notch. In such cases, the notches are often discontinuous; the notch base or part of the erosional profile has disappeared, dissolved by fresh water arrivals. Often, the presence of organic

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Level, Fig. 2 The basic characteristics of a tidal notch profile: A, C the notch roof and floor, respectively, B the vertex, D the inward depth, and H the height



accretions, developed on the notch floor, may transform the initial tidal notch into a surf notch (Pirazzoli 1986). In this case, its use as a sea-level indicator will depend on the species forming the organic accretions, rather than on the shape of the notch.

Tidal notches are of special interest, because they provide information regarding sea level changes, as well as tectonic displacements. More specifically, the profile of a tidal notch provides information on how long the mean sea level has remained constant at the vertex position (maximum retreat point), as well as on the speed of RSL change (e.g., gradual or rapid) (Pirazzoli 1986; Evelpidou et al. 2011). Tidal notches are considered as excellent sea-level indicators and have often been used in the Mediterranean and elsewhere to decipher relative sea level changes and vertical tectonic movements (e.g., Evelpidou et al. 2011; Evelpidou and Pirazzoli 2014, among many others).

Bioerosion Rates

Several estimates of erosion rates that are attributed to biologic activity have been noted in the literature (e.g., Pirazzoli 1986, Table 1; Evelpidou and Pirazzoli 2016 and references therein). In carbonate rocks the deepening of the notch profile is concentrated biologically in the midlittoral zone, where eroding Cyanobacteria, patellaceous gastropods (limpets) and chitons are abundant (Laborel and Laborel-Deguen 2005). Detailed estimations on rates of surface retreat in an intertidal notch show a range varying from less than 0.1 mm/a to 5 mm/a, depending on lithology, location and probably duration of bioerosion (for references, see Pirazzoli 1986; Evelpidou et al. 2011; Moses et al. 2015; Evelpidou and Pirazzoli 2016). Bioerosion rates reported from the Mediterranean generally vary between 0.2 and 1 mm/a (e.g., Laborel and Laborel-Deguen 2005; Evelpidou and Pirazzoli 2016).

Modification of a Tidal Notch Due to Relative Sea-Level Changes

If a movement of the relative SL occurs, the notch profile will be modified. In the case of emergence due to relative sea level fall, several results are possible (Fig. 3). A sudden emergence (e.g., coseismic) greater than the tidal range is a unique case, in which, the notch, completely emerged, is preserved from further marine modification while a new notch develops lower, in the new intertidal zone. A rapid relative SL fall with a vertical displacement smaller than the tidal range will allow only the notch roof to be preserved, while the floor is attacked and degraded from the new low-tide level; in this case, the height of the notch will increase.

If a rapid relative sea level change occurs faster than the local bioerosion rate, then the following cases may be found: In the case of relative sea level fall, the tidal notch will be preserved completely emerged; but in the case of sea-level rise two cases are possible: (1) if the SL rise is greater than the local tidal range, then the tidal notch will be preserved completely submerged (2) if the SL rise is smaller than the local tidal range, the tidal notch will be modified because its vertex will be displaced upward proportionally to the vertical displacement.

If multiple sudden vertical movements occur, successive overlapping indentations will produce so-called ripple notches (Fig. 4). A gradual slow rate of emergence (e.g., 1 mm/year) or a gradual SL oscillation will result in a symmetrical notch profile; however, the height of the profile will adapt to the amplitude of change. A gradual emergence, following a period of stable SL, will cause only the floor of the notch to be modified. In the case of a gradual emergence between two periods of stable SL, the notch will assume a reclined W-shaped profile. Many combinations of movements

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Level, Fig. 3 In case of emergence due to relative sea level fall, the tidal notch may be preserved uplifted and depending on the conditions different profile types will be developed

**Geomorphology and Sea**

Level, Fig. 4 Ripple notches may develop if multiple sudden vertical movements occur in a relatively short time period



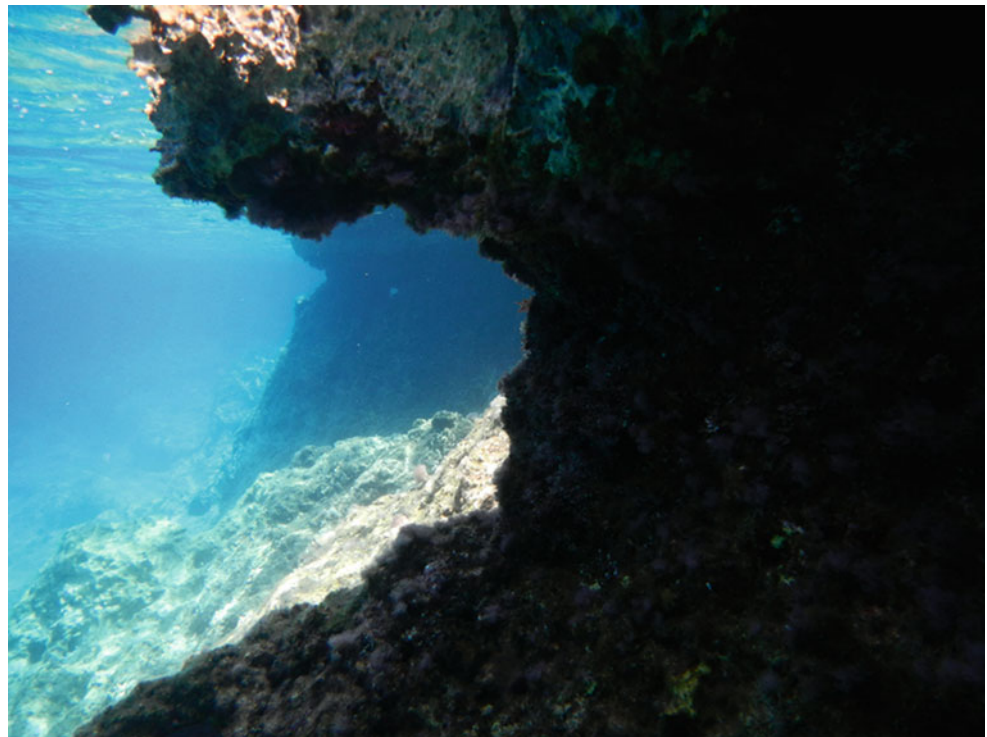
leading to submergence instead of emergence are possible. Very recent emergence may be indicated by limpet and chiton marks with an abrupt limit below the roof.

In the case of gradual submergence at a rate smaller than the bioerosion rate, the vertex of the notch profile will be displaced downward and the height of the notch will increase (Fig. 5).

A special case is gradual subsidence occurring at a rate exceeding the local bioerosion rate, as during the last two centuries; according to recent studies (e.g., Kemp et al. 2011, and references therein) the global sea level has risen by 6 cm during the nineteenth century and by 19 cm in the twentieth century, at an average rate of 2.1 mm/year. With such a sea-level rise, greater than the tidal range (e.g., in

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Level, Fig. 5 In case of gradual submergence at a rate smaller than the bioerosion rate, the height of the notch vertex will increase



most of the Aegean Sea, this rate exceeds the bioerosion rate), the consequences will be similar to those of a coseismic uplift of the same amount, i.e., the submergence of the tidal notch that was developing up to the nineteenth century (that has been called “modern”), while no new tidal notch will be developing near the present sea level (Evelpidou et al. 2012).

Other Types of Marine Erosion Notches

Solution Notches

It is well known that fresh water arrivals may cause phenomena of dissolution on limestone rocks. According to Higgins (1980), solution notches in calcareous rocks are only present near coastal springs, where surface seawaters are diluted locally by freshwater. In the case of a relative sea-level rise along with fresh water arrivals on a carbonate coast, the profile of a tidal notch will initially start its development, but dissolution will soon occur, tending to make the floor of the tidal notch disappear, while the roof of the notch remains. The continuation of sea-level rise will result in an upward gradual displacement of the notch roof, which will remain isolated, testifying to the former presence of a tidal notch. In a paper examining tidal notches in the Gulf of Corinth (Greece), Evelpidou et al. (2011) have proposed to call “visor” such remnants of a tidal notch by dissolution.

Structural Notches

Wave erosion exerted on weaker layers of a cliff or of a rock formation may leave a notch-like mark that is called structural notch; such features are not related to sea level and, thus, are not used as sea-level indicators.

Surf Notches

When water turbulence increases, organic incrustations tend to develop on the floor of a tidal notch (Pirazzoli 1986) and to protect from further erosion and a bench (surf bench) will begin to form. In these cases, the tidal notch will become a surf notch and its accuracy as sea-level indicator will depend on the species of the organic accretion. In the Eastern Mediterranean such surf benches are often called “trottoirs,” which are found usually not more than 0.2–0.4 m above MSL (Pirazzoli 2005).

Abrasion Notches

Pebbles or sand transported by waves may cause abrasion on a notch profile, producing a polished surface. In these cases the notch can be called an abrasion notch. According to Kelletat (2005), these types of notches are occasionally found nearby coastal caves in areas of less resistant rocks.

Wave-Cut Notches

The term wave-cut notch is sometimes found in the coastal literature. However, such vague term may refer to any type of marine notches.

Beachrocks

Beachrocks are defined as hard sedimentary formations, which consist of coastal sediments (sand, pebbles, etc.), lithified relatively quickly through the precipitation of carbonate cement (Fig. 6). Field surveys and laboratory experiments (e.g., Vousdoukas et al. 2007 and references therein) have shown that lithification is rapid, ranging from a few years to decades.

Beachrocks are found mainly in the Mediterranean, the Caribbean, on tropical and subtropical Atlantic coasts, as well as in the atolls of both the Pacific and Indian Ocean. As beachrocks are often found near sea level, it was considered they are forming in the intertidal zone and that their exposure is owed to coastal retreat.

Beachrocks typically form in the interface between seawater and meteoric water, i.e., the mixing zone. In particular, beachrock cements consist of metastable carbonate phases (e.g., Vousdoukas et al. 2007 and references therein): high magnesian calcite (HMC) or aragonite, depending on the physicochemical factors of the diagenetic environment.

Conflicting theories about the formation zone of beachrocks may be found in the literature. Some authors suggest their development in the intertidal zone (e.g., see Vousdoukas et al. 2007 and references therein), while Kelletat (2006) suggests a supratidal genesis in the spray zone, since beachrocks in microtidal areas often present a thickness much greater than the tidal range. Beachrocks are frequently

used for the study of coastal evolution, as well as for the relative sea level changes both in the Mediterranean and elsewhere (e.g., Karkani et al. 2017).

Beachrocks Formation

The cement of beachrocks is indicative of their development zone, a chemically active zone. The sedimentary layer, which is closest to the aquifer, is cemented firstly and faster, while surfaces where the microbes are active are preferable.

In microtidal or a-tidal areas, beachrocks may be characterized by small thickness, corresponding to the seasonal variation of the mean sea level (MSL), while in areas with higher tidal range, such as Australia, along the Great Barrier Reef, the thickness of the cemented layers may exceed 3 m.

Diagenesis takes place below the surface, responding to changes of the height of the aquifer, the temperature, or the pressure (Mauz et al. 2015). It involves processes, such as dissolution, re-precipitation, and re-crystallization to ultimately result in chemical stability. The study of the mineralogy and microstratigraphy of the cement can provide information regarding the formation zone of beachrocks, which is particularly valuable when beachrocks are used as sea-level indicators (Mauz et al. 2015). For example, high magnesian calcite (HMC) and aragonite are developed in circumgranular rims in meniscus fabric in the vadose environment, while in the meteoric zone they are developed in symmetrical crusts (Mauz et al. 2015).

Geomorphology and Sea

Level, Fig. 6 Beachrocks represent former shorelines and are often used to understand sea level fluctuations together with other sea-level indicators



Beachrocks as Indicators of Sea Level Changes-Concerns

Altitude Relationships

Beachrocks have proven to be particularly useful when no other sea-level indicators are present or when they are combined with other available indicators. The mineralogy and the morphology of the cement are indicative of the diagenetic environment, so the examination of microstratigraphy, and the characteristics of cement allow the determination of the spatial relationship between the coastline and the formation zone of the beachrock (Mauz et al. 2015).

Recently, Mauz et al. (2015) proposed a more reliable methodology for using beachrocks as sea-level indicators by associating the cement and the sedimentary structures with a particular tidal zone and assigning an error to the indicator, depending on the aforementioned characteristics. As already noted, beachrocks with large thickness have been observed in microtidal areas, so when they are used as sea-level indicators, without mineralogical information, a relatively large error must be expected, which may exceed 1 m (Mauz et al. 2015).

Dating

Accurate dating of beachrocks is a particularly difficult task, although biogenic material, such as shells or corals, can be easily dated with radiocarbon and, if special attention is given to avoid any re-crystallization, reliable results may be achieved (e.g., Mauz et al. 2015). In the case of biogenic material, the age attributed is the time of death of the

organism. After the organism died, it was transported from its in situ position, deposited on a beach, and was later cemented into the beachrock. Although, in many cases, it is possible this sequence to have taken place within a few years, it is likely the time difference between the death of the organism and its incorporation to the beachrock to reach hundreds of years. Therefore, the age attributed by dating an organism can be considered as the maximum age for the lithification of the beachrock (Fig. 7).

A minimum age for this process is possible only by dating the cement. However, several difficulties arise, mainly in the extraction of enough material. In addition, in many cases a beachrock may have undergone multiple diagenetic phases and therefore consist of different cement generations.

During the last decade, many efforts have taken place for dating beachrocks using the method of optically stimulated luminescence (OSL), aiming to reconstruct palaeoshorelines and sea level changes.

According to absolute datings, the beachrocks of the Eastern Mediterranean were mainly formed between 1000 and 5000 years ago (see Voutsoukas et al. 2007, Table 1).

Other Geomorphological SLI

Potholes

Potholes are vertical, circular, and cylindrical erosional features in consolidated rocks, found both in fluvial and coastal

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Level, Fig. 7 Although biogenic material may be found incorporated into the beachrock, its age should be considered as the maximum age for the lithification



environments. They are developed in zones of breaking rocks, where sand or pebbles act as abrasion tools within depressions, physically excavating the rock (e.g., Taboroši and Kázmér 2013). Their size, diameter, and depth range from a few dm to a few m. Potholes are mostly found at or above sea level, in rock formations frequently scoured by waves; however, they may also develop in deeper water, in areas of rough surf (Taboroši and Kázmér 2013). As they may be formed at various levels in the midlittoral or supralittoral zones as well as in a fluvial environment, they should not be used as sea-level indicators.

Honeycombs and Tafoni

Honeycombs are small cavernous features that may occur in a variety of medium- to coarse-grained rocks, assuming cell-like features, with deep chambers separated by thin walls, which may be produced by several physical processes (Fig. 8). They may form in the spray zone above high tide and although they may indicate the proximity of sea level below them, their accuracy is poor (Pirazzoli 2005). Tafoni are typically cavernous weathering features, which are several cubic meters in volume, and have arch-shaped entrances, concave inner walls, overhanging margins (visors) and fairly smooth, gently sloping, debris-covered floors (Cooke et al. 1993). Their cavernous hollows are believed to largely result from granular disintegration and flaking stimulated by a range of possible processes such as hydration, dissolution, and precipitation of calcareous and saline cements, wetting and drying, rain-wash, and wind deflation (Fig. 9). As they may be found not only in coastal outcrops but also in inland deserts, they should be avoided, alone, as sea-level indicators.

Geomorphology and Sea Level, Fig. 8 Honeycombs may be formed by the sea spray. Although they may indicate the proximity of sea level below them, their accuracy is poor since, depending on the wind conditions, they may form far away from sea level

Sea Caves and Arches

Sea caves are cavities excavated by erosion into cliffs by wave action. They usually occur in the weaker parts of rock formations, following soft layers or unconformable strata, or they may have a karstic origin (Pirazzoli 2005). If their floor is regular and flat, it may be related to a palaeoshoreline indicating a former low tide position (Pirazzoli 2007, Fig. 7) (Fig. 10).

A sea arch is a remnant of a cave opening to the sea on both sides of a headland, beneath a connecting bridge. Both features are commonly inaccurate sea-level indicators.

Algal Rims, Trottoirs

Algal rims may be reef-like structures developing on the outer part of reef flats (mainly *Porolithon*) or on the rocky substrates on coasts subjected to regular wave action (Laborel 2005). In subtropical areas, algal rims (mainly *Neogoniolithon*) may contribute, together with the vermetid *Dendropoma* to the development of trottoirs and even to microreefs growing up to sea level. They are usually easy to date and are considered as excellent sea-level indicators (up to ± 0.1 m accuracy) (Pirazzoli 2007).

Trottoirs are crusts of vermetids or calcareous algae that cover a rocky substratum, forming an almost horizontal line at sea level (Pirazzoli 2007). Trottoirs are also considered as excellent sea-level indicators, especially in microtidal areas, and they are relatively common in the Mediterranean and the tropics (Pirazzoli 2007).

Cornices

On the rocky coasts of the northwestern Mediterranean the calcareous alga *Lithophyllum likenoides* (Philippi) forms a



Geomorphology and Sea

Level, Fig. 9 Tafoni, cavernous weathering features, are not found exclusively in the coastal zone, so they should not be considered as sea-level indicators

**Geomorphology and Sea**

Level, Fig. 10 If the floor of a sea cave is regular and flat, the cave may be related to a palaeoshoreline, indicating a former low tide position



solid cornice (trottoir), the upper level of which is slightly above MSL (Morhange et al. 1982) that may be used as an accurate sea-level indicator. Cornices are bioconstructional features, forming a narrow intertidal reef by *Lithophyllum*, *Dendropoma*, coralline algae, and occasionally also by ostraoids (Pirazzoli 2007).

Summary

Past sea level fluctuations may be studied through various types of indicators, such as geomorphological. A number of geomorphological indicators may allow for past relative sea level reconstructions with variable accuracy. Furthermore,

often there are advantages and disadvantages in the geomorphological feature used to unravel the sea level history of a particular area. The combination of different types of indicators is a powerful tool, not only to verify the results of a particular study but also to improve the accuracy of sea level reconstructions.

Cross-References

- Beachrock
- Changing Sea Levels
- Geochronology
- Holocene Coastal Geomorphology
- Paleocoastlines
- Sea-Level Indicators, Biologic
- Sea-Level Indicators, Geomorphic
- Submerged Coasts

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Geotextile Applications

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The technological term “geotextile” has now come to stay in the context of coastal engineering. Perhaps, the original term “filter fabric” is more expressive, but it has fallen into comparative disuse. Geotextiles are really fabric materials used in a “geo,” that is an earth-related, engineering context. These fabrics may be based on natural fibers or man-made synthetic fibers. The latter type of geotextiles are referred to as “geosynthetics.” Geosynthetic materials are generally non-biodegradable, whereas natural geotextiles are often biodegradable. Biodegradability would be an advantage or a disadvantage, depending upon the user needs and use contexts.

Most of the man-made geotextiles (i.e., geosynthetics) are made from polymers of various chemical types. Some common polymers used are polyethylene (PE), polypropylene (PP), polyvinylchloride (PVC), nylons, polystyrene (PS), and polyesters. Natural geotextiles are made from natural fibers, which are also polymers, largely based on cellulose and lignin.

Historical Development

Pioneers of coastal engineering introduced the use of geosynthetics in the 1950s in the United States. According to Bruun (2000), some of the first coastal geotextile research in the United States was carried out at the University of Florida, Gainesville, where plastic filters were tested. Work on the use of nonwoven fabrics for soil reinforcement was done at the Rhone-Poulence textiles in France. Early development work on geotextiles was also conducted in the Netherlands, Britain, Germany, and the United States. Currently, the geotextile industry is quite active in research and development in the United States, through the establishment of the Geosynthetic Research Institute at Drexel University. Additional details on the history of geotextiles can be obtained from Heerten and Kohlhasse (2000).

Material Types and Chemistry

Geotextiles are only a sub-family in the larger family of “Geosynthetics.” Other sub-families include geogrids, geonets, geomembranes, geosynthetic clay liners (GCL), geopipes, and geocomposites. Geotextiles are woven or nonwoven fabrics. Geogrids are plastics formed into open grid-like configuration. Geonets are formed by continuous extrusion of sets of polymeric ribs which when opened give rise to a net-like configuration. Geomembranes are very thin sheets of polymeric materials. GCL’s are thin layers of bentonite clay sandwiched between geotextiles or geonets. Geopipes are simply buried plastic pipes and represent one of the oldest members of the family of geosynthetics. On the other hand, geocomposites are the newest member of the family and are composed of either geotextiles and geonets, or geotextiles and geogrids, or geogrids and geomembranes, etc.

The material used for fabricating geotextiles is invariably some kind of polymer. A “polymer” is a giant molecule of high molecular weight, obtained by polymerizing one or more smaller units called “monomers.” For example, the simple chemical “ethylene” is a monomer. When it is polymerized under suitable conditions, we get “polyethylene” (PE, or polythene). Similarly, the polymerization of simple monomers such as propylene, vinylchloride, or styrene produces PP, PVC, or PS, respectively. In some cases, two monomers are involved in polymerization. Thus, nylons are polyamide polymers formed by the polymerization of diamines and dicarboxylic acids (e.g., hexamethylene diamine and adipic acid). Polyesters such as polyethylene terephthalate (PET) are formed by the polymerization of ethylene glycol and terephthalic acid or derivatives or analogues of these monomers. Polymers commonly used in the geotextile industry include PE, PP, PVC, PET, PS, PA (nylons), and cellulose.

Cellulose is a naturally occurring polymer based on glucose as the monomer.

Another recent development is the use of natural geotextiles, which include fabrics formed using coir, the natural fiber from the husk of coconuts. These geotextiles have exceptional strength due to the high content of a cellulose lignin polymer in them. Besides, they are naturally degradable, which may be attractive in certain applications such as natural stream bank restoration. Other natural fibers include jute, sisal, mixed coir-jute, and mixed sisal-jute.

Geotextile Applications

Barrett (1966), in his well-known and classic paper, discusses the use of fabric materials as reinforcements behind precast concrete seawalls, and under precast erosion control blocks. At present there are several application domains for geotextiles, as discussed below (Heerten and Kohlhasse 2000): (1) filters in erosion control structures (revetments, coastal dikes, etc); (2) separators in the foundation of groins and breakwaters, (3) fabric forms for sand filled bags or tubes as construction elements for groins, dikes, and dunes; (4) flexible scour protection mats at different offshore or coastal structures; (5) reinforcement in dredged material sites; (6) membranes for use in landfill caps; (7) bags for disposing dredged material from navigation channels; and (8) dikes for river training and coastal structures.

Koerner (1998) classifies the applications into six broad categories: (1) *Separation*-is the usage of geotextile to provide physical separation between two material types, while retaining their individual integrity and functions; (2) *Reinforcement*-involves using the geotextile to improve on the inherent strength of the system; (3) *Filtration*-is the use of the geotextile to retain the soil behind it while providing an outlet for dissipation of hydraulic forces; (4) *Drainage*-involves the concept of using the geotextile to provide sufficient fluid movement through it without much loss of the soil behind it; (5) *Containment*-is the concept of providing isolation between two surfaces with the aid of an impervious geotextile; and (6) *Combined uses*-which is the use of the geotextile for any combination of the above listed uses.

Design Considerations

Important general considerations include the following: (1) *Function*-the type of use the geotextile should provide; (2) *Specifications*-the engineering properties that would be needed to satisfy that function; (3) *Availability*-the supply versus demand of that particular type of geotextile in the region; and (4) *Cost*-the price of obtaining and installing the geotextile in place. For small projects with financial

constraints, design by cost and availability is used, which tries to obtain an optimal balance between funds available, area to be covered with geotextile, and unit price of the geotextile.

Key design properties of interest include the following:

- (1) *Physical properties*-includes specific gravity, mass per unit area of the geotextile, thickness, and stiffness;
- (2) *Mechanical properties*-includes compressibility, tensile strength, seam strength, burst strength, and elongation;
- (3) *Hydraulic properties*-includes porosity, percent openings in the geotextile, an equivalent opening size, permeability, and soil retention;
- (4) *Endurance properties*-includes potential for damage during installation, long-term strength loss, abrasion potential, and clogging potential; and
- (5) *Degradation properties*-includes potential for degradation from sunlight, temperature, oxidation, biological action, and hydrolysis.

Various international organizations provide standards for estimating the above properties, with the American Society for Testing and Materials (ASTM 2000) providing the lead. Other leading organizations include the International Standards Organization (ISO), Permanent International Association of Navigation Congresses (PIANC), Geosynthetic Research Institute (GSI) and British Standards Institution (BSI).

Koerner (1998) discusses key aspects of design considerations for the many functional uses of geotextiles, a brief summary of which is presented here

- *Separation design criteria*: Separation function is primal to all geotextiles. Indeed many other functions depend, in their turn, on separation. The use of geotextiles as separators is illustrated by their placement between a soil subgrade (beneath) and a stone base (above). The important criteria here are burst resistance, tensile strength requirements, puncture resistance, and impact resistance.
- *Soil reinforcement design criteria*: Reinforcement of soil is another cardinal application of geotextiles. Seawall reinforcement, reinforcement of embankments, foundation reinforcement and in situ slope reinforcement are the important specific cases. Key properties include tensile strength, puncture resistance, and impact resistance.
- *Filtration design criteria*: Geotextile filters are used behind retention walls, around underdrains, beneath erosion control structures and as silt fences. Adequate fluid permeability and soil retention are the main requisites.
- *Drainage design criteria*: The two main design variations here are gravity drainage design and pressure drainage design. Various manufacturing textiles such as woven silt film, woven nonfilament, nonwoven heatbonded, nonwoven resin-bonded, nonwoven needle-punched, and hybrid systems, are used. Nonwoven needle-punched geotextiles are suited for drainage of fine-grained soil masses such as clays and silt. A high permeability and high soil retention are essential to this application.

- *Containment design criteria*: Application of geotextiles in containment include coastal landfills and other waste barriers. Key properties include low fluid permeability and high soil retention.

Summary

The use of geotextiles in coastal areas has been practiced for a very long time for shore protection needs. However, advances in manufacturing and marketing has led to innovative uses of geotextiles in the coastal zone in recent years. These include the use of sand filled geobags as the core of dunes, sand filled geobag breakwaters and reefs, geobag spur dikes, and deep ocean placement of contaminated sediments in geobags. The future of the geotextile seems limited only by human imagination.

Cross-References

- ▶ [Bioengineered Shore Protection](#)
- ▶ [Capping of Contaminated Coastal Areas](#)
- ▶ [History of Coastal Geomorphology](#)
- ▶ [Navigation Structures](#)
- ▶ [Shore Protection Structures](#)

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Glaciated Coasts

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Definition

Glaciated coasts are dominated or closely affected by glaciers and glacier processes. They include: (1) modern systems in

direct contact with glacial ice; (2) proglacial systems, in front of or adjacent to active glaciers; and (3) paraglacial coasts, formerly glaciated systems that are controlled by glacial geomorphology, sediment sources, and/or isostatic responses.

Introduction

There are a wide variety of glaciated coasts, usually with both erosional and depositional features. Coasts may be actively glaciated systems, such as Antarctica, southeastern Alaska, and Greenland, or formerly glaciated regions, such as Scandinavia, Scotland, northeastern Canada and New England, the Pacific Northwest, and southern Chile (Table 1). Formerly glaciated, or paraglacial coasts (FitzGerald and van Heteren 1999; Slaymaker 2009), are modified by modern coastal processes and by changes during deglaciation such as permafrost action, outwash, isostatic rebound, and proglacial lacustrine and marine deposition and erosion. Many of the morphological features and sediment accumulations are common to both actively glaciated coasts and to paraglacial coasts. In the discussion below they are combined by morphodynamics, rather than separated by active or relict environments.

Glaciated coasts characteristically have moderate to high relief, with abundant bedrock outcrop. They occur in cold-temperate to Polar Regions where modern glaciers and ice sheets are stable, or where perennial ice was stable down to sea level in glacial climates. Formerly glaciated coasts are found as far as from the poles as 40° south and north as a result of the Wisconsin (Würm) Last Glacial Maximum (LGM) 20–22 ka ¹⁴C (Denton and Hughes 1981; Andersen and Borns 1994). Earlier Quaternary glaciations as well as Precambrian and Paleozoic glaciations may have extended even farther from the poles. At present, outlet glaciers from ice sheets meet the coast in Antarctica, Svalbard (Fig. 1), and Greenland, extending to a maximum of 63°N (southern Greenland) and 63°S (Antarctic Peninsula). Valley and piedmont glaciers extend to the sea in Alaska (58°N), Baffin Island (71°N), and Chile (47°S). This variability in latitudes demonstrates that relief, precipitation, ocean currents, and other climatic factors must be considered in evaluating changes in extent of ice sheets.

Study of the former extent of ice sheets and valley glaciers on coasts such as Antarctica, New Zealand, Chile, Alaska, and northeastern Canada provides an important proxy for climate change. Of particular interest are the relationships of events in the northern and southern hemisphere (Moreno et al. 2001), and to potential driving forces for climate change such as the Milankovitch theory of variations in insolation due to Earth orbital parameters. Other theories of climate forcing include dust, atmospheric CO₂, ocean currents, solar variability, internal ice-sheet dynamics, and deglacial freshwater

Glaciated Coasts, Table 1 Glaciated coasts

Morphodynamic types	Examples
I. Active glacial coasts – directly influenced by glaciers	
A. Continental ice-sheet margins	
1. High-relief bedrock and ice cliffs	Much of Antarctica, Greenland
2. Ice shelf	Ross ice shelf, Ronne-Filchner ice shelf
B. Outlet valley glacier margins – tidewater glaciers	
1. Fjords	British Columbia, SE Alaska, Patagonia, Ellesmere Island, Baffin Island
2. Cross-shelf troughs	Pine Island sound Antarctica, Southwest Greenland
C. Proglacial margins – sedimentary coasts actively receiving glacial sediments	
1. Morainal coasts	Malaspina glacier Southeast Alaska
2. Outwash coasts	
(a) Alluvial fan and valley train, Sandur	South-Central Iceland
(b) Deltas	Copper River Delta Alaska
3. Glaciomarine	Svalbard, Southeast Alaska, northern Baffin Bay
II. Paraglacial coasts – formerly glaciated or distant from active glaciers	
A. Glacially eroded bedrock framework	Maine, Atlantic Canada, Sweden
1. Glacially streamlined bedrock peninsulas and embayments of moderate relief (fjord coast)	
2. Relict fjord coasts	Norway, southern Chile, Southwest New Zealand
B. Glacial sedimentary coasts	
1. Terminal moraines	Long Island New York
2. Recessional and minor moraines	
3. Drumlins	
4. Outwash fans	
5. Beach, barrier, and tidal flat coasts built from reworked glacial and glaciomarine sources	New England (see FitzGerald and van Heteren 1999), western Denmark, isles de la Madeleine central gulf of St. Lawrence
C. Isostatic uplift coasts	Maine, Newfoundland, Svalbard, Sweden, Finland
Note that many glacial and paraglacial coasts may be overprinted by isostatic deformation	

Glaciated Coasts, Fig. 1

Coastal outlet glacier with terminal moraine at present sea level, Spitzbergen (June, 1976, courtesy of J.M. Demarest)



flooding into the oceans (e.g., Broecker et al. 1968; Berger 1988; MacAyeal 1993; Denton and Hendy 1994; Broecker 1997; Denton and Hall 2000; Clark et al. 2001). There is a high priority for studies of sensitive areas of rapid change, such as Pine Island glacier and the Ronne Ice Shelf. Hughes (1981) identified Pine Island outlet glacier as the “weak underbelly of the Antarctic Ice Sheet,” suggesting that much of the West Antarctic Ice Sheet could be drawn down through this outlet. Recent observations of rapid change and identification of warm seawater intrusion in the deep trough under Pine Island Glacier ice shelf could be indicators of accelerating losses (Jacobs et al. 2011; Favier et al. 2014). Most of southeast Alaska and Patagonia mid-latitude glaciers are also undergoing rapid retreat and volume loss because of climate change (Larsen et al. 2007), undergoing feedbacks in basal conditions accelerating retreat (Pfeffer 2007). All of these increases of melting and calving of ice have the potential to accelerate global sea-level rise and contribute to complex feedbacks in climate change.

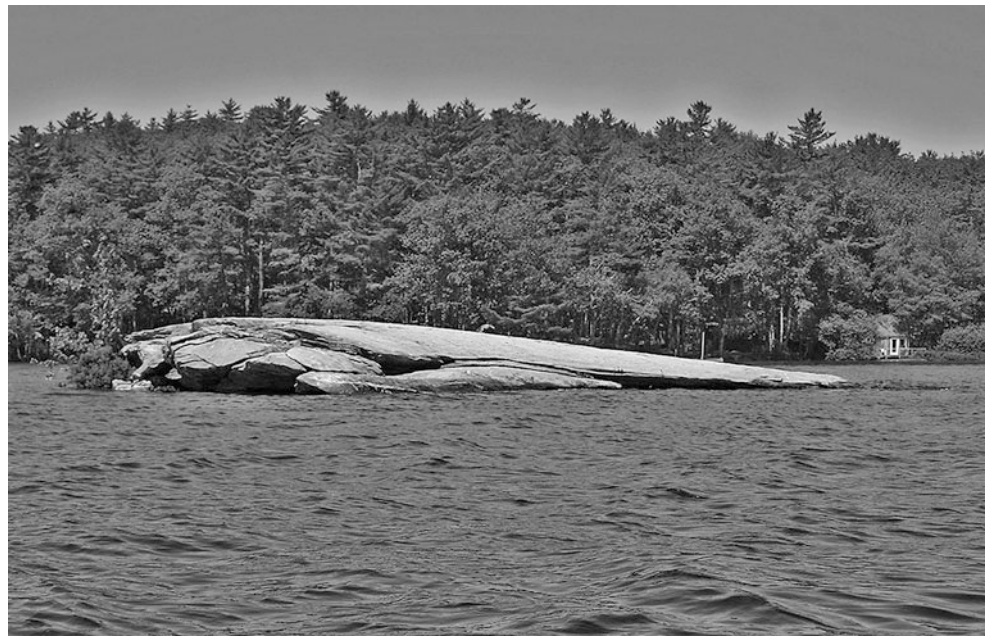
Glaciated Coast Erosional Features

Some characteristic erosional landforms include glacial troughs and fjords. Glacial troughs are immense channels cut by ice streams across the continental shelf in formerly glaciated areas, such as the Laurentian Channel off the Gulf of St. Lawrence and the Northeast Channel off the Gulf of Maine. Other troughs representing greater extent of present ice sheets are found around Antarctic, such as in the Ross Sea and Amundsen Sea. Besides the major ice-stream erosional features, ice sheets also channel flow around resistant rock outcrops and into less resistant valleys, creating characteristic

streamlined hills and linear valleys. Erosional resistance is controlled by rock type (igneous and metamorphic vs. sedimentary), grain size, weathering, spacing and frequency of fractures and bedding planes, and structural orientation relative to ice flow. Small-scale streamlined outcrops, *roches moutonnées*, may form islands (Fig. 2). These and less regular small islands are known as *skerries*. Large-scale streamlining of hills and valleys creates a rolling shoreline of complex peninsulas and embayments, known as *fjord coasts*, as in Maine, Sweden, and other formerly glaciated regions (e.g., Andersen and Borns 1994).

The classic coastal glacial erosion feature is the fjord. Fjords are steep-walled embayments hundreds of meters deep, tens of kilometers long, and kilometers wide. Their mouths are often restricted by a lip composed of bedrock and/or till. This lip may lead to a salinity density stratification in the fjord and also acts a dam that contains abundant proglacial and glaciomarine sediments within the deeper basins. Fjords are formed by valley glaciers, where streaming flow incises more efficiently vertically than laterally and are subsequently drowned by glacial retreat and rising post-glacial sea level. Fjords may be found on actively glaciated coasts, but are also prominent relict geomorphic features. Their morphology is a major control on the physical oceanography and sedimentation in fjords, with or without active glacial input. Sedimentation in fjords is strongly controlled by geometry of the glacial erosion, degree of continuing glacial inflow, and by circulation effects of stratification. Commonly there are coarse deltaic sediments and morainal banks at the head of relict fjords, with fine muddy basins in the remainder of system. Rapid sediment influx creates instability. Slumps, subaqueous fans, and channels created by sediment gravity flows are common (Prior et al. 1981; McCann and Kostaschuk 1987).

Glaciated Coasts, Fig. 2 Roche moutonnée island approximately 15 m long, Pemaquid Pond, Maine, USA, formed on a resistant pegmatite dike intruding schist. Last Wisconsin ice flow was from right to left (175° T). Note the characteristic smooth, striated and polished stoss side (R) and jagged, plucked lee side (L) (D.F. Belknap, 06/19/2013)



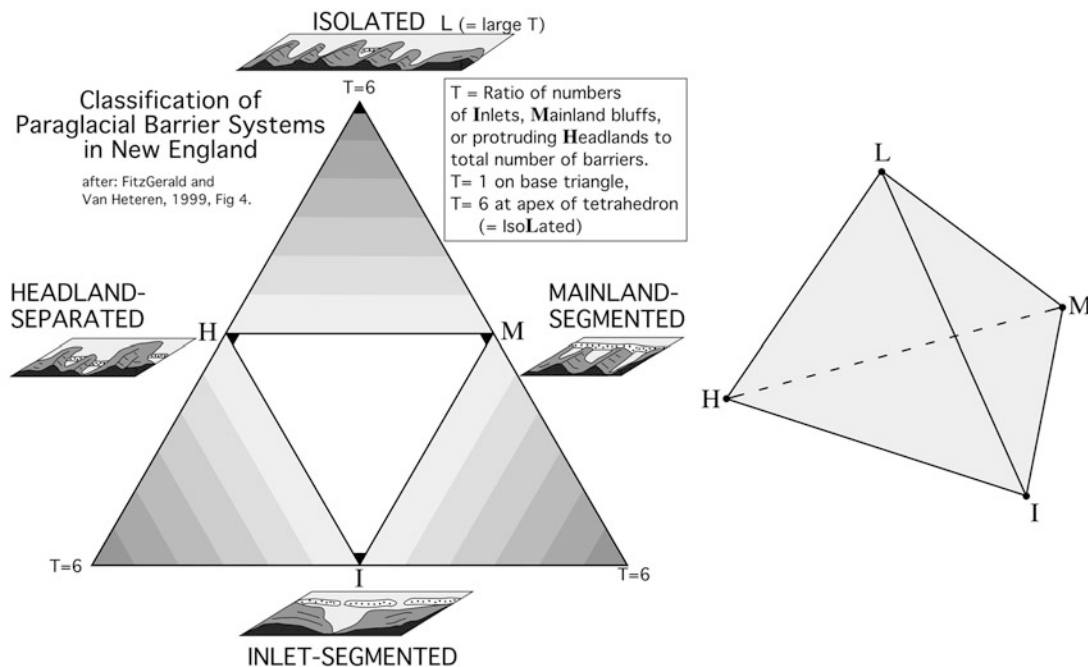
Glacial Depositional Features

Glaciers deposit sediments at their bases, sides, and front. Landforms created by these deposits may form important coastlines on their own, such as Cape Cod, Martha's Vineyard, and Nantucket. Moraines and outwash plains core the outer and inner portions of the well-known "flexed arm" of Cape Cod (Oldale 1992). Reworking of till and stratified outwash from these sources provides abundant sand and gravel to the modern beach and barrier systems (Fisher 1987; Leatherman 1987). Much of the southern New England coast is formed by and related to terminal moraines. The Long Island Atlantic shore demonstrates the transition from a glacial source of eroding bluffs, the Ronkonkoma Moraine at Montauk Point and the Hamptons (FitzGerald et al. 1994), to a wave-dominated mixed-energy barrier-lagoon system southwest of South Hampton.

All of New England can be considered a paraglacial coastal system, with a strong geomorphic signature of glacial erosion and deposition only partially modified by Holocene processes. FitzGerald and van Heteren (1999) classified these paraglacial beach and barrier-lagoon systems of the New England with respect to underlying bedrock topography and glacially derived sediments (Fig. 3). If there is sufficient regional sediment supply, barriers may form a more-or-less continuous barrier, but usually with separated lagoons (M). These barriers may be segmented into separate barriers by inlets (I). If sediment is of less volume or more localized, the barriers may be headland separated (H). The apex of the tetrahedron represents the degree of isolation of the barriers by individual factors or combinations of M, I, and H.

Drumlins are less continuous than moraines, often scattered in fields. Johnson (1919) describes a model for evolution of Boston Harbor and the Massachusetts Bay coast as erosion of drumlins supplied coarse sediment for barriers, until the drumlins were consumed. The shoreline system then jumped back to the succeeding sets of drumlins and their sediment sources. This coastal evolution of drumlin fields is also found in Clew Bay, western Ireland. In settings where bedrock topography is more important for forming embayments, but glacial sediment sources such as drumlins, moraines, eskers, and kames provide the sediment for beaches and barriers, a similar saltating shoreline mechanism may prevail. Boyd et al. (1987) and Forbes et al. (1991) show that the southern coast of Nova Scotia has retreated in this manner during Holocene transgression. Boyd et al. (1987) describe a cyclical model of coarse gravel barrier formation, retreat, and destruction based on the availability of coarse sediment from discrete drumlin sources during shoreline retreat. The Avalon Peninsula of Newfoundland has outstanding examples of drumlins and moraines sourcing baymouth barriers (Forbes et al. 1995) (Fig. 4). Pocket beaches and extended coastlines near these proximal glacial deposits are dominated by boulders, gravel and sand (Kelley et al. 2001). They may form pure gravel beaches, or mixed sand and gravel systems.

Sediment is also brought to the coast as glacial outwash, in valley-train deposits, in deltas and subaqueous outwash fans. These are generally sand and finer gravel but may include gravel and boulders carried by floods and ice rafts. The southern coast of Iceland contains numerous sandurs (outwash fans) that provide the basis for barrier and beach systems (Nummedal et al. 1987). The southeastern coast of Alaska contains a rich variety of glaciated coastline types,



Glaciated Coasts, Fig. 3 Tetrahedral classification of New England paraglacial barrier systems on the basis of separation by headlands (H), and segmentation by mainland (M) and inlets (I) compared to degree of isolation (L), after FitzGerald and van Heteren (1999, Fig. 4)

Glaciated Coasts, Fig. 4

Drumlins sourcing gravel barriers,
Holyrood Bay, Newfoundland
(D.F. Belknap 5/27/2000)



including outwash plains, direct deposition from the Malaspina piedmont glacier (Gustavson and Boothroyd 1987) as moraines that are reworked in the littoral zone, and fjords with retreating valley glaciers (Molnia 1980; Powell 1983). Glaciers supply abundant sediment to deltas such as the Copper River Delta of southeastern Alaska, which has some of the highest natural concentrations of suspended

sediment on earth. The Childs Glacier and Miles Glacier calve directly into the Copper River flowage, while the Sheridan Glacier (Fig. 5) and several others contribute sediment to the margins of the delta.

The finest glacial sediment is rock flour: silt and clay particles carried primarily in suspension. Abundant rock flour causes rapid infilling of lagoons and embayments (up to several

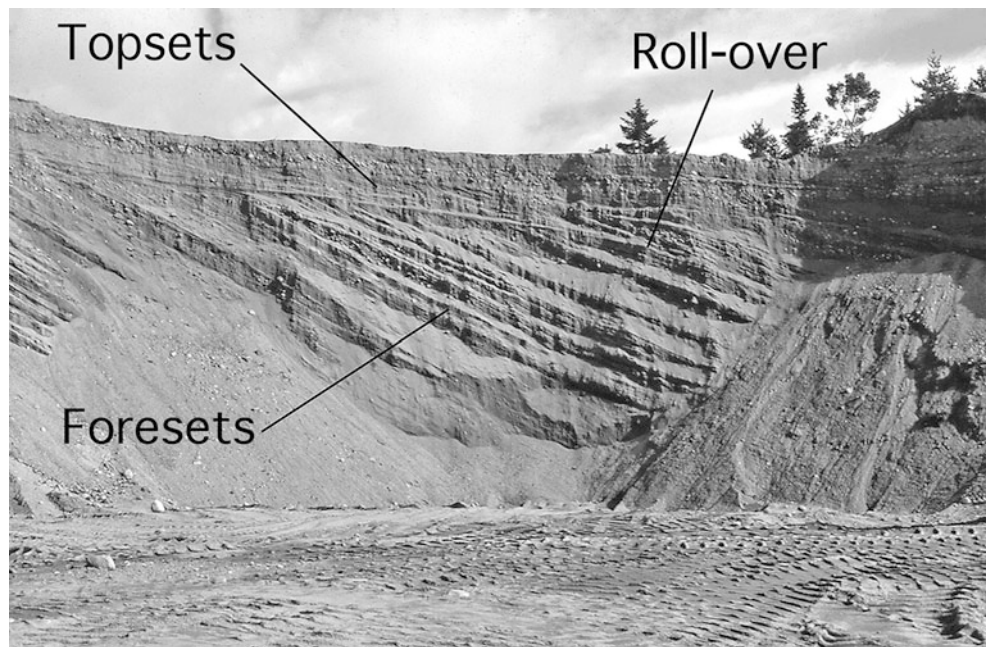
meters/yr) in proglacial and glaciomarine environments. The rapid accumulation of proglacial and glaciomarine fine sediments creates a concentric drape over underlying bedrock or previously deposited glacial sediments, resulting in a distinctive signature in seismic reflection profiles (Piper et al. 1983; Belknap and Shipp 1991). Later slower sedimentation and reworking produces ponded, flat-lying accumulations. Uplift of the glaciated coastline exposes glaciomarine mud, deltas, and outwash fans, as well as moraines and drumlins to wave attack. Much of the coast of Maine, Puget Sound, the Baltic Sea, and similar locations consist of eroding bluffs of these materials. Clay bluffs in particular may be highly unstable, tending to fail as retrogressive slumps.

Glaciomarine deltas form where abundant coarse sediment causes progradation into the sea. On glaciated coasts, especially during deglacial phases, isostatic rebound and relative sea-level change must be in balance for short periods in order to allow deltas, and other shoreline indicators, to stabilize. Glaciomarine deltas are commonly of the classic Gilbert delta type (Fig. 6). Topsets are braided stream deposits at the top of the section. These streams incised and reworked underlying units while the delta was prograding. Foresets are avalanche deposits at the terminus of the traction conveyor, the mouth of the stream. Sometimes it is possible to find the “roll-over” point where the process changed from traction to avalanching, allowing confident reconstruction of shoreline position

Glaciated Coasts, Fig. 5
Sheridan Glacier terminus,
Copper River Delta, Alaska,
(D.F. Belknap 07/18/2009)



Glaciated Coasts, Fig. 6 Gravel pit exposure into a glaciomarine highstand delta, Lamoine, Maine, USA exemplifying the Gilbert Delta stratigraphy. The top of the exposure is approximately 70 m above present sea level (D.F. Belknap 10/12/1982). The topset-foreset contact approximates contemporaneous sea level. The “rollover” shows the preserved delta front in transition from near horizontal topsets to foresets inclined near the angle of repose (30–33°). Bottomsets are not exposed in this pit



and relative sea level at that point (Fig. 5). The third component is bottomsets, the suspension and sediment-gravity flow deposits found basinward of the foresets.

Glaciomarine Environments

There are two major types of glaciomarine environments, tidewater glaciers and ice shelves (Powell 1984). Tidewater glaciers have ice fronts in seawater in restricted embayments, fjords in particular, and produce smaller icebergs. Tidewater glaciers terminate in the ocean at an ice cliff, which is a calving front at or near grounding line (the position where the ice floats out of contact with the bottom) (Fig. 7), either in the open ocean (Pfirman and Solheim 1989) or in protected embayments (Powell 1983). Morainal banks form at the grounding line, and outwash fans form as ice retreats, creating an overlapping and interfingering set of proximal facies. Many Alaskan coastal glaciers, such as in Glacier Bay, are presently retreating rapidly through this process of calving of icebergs. There are several lithofacies found in this environment seaward of the grounding line. Morainal banks are composed of diamicton, gravel, and marine sand and mud. Stratified sand and gravel subaqueous outwash fans interfinger with till and distal glaciomarine mud. These fans may be narrowly restricted in fjords, or broad composites under continental ice-sheet margins (Ashley et al. 1991). Fans also are commonly channeled through traction under density underflows and may be sites of active slumping. Active ice push can deform and override morainal banks and stratified deposits. Iceberg drop and dump deposits are a common admixture to this and more distal environments. Laminated sand and mud are characteristic of basins farther from the grounding line. These are produced from density overflows, interflows and underflows. In the marine

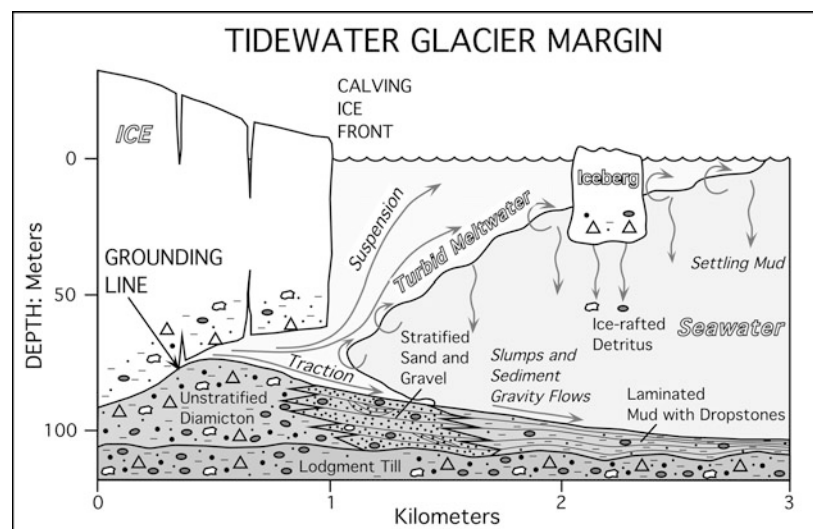
environment, however, the latter are rare, localized close to the source because of the lower density of fresh meltwater, unless charged with very high concentrations of sediment (2–3 gm/l). Laminated deposits are easily characterized as glaciomarine when they contain abundant ice-rafted detritus.

Ice shelves terminate in the ocean, with the ice edge distant from the grounding line. Ice shelves are broad and flat and are restricted primarily to Antarctica today. Ice shelves calve immense tabular icebergs hundreds to thousands of square kilometers in area. Polar ice shelves are cold and do not produce abundant sediment. The Ross Sea and Amundsen Sea, for example, are blanketed by marine biogenic sediments draped over glacial deposits left by Holocene grounding-line retreat (Anderson et al. 1983; Kellogg and Kellogg 1988; Jacobs 1989). Ice shelves that were abundantly productive of sediment were found during deglaciation after the Wisconsin (Würm) Last Glacial Maximum, in particular where fed by ice streams such as in the Scotian Shelf and Gulf of Maine (Bacchus and Belknap 1997; Schnitker et al. 2001).

Glacioisostatic Rebound

The release of the immense weight of ice sheets causes rapid isostatic readjustment. As a consequence, raised beaches can be found preserved around many formerly glaciated regions such as Hudson Bay (Fairbridge and Hillaire-Marcel 1977), Newfoundland (Liverman 1994), Spitsbergen (Boulton et al. 1982), and Scandinavia (Granö and Roto 1989; Andersen and Borns 1994). In Maine, raised beaches formed on the outer edges of glaciomarine deltas (Fig. 8) were raised to elevations 60–130 m above present sea level (Thompson 1982). Recent work with satellite data shows that the

Glaciated Coasts, Fig. 7 Model of a tidewater glacier calving margin, with morainal bank, stratified subaqueous fan, and distal mud and dropstone facies, after Thompson (1982, Fig. 3, p. 215) and Pfirman and Solheim (1989)



Glaciated Coasts, Fig. 8

Highstand shoreline formed ca. 15 ka in sand and gravel at the front of a glaciomarine delta on East Base, 75 m above present sea level (D.F. Belknap 10/12/1982)



southeast Alaskan glacial terrain is undergoing much more rapid uplift than that predicted by large-scale global models (Larsen et al. 2005), up to 32 mm/yr. (3.2 m/century). This has profound influence on coastal systems. Understanding isostasy through study of deformed glacial coastlines in turn has important implications for modeling the deep Earth, such as structure and viscosity of the mantle, and thickness and elastic/plastic properties of the lithosphere. These reconstructions also help constrain poorly understood properties of former ice sheets, in particular their thickness.

Summary

Glaciated coasts are heterogeneous in morphology and sediment type. They occur in higher latitudes today, but the influence of former episodes of glaciation imposes a strong control on coastal evolution through preserved erosional and depositional landforms, and the variety of sediment types. One association that typifies the glaciated coastline is coarse gravel beaches adjacent to muddy embayments. Lack of long-term stability is another important characteristic, driven by isostatic changes, laterally variable sources of fine and coarse sediment, and a complex, deranged landscape upon which coastal environments form and migrate. Glaciated coasts are not only a widespread littoral environment in the modern and during glacial times but study of changing distributions and types can supply critical proxy information for reconstruction of global climate change.

Cross-References

- ▶ [Boulder Barricades](#)
- ▶ [Changing Sea Levels](#)
- ▶ [Climate Patterns in the Coastal Zone](#)
- ▶ [Gravel Barriers](#)
- ▶ [Isostasy](#)
- ▶ [Paraglacial Coasts](#)

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Global Positioning Systems (GPS)

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Synonyms

Global Navigation Satellite System (GNSS)

Definition

Developed in the 1970s by the US Department of Defense (DOD) and managed as a shared global utility by the US Air Force, the Global Positioning System (GPS) is a satellite-based navigation system that allows the determination of precise location, time, and velocity anywhere on earth (US Government, GPS 2017). Coastal scientists often use GPS receivers to collect field data for later analysis and mapping using standard or specialized software including [see in this volume] *Geographic Information Systems* (GIS). The major components of the GPS consist of the space segment, the control segment, and the user segment.

The space segment includes a constellation of radio signal-transmitting satellites complete with computers and atomic clocks. They orbit the earth in 12 h from an altitude of 20,200 km (12,550 miles), sending signals in less than 70 ms (0.06 s) (Trimble Navigation 2017). A full constellation consists of 24 satellites so that at least 4 are visible at any time from any location on the earth; the extras serve as backups (U.S. Government, GPS 2017).

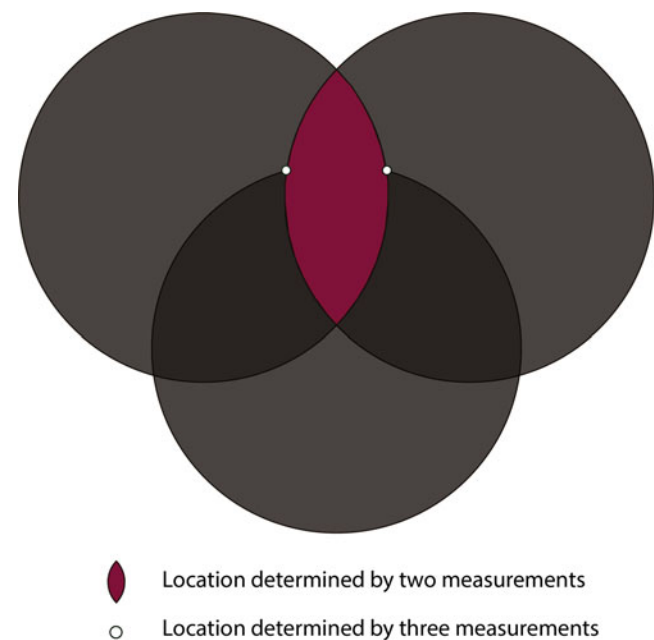
The control segment consists of a master control station with command and control antennas and monitoring sites around the globe (U.S. Government, GPS 2017). Master control at Schriever Air Force Base in Colorado monitors the satellite system and compiles an ephemeris that includes advance calculations of the satellite's orbits; forecasts of the ionospheric and atmospheric conditions are also uploaded to each satellite's computer. The control center additionally monitors the functioning of each satellite and readjusts the orbits or the atomic clocks when necessary.

The user segment consists of the global user community and the billions of receivers in use worldwide (U.S. Government, GPS 2017; Milner 2016). Types of GPS receivers differ from sophisticated surveying models to portable commercial units and now most commonly smartphones. Although GPS receivers vary in their components and thus their capabilities, they will determine location as X, Y, or Z coordinates, typically expressed as longitude, latitude, and altitude. In addition the GPS provides a fourth dimension – time. Each GPS satellite contains multiple atomic clocks that provide very

precise time data via the GPS signals. GPS receivers synchronize to the atomic clocks enabling users to determine the time to within 100 billionths of a second (U.S. Government, GPS 2017).

Satellite Ranging/Calculating Position

A GPS receiver is used to measure distance using the satellite's radio signals. Both the receiver and the satellite generate an identical pseudorandom code that is transmitted simultaneously and allows the determination of signal travel time from the satellites. There are two types of pseudorandom code: C/A code and P code. Civilians use the lower-frequency C/A code and the P code is reserved for the DOD (US Government, GPS 2017; Hurn 1989; Van Sickle 2015). When the receiver picks up a transmission from one satellite, it computes the distance by multiplying the velocity (speed of light) by travel time. The geometric principle of trilateration allows one to accurately measure a position with just three distance measurements (Fig. 1). Calculating the distance from the first satellite narrows the possible location to the intersection of a sphere (with a radius of the distance) and the sphere of the earth. A second distance measurement identifies a second intersecting sphere that further reduces the possible location. A third distance measurement produces a sphere that intersects the others at two possible locations: the points where the third sphere intersects the circle of the first two spheres. In three-dimensional space, trilateration requires a fourth measurement to positively identify one location and its



Global Positioning Systems (GPS), Fig. 1 Trilateration with three distance measurements

altitude. One of the two locations identified with three measurements is often so obviously wrong, however, that a fourth measurement is not needed (Hurn 1989).

Sources of Error

There are several types of error that can influence the accuracy of a typical GPS measurement.

The control segment of GPS provides corrections for some of the errors. Although the satellites carry atomic clocks that keep very accurate time, minimal clock error is still possible and will influence travel time calculations. Additionally, the less accurate receiver clocks may introduce error as well.

Another type of error occurs because radio signals only travel at the speed of light in a vacuum; the ionosphere and troposphere can therefore slow down the GPS signals. The assumption that the signals will travel in a straight line creates another potential for error. Multipath error is caused when a signal is reflected or bounced off of a building, for example, before reaching the receiver (Leick et al. 2015). GPS receivers are capable of processing the signals and determining that the earliest signal to arrive is the best choice.

Because satellites are constantly moving, their orbital location is monitored continually and the data is stored in an ephemeris file. Small orbital errors occur because the satellite location is recorded only every hour (Trimble Navigation 2017).

A geometric dilution of precision (GDOP) error is related to the angle between the satellite and the receiver. Since a wider angle between the satellite and the receiver allows for more precise measurement, good quality receivers analyze the positions of the satellites and choose the best four satellites (Hurn 1989).

Standard horizontal accuracy varies by the quality of the GPS receiver, while vertical accuracy is more difficult because satellite signals are not detected below the visible horizon of the earth (Table 1).

If the task requires accuracy better than that produced by standard GPS methods, differential GPS (DGPS) methods may provide accuracy to within a few centimeters.

Differential GPS and Augmentation Systems

Differential GPS provides a simple solution that corrects for many sources of error and can provide accuracy to a few centimeters for stationary objects, depending on the type of receiver. DGPS relies on two receivers: one receiver is stationary (the base or reference station) and transmits from a known location; the second receiver is the roving device that collects locational data in the field. The receivers both transmit or receive pseudorandom code from the satellites, but, since the base station location is already known, it uses that information to compute the travel time that each satellite signal should take. It then measures the actual travel time (if the satellite signal traveled in a straight line with no interference), and that difference becomes the error correction factor (Steede-Terry 2000).

Sources for differential correction are known as augmentation systems and the options are expanding. Base stations and wide-area differential GPS (WADGPS) services are often established by government agencies, such as the US Coast Guard, the Federal Aeronautics Administration, NASA, and the National Oceanic and Atmospheric Administration, and there is now an international GNSS service as well as other private augmentation services. Their receivers continually compute correction factors as the satellites orbit overhead. The correction factors are encoded in a computer file and used for post-processing. If a user requires “real-time” correction because they are navigating, for example, some receivers are designed to receive radio signals or corrections directly (U.S. Government, GPS 2017).

Conclusion

The US government is continually investing in the modernization of the space and control segments of the GPS in an effort to maintain leadership in satellite navigation. It plans, for example, to implement two to three additional civilian signals that will improve reliability for coastal applications with increased accuracy and availability. Meanwhile other governments have established their own satellite systems including China’s BeiDou, Europe’s Galileo, Russia’s GLONASS, Japan’s QZSS, and India’s IRNSS (U.S. Government, GPS 2017; Leick et al. 2015; Milner 2016). Thus the term GPS no longer describes the totality of the space and control segments of the satellite navigation market; the term Global Navigation Satellite Systems (GNSS) now includes the expansion of systems by other governments and the private sector into the global marketplace. The GPS/GNSS market is estimated to grow by 10% annually from 2016 to 2022 (Research and Markets 2016). Augmentation systems, both government and private, will also continue to expand their range of tools and support to the growing user community (U.S. Government, GPS 2017).

Global Positioning Systems (GPS), Table 1 Source (Trimble Navigation 2017)

Typical error in meters (per satellites)	Standard GPS	Differential GPS
Satellite clocks	1.5	0
Orbit errors	2.5	0
Ionosphere	5	0.4
Troposphere	0.5	0.2
Multipath	0.6	0.6
	10.1	1.2

Despite its continuing importance for determining accurate location and uses in navigation, mapping, tracking, and surveying, supplying very accurate time has increased its usefulness across all sectors (Milner 2016; Trimble Navigation 2017). In studies of dynamic coastal systems, the application of GPS and GNSS technology will keep expanding and evolving as scientists make use of these improvements.

Cross-References

- Beach and Nearshore Instrumentation
- Coastal Boundaries
- Erosion: Historical Analysis and Forecasting
- Geodesy
- Geographic Information Systems (GIS)
- Monitoring Coastal Geomorphology
- Nearshore Geomorphological Mapping
- Photogrammetry
- RADARSAT-2
- Remote Sensing of Coastal Environments
- Remote Sensing: Wetlands Classification
- Shoreline and Coastal Terrain Mapping

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Global Vulnerability Analysis

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Introduction

Climate can have great influence on our lives as shown by the great damage and loss of life in events such as Hurricane

Mitch in Central America, the 1999 cyclone in Orissa, India, and the flooding in Mozambique in 2000. Such events could be intensified by climate change, making this issue a major challenge for the twenty-first century. This widespread concern has generated a global policy response including the Intergovernmental Panel on Climate Change (IPCC) and the United Nations Framework Convention on Climate Change (UNFCCC), whose signatories are committed, among other things, to “avoid dangerous climate change.” The key policy issue is the relative merits of reducing greenhouse gas emissions (usually termed mitigation) and/or adapting to the impacts of climate change, with a mixed response being most realistic.

A major consequence of climate change is global sea-level rise that could cause serious impacts around the world’s coast. In the context of coastal zones, the goal of vulnerability analysis for sea-level rise (and other coastal implications of climate change) is to assess the potential impacts on coastal populations and the related protection systems and coastal resources, including the ability to adapt to these changes. A range of methods for such analyses has been developed and these have been extensively applied at the national and sub-national level (e.g., IPCC CZMS 1992; Klein and Nicholls 1999). These varied studies are often based on different assumptions and scenarios, so they are difficult to synthesize to the larger scales most pertinent to the policy debate outlined above. Therefore, there have also been efforts at vulnerability analysis at the regional and global scale.

The first global vulnerability analysis was completed in 1992 and evaluated: (1) increased flood risk and potential response costs; (2) losses of coastal wetlands; and (3) changes in rice production, assuming a 1-m global rise in sea level (Hoozemans et al. 1992). This was rapidly updated with a second edition (Hoozemans et al. 1993). Here, only results for this second edition are discussed and henceforth this analysis is termed GVA1. The IPCC Common Methodology (IPCC CZMS 1992) was followed throughout. These results influenced the United Nations Conference on Environment and Development (Rio de Janeiro, Brazil, 1992) (IPCC CZMS 1992), and the World Coast Conference (Noordwijk, the Netherlands, 1993) (WCC’93 1994), and are included in the IPCC Second Assessment Report (Bijlsma et al. 1996). Subsequently, Nicholls et al. (1999) made a major improvement relative to GVA1 for the flood and wetland analysis. This was upgraded to a dynamic form, including improved impact algorithms, which can consider variable sea-level rise scenarios, the implications of growing coastal populations, and rising living standards (this analysis is henceforth termed GVA2). It is widely cited within the IPCC Third Assessment Report and has also contributed to a series of impact studies based on common climate and socioeconomic scenarios (e.g., Parry and Livermore 1999).

This following description first outlines the key concepts of global vulnerability assessment. It then presents some selected methods of analysis, considering coastal population and flood risk, adaptation to increased flood risk and wetland loss. The results together with their validation and use are then considered, including the differences between the methods used. Lastly, possible developments in the nearfuture are considered.

Concepts, Constraints, and Approaches

In the present context, *vulnerability* is defined as the degree of capability to cope with the consequences of climate change and sea-level rise (Klein and Nicholls 1999). As such, the concept of vulnerability comprises:

- the susceptibility of a coastal area to the physical and ecological changes imposed by sea-level rise;
- the potential impacts of these natural system changes on the socioeconomic system;
- the capacity to cope with the impacts, including the possibilities to prevent or reduce impacts via adaptation measures. (This last factor is often termed “adaptive capacity”).

However, there are four main barriers to the comprehensive vulnerability assessment, irrespective of the scale of assessment (Nicholls and Mimura 1998):

- incomplete knowledge of the relevant processes affected by sea-level rise and their interactions;
- insufficient data on existing conditions;
- difficulty in developing the local and regional scenarios of future change, including climate change;
- the lack of appropriate analytical methodologies for some impacts.

For the global assessments, the availability of consistent and complete global databases on (1) the distribution, density and present status of the impacted resources and (2) the nature and probability of the impacting hazardous events was a major constraint. Coverage at a global scale was sometimes incomplete due to regional gaps, or only coarse resolution data was available.

All these problems necessitate careful consideration of what can realistically and usefully be assessed. After considering these limitations, global assessments have evaluated fairly simple parameters to date, with the main focus on impact assessment rather than adaptation assessment. GVA1 was limited to an assessment of the potential impacts on three distinct elements in the coastal zone and one possible adaptive response

- *risk to population (and adaptation potential)*—population at risk of flooding and also potential protection upgrade costs;
- *ecosystem loss*—coastal wetlands of international importance at loss;
- *agricultural impacts*—rice production at change (in south, southeast, and east Asia only).

GVA2 refined the methods and results for the first two elements, but the underlying data is the same. Only these elements will be considered here.

In order to assess the vulnerability of a coastal zone to sea-level rise we need uniform procedures to compare and to integrate national and regional studies. To determine impacts in measurable and objective terms, the concept of *values at risk*, *values at loss*, and *values at change* were used. The concept of risk is defined as the consequence of natural hazardous events multiplied by the probability of the occurrence of these events, *excluding* the system response. The concepts of loss and of change are defined as the consequences of natural hazardous events multiplied by the probability of the occurrence of these events, *including* the system response. The “risk”-approach is considered appropriate to assess the consequences of sea-level rise on the probability of episodic hazards such as flood impacts on the coastal population and economy. System response is excluded because it is difficult to predict how flood events may change the behavior of the population in the long term. The concepts of “loss” and “change” are appropriate for impacts on ecosystems and agricultural production, respectively, because it is the longterm consequences of sea-level rise that are the most important factors influencing the magnitude of the impacts.

To examine the capacity to cope with flooding, a set of protection measures was developed with these impact studies to enable the comparison of the impacts of sea-level rise “with-and without increased protection measures.” While one may be susceptible to increased flooding, if one can easily afford to upgrade defenses, there is little cause for concern and overall vulnerability is low.

Lastly, the scale of assessment needs to be considered. Much of the available data for these studies was only available at national resolution and was of uncertain quality. Therefore, some of the underlying data as well as the assumptions about physical processes, physical and socioeconomic boundary conditions, limit the accuracy of the nationalscale results. This is especially true since the last major revision of the underlying databases was in 1993. However, the errors appear to be unbiased so regional and global estimates are expected to be more robust (Nicholls 1995). Therefore, all the results are aggregated to 20 regions and the global scale, and this is the output of the analysis that has been utilized elsewhere. The national data and results are available in

Hoozemans et al. (1993) for reference purposes, although the limits of the accuracy and validity of these detailed results should be born in mind.

Some of the methods are now considered in more detail.

Coastal Population and Flood Risk

Storm surges are temporary extreme sea levels caused by unusual meteorological conditions. The resulting coastal flooding is a major issue damaging livelihoods, causing great distress and in the extreme, loss of life. As many as two million people may have been killed by storm surges in the last 200 years, mainly in south Asia (Nicholls et al. 1995). Sea-level rise will raise the mean water level, and hence allow a given surge to flood to greater depths and penetrate further inland. Changes in storm tracks, frequencies, and intensity would also change surge characteristics, but in the absence of credible scenarios, this factor is considered constant in time within this analysis.

The concept of risk is considered appropriate in the context of assessing the consequences of sea-level rise on flooding for the population in the coastal zone. As rising sea levels intensify flood hazards, some human response might be expected. However, this response is not considered as such a prediction was considered unrealistic given that it involves human choice (ranging from migration to increased protection). Therefore, a high Population at Risk (PaR) indicates the need for some kind of a response. Possible protection costs against flooding are considered in the next section.

Based on the definition of risk, PaR is defined as the product of the population density in a certain risk zone and the probability of a hazardous flooding event in this risk zone. The resulting number is interpreted as the average number of people expected to be subject to flooding events per time unit (/year). Hence, PaR has also been termed “average annual people flooded” (Nicholls et al. 1999). The “risk”-value is able to reflect changes in:

- the population living in the risk zone (coastal flood plain);
- the flood frequency due to sea-level rise;
- the protection standard of defenses.

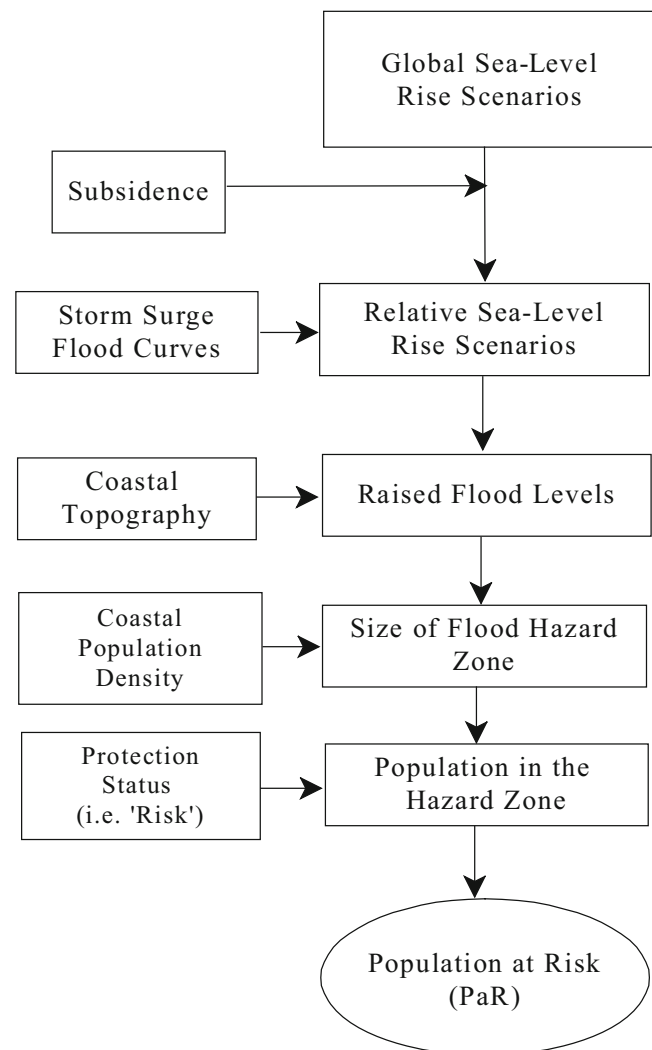
As a general approach, the following steps were undertaken in GVA1 to determine the PaR for the various scenarios:

- Assessment of the height of the maximum flood level theoretically threatening the low-lying coastal zone, taking into account present and possible future extreme *hydraulic and geophysical conditions*
- Determination of the *flood-prone area* and calculation of the area contained between the coastline and the maximum flood level.
- Assessment of the *present state of protection* against flooding.

- Determination of the coastal *population densities* for the present and future state.
- Determination of the PaR with and without measures, with and without sea-level rise, and for conditions in the years, 1990 and 2020.

To estimate global PaR with a reasonable accuracy, the world's coast was divided into 192 coastal zones based on the 181 coastal countries (as existed in the early 1990s). For each of the 192 coastal zones, a database was developed and an identical stepwise calculation scheme was followed to arrive at a coastal zone-specific PaR-number for each scenario (e.g., Fig. 1). The database contained the following elements:

1. the maximum area of the coastal flood plain after sea-level rise;



Global Vulnerability Analysis, Fig. 1 The flood model algorithm as used in GVA2. (Modified from Nicholls 2000)

2. the flood exceedance curve for storm surges from a 1 in 1 year event to a 1 in 1,000 year event;
3. the average coastal population density in 1990;
4. the occurrence or absence of subsidence; and
5. the standard of coastal protection.

Three fundamental assumptions are that (1) the coastal flood plain has a constant slope; (2) the population is distributed uniformly across the coastal zone; and (3) if a sea defense is exceeded by a surge, the entire area behind the sea defense is flooded.

Calculations proceed as shown in Fig. 1. Estimates of four storm surge elevations (1 in 1 year, 1 in 10 years, 1 in 100 years, and 1 in 1,000 years) are raised by the relative sea-level rise scenario and converted to the corresponding land areas threatened by these different probability floods assuming a uniform coastal slope. These areas are then converted to people in the hazard zone using the average population density for the coastal area. Lastly, the standard of protection is used to calculate PaR. These national estimates are then aggregated to regional and global results.

In GVA1, a 1-m rise in sea level was imposed on the 1990 (and 2020) world. Coastal population density was estimated and used directly. In subsiding coastal areas, 15 cm of subsidence was assumed. Protection standards were estimated indirectly as discussed below. Lastly, only impacts of the expansion of the flood plain were considered, although this was amended by Baarse (1995).

In GVA2, a more dynamic approach was followed in which climate and socioeconomic scenarios both reflect realistic timescales. The 1990 coastal population density was increased (or decreased) at twice the rate of national growth. This is simply projecting present trends (Bijlsma et al. 1996). In coastal areas subject to subsidence, a uniform subsidence of 15 cm/century was applied to the entire coastal area, although it is recognized that this is only a first approximation.

There are no global databases on the level of flood protection. Therefore, GVA1 adapted the World Bank classification of less developed, middle and high developed nations to estimate this parameter indirectly and used the GNP/capita in 1989 as an “ability-to-pay” parameter (Table 1). GVA2 used the same concept, but the algorithm was improved to reflect: (1) existing defense standards; (2) the greater costs of protecting deltaic areas against flooding; and (3) the increasing risk of flooding within the coastal flood plain as sea levels rise. The minimum standard of protection in 1990 was assumed to be 1 in 10 years, reflecting that people do not choose to live in highly flood-prone areas. Deltaic areas have a longer land–water interface than elsewhere, and a greater need for water management within the extensive low-lying areas that are protected, substantially raising protection costs.

Global Vulnerability Analysis, Table 1 Protection Classes used in CVA1

GNP/capita (US\$) (or ability-to-pay)	Protection class (PC)	Protection status	Design frequency
<600	PC 1	low	1/1 to 1/10
600–2400	PC 2	medium	1/10 to 1/100
>2400	PC 3	high	1/100 to 1/1000

Global Vulnerability Analysis, Table 2 Revised protection classes used in CVA2, allowing for deltaic and non-deltaic coasts

GNP/capita (US\$)		Protection class (PC)	Protection status	Design frequency
If deltaic coast	If non- deltaic coast			
<2400	<600	PCI	low	1/10
2400–5000	600–2400	PC 2	medium	1/100
>5000	2400–5000	PC 3	high	1/1000
–	>5000	PC 4	very high	1/1000

Based on expert judgment, the protection classes shown in Table 2 were selected. Lastly, the increase in flood risk within the existing flood plain produced by sea-level rise is estimated by reducing the protection class as sea-level rises.

In GVA2, two protection scenarios are considered:

- constant protection (i.e., constant 1990 levels); and
- evolving protection in phase with increasing GNP/capita.

The evolving protection scenario is more realistic based on observed trends during the twentieth century. It should be noted that evolving protection only included measures that would be implemented *without* sea-level rise—that is, there are no proactive adaptation measures to anticipate sea-level rise. These two protection scenarios allow us to examine how such evolving protection might reduce vulnerability to sea-level rise.

Adaptation to Flooding

To estimate realistic first order national-scale protection costs within the constraints of the GVA1, a simple modular approach was adopted (Hoozemans and Hulsbergen 1995). This assumed that the protection class is upgraded by one class (e.g., PC 1 increases to PC 2—Table 1). Then the revised PaR is calculated together with the protection costs as outlined below. The regional protection costs are compared with the regional GNP to quantify their relative cost, and hence the relative capacity to implement such measures.

The method aims to address the wide range of existing coastal defense types and their related costs. The following factors were used:

- the lengths of low-lying coastline (or coastal areas) to be protected,
- a set of six standardized coastal defense measures to be applied,
- standard unit costs for each type of defense measure,
- individual national cost factors, to take account of local cost factors.

For each country, the national costs (CN) were found by applying the following summation at the national scale for all partial stretches of coastline in that country or coastal area which need protection:

$$C_N = \sum l_c \cdot c_m \cdot c_f \quad (1)$$

where l_c is the coastal length, c_m is the unit defense measure cost, and c_f is the national cost factor.

Regionally and globally, the aggregated cost is found by summing the respective national costs. This approach does *not* produce a basis for national coastal defense planning. However, the modular setup provides a realistic and practical framework for subsequent, more detailed analyses, to improve the accuracy and local relevance of the individual modules.

For each coastal country, an evaluation was made of the present types of sea defenses. It was assumed that new or upgraded defenses will be based on this experience, and hence the preferred type of defense options were selected by expert judgment. This selection should also account for matters like soil conditions, elevation and wave-exposure of the shore, the availability of construction materials, and the value of the direct hinterland.

The length of coastline vulnerable to flooding was determined from the earlier World Coast Estimate (WCE) study by adding the length of low coast, the length of city waterfronts, and the length of low coast of islands. Because there were no suitable global databases on geomorphology/defense status (e.g., dunes, dikes, salt marshes) within each country, a typical coastline was selected, based on the dominant coastline type per country.

Six types of defense measure were considered:

1. *Stone-protected sea dike*
2. *Clay-covered sea dike*
3. *Sand dune*
4. *Tourist beach maintenance (beach nourishment);*
5. *Harbors and industrial areas (upgrade);*
6. *Elevation of low-lying small islands.*

Defense measures 1, 2, and 3 apply in most cases. However, the effective implementation of any measures requires a well-functioning technical and organizational infrastructure

(as this is a key element of adaptive capacity). More demanding coastal defense works like those used to close off large estuaries were deliberately omitted from the standard list, although their application will be the most economical solution in some cases, as experience in England, the Netherlands, and Japan illustrates.

Cost estimates for the standard protection measures were established in 1990 US dollars. They are based on the following assumptions and conditions, which draw strongly on Dutch technical experience:

- Standard defense constructions are defined for each situation, including dimensions, construction material, and construction methods.
- The schematized designs are based on well-established procedures.
- Standard unit costs are derived from the Dutch situation, including provisions for all the costs, including design, execution, taxes, etc.
- Construction methods and cost estimates are based on the assumption of construction in one continuous operation per project.
- The hydraulic regime determined in the flood analysis is considered in design, increasing adaptation costs in areas with large surges.

Coastal Wetlands and Loss

Natural systems may also be impacted by sea-level rise. Coastal wetlands are defined as salt marshes, mangroves, and associated unvegetated intertidal areas (and here exclude features such as coral reefs and shallow-water sea grasses). Wetlands are not impacted by short-term fluctuations in sea level such as tides and surges, but they are susceptible to long-term sea-level rise and show a dynamic and nonlinear response (Nicholls et al. 1999). Therefore, we are considering potential losses. In this case, it is important to consider the wetland response to sea-level rise to make credible impact estimates.

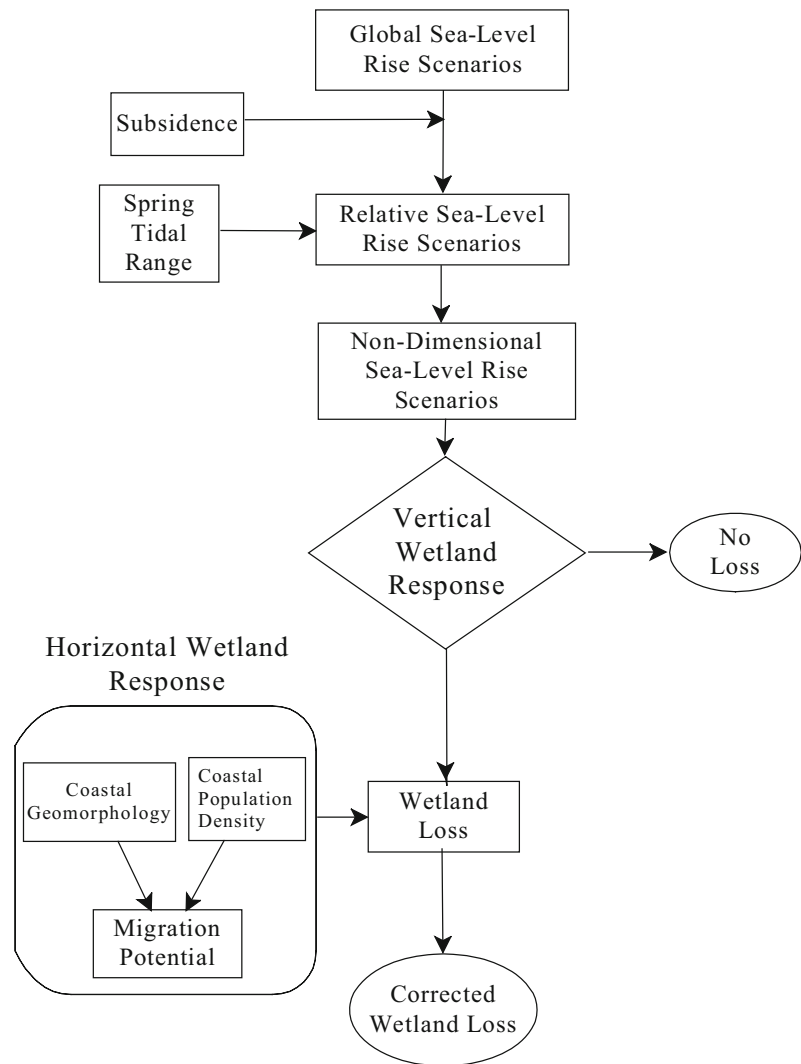
All the evidence shows that coastal areas with a small tidal range are more susceptible than similar areas with a large tidal range. Loss of coastal wetlands due to sea-level rise can be offset by inland wetland migration (upland conversion to wetland). However, in coastal areas without suitable low-lying areas, or in low-lying areas protected against flooding, wetland migration cannot occur, producing a coastal squeeze.

A database of the type, area, and location of most coastal wetlands of international importance was created as part of GVA1. It mainly comprised sites recognized by the Ramsar Treaty. It is missing data for certain regions such as Canada, the Gulf States, and the former Soviet Union.

GVA1 identified those wetlands that might be threatened by a 1-m rise in sea level, but did not project actual

Global Vulnerability Analysis,

Fig. 2 The wetland loss model algorithm used in GVA2. (Modified from Nicholls et al. 1999)



losses. These threatened wetlands were identified based on coastal geomorphology, tidal range, and local population density.

A nonlinear model of coastal wetland response to sea-level rise was developed in GVA2 (Fig. 2). The modeling effort is split into two parts (1) vertical accretion and (2) wetland migration. To model vertical accretion, the availability of sediment/biomass for vertical accretion is parameterized using critical values of non-dimensional relative sea-level rise ($RSLR^*$):

$$RSLR^* = RSLR/TR \quad (2)$$

where $RSLR$ is the relative sea-level rise scenario and TR is the tidal range on spring tides. Hence, wetlands in areas with a low tidal range are more susceptible to sea-level rise than wetlands in areas with a higher tidal range. (The rate of relative sea-level rise was implicit being defined by the 95-year period of interest). A critical value of $RSLR^*$.

($RSLR^*_{crit}$) distinguishes two distinct wetland responses to sea-level rise in terms of vertical accretion:

1. $RSLR^* \leq RSLR^*_{crit}$, no wetland loss as wetland accretion $\geq$ sea-level rise; and
2. $RSLR^* > RSLR^*_{crit}$, partial or total wetland loss as wetland accretion $<$ sea-level rise.

If wetland loss occurs, it is modeled linearly using the excess sea-level rise up to $RSLR^* = RSLR^*_{crit} + 1$. Above this rise, (near-) total loss is assumed and wetlands will only survive if there is inland wetland migration. This simple model captures the nonlinear response of wetland systems to sea-level rise and the association of increasing tidal range with lower susceptibility to loss. The literature stresses the large uncertainties concerning quantitative wetland response to sea-level rise, so a range of values for $RSLR^*_{crit}$ which encompasses the available information were selected (Nicholls et al. 1999). The wetland sites are aggregated to

the 192 coastal areas defined in the flood analysis, except for eight continuous national coasts that were subdivided because of a large variation in tidal range.

To model wetland migration, the same approach as GVA1 was used. The natural potential for the migration of the coastal wetlands under sea-level rise was evaluated for each wetland site using the global coastal geomorphic map of Valentin (1954) (showing the limited work on global-scale coastal typology in the last 40 years!). Three classes of migration behavior are recognized: (1) migration is possible; (2) migration is impossible; and (3) migration is uncertain. In the latter cases, losses were calculated assuming both migration and no migration and this contributes to the uncertainty between the low and high range of the results. In areas where migration is possible, the population density was estimated for the 2080s in a consistent manner to the flood analysis. If this population density exceeded 10 inhabitants/km², it was assumed that wetland migration would be prevented by flood protection structures. In areas where wetland migration is possible, wetland losses are assumed to be zero (i.e. wetland migration compensates for any losses due to inundation).

Validation/Interpretation of the GVA

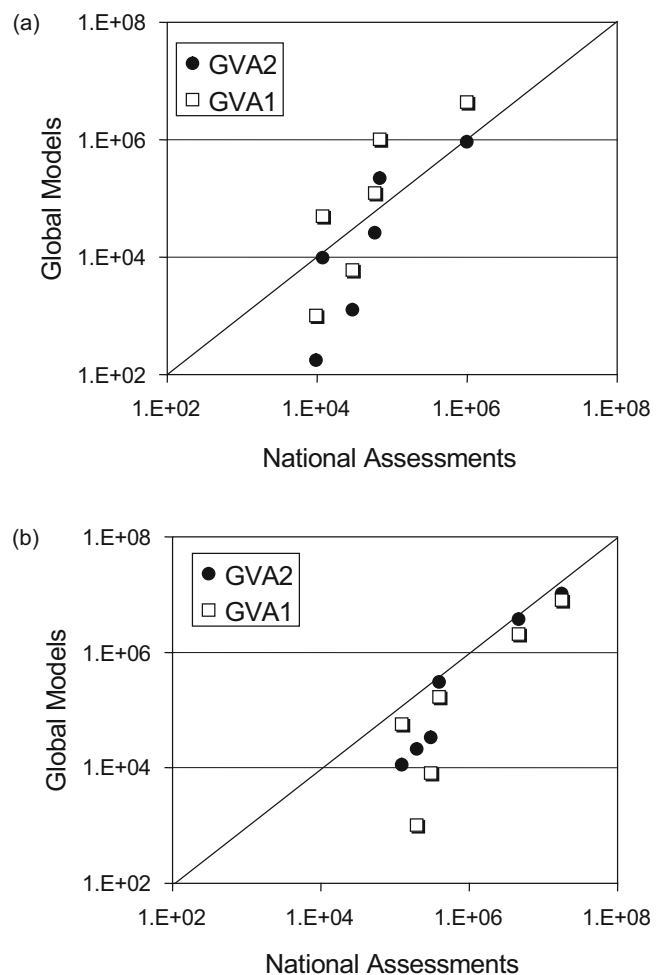
Validation of any model is a critical step, which increases confidence in the absolute quality and interpretation of the results. However, global assessments can also be interpreted as relative impact measures. Given that global assessments provide internally consistent results, the relative impacts may be more reliable than the absolute impacts. Therefore, it is suggested that users interpret the results from both an absolute and a relative perspective.

For the flood analysis, an independent data set of the impact parameters was developed via national-scale vulnerability assessments (Nicholls 1995, 2000). While these national-scale results consider the impacts of sea-level rise on the 1990 world without any socioeconomic changes, the results can be used to validate the global flood model for these scenarios. In broad terms, the results for GVA2 are of the right order of magnitude for the three parameters assessed, and are also an improvement over the results in GVA1 (e.g. Figure 3). Therefore, the changes to the methods in GVA2 are justified.

Validation of the protection cost estimates suggested that they are broadly reasonable (Nicholls 1995). The validation of the wetland loss models remains limited due to a lack of suitable calibration data.

Results

Globally, about 200 million people live in the coastal flood plain (below the 1 in 1,000 year flood elevation). GVA1 estimated that PaR is 50 million people/year in 1990. Most of these people live in deltaic areas in the developing world.

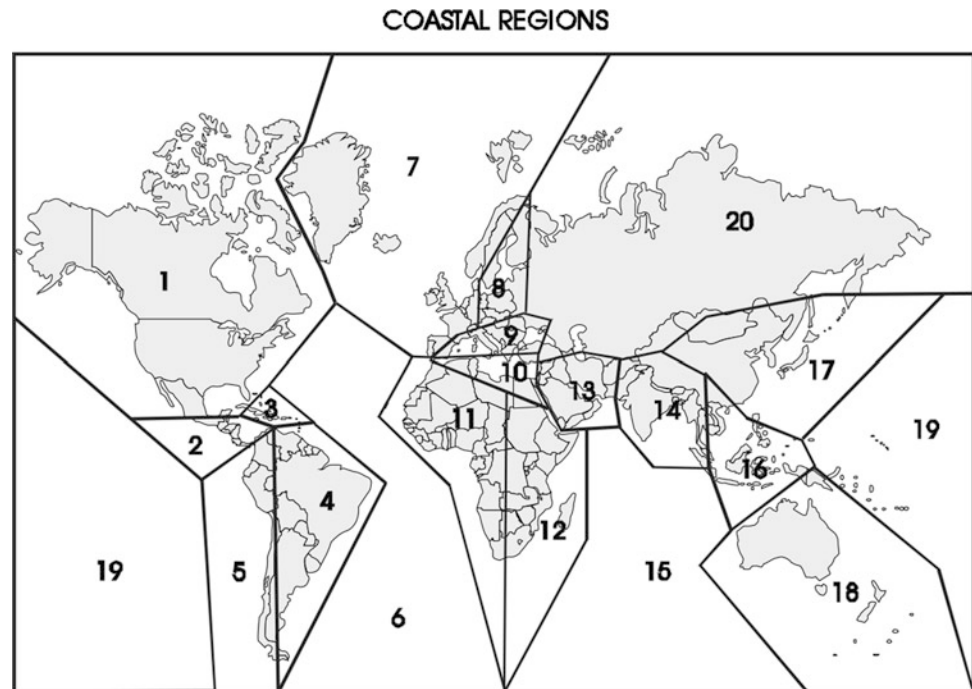


Global Vulnerability Analysis, Fig. 3 National PaR-estimates from GVA1 and GVA2 against independent national-scale vulnerability assessments for six countries (Egypt, Germany, Guyana, the Netherlands, Poland, and Vietnam): (a) no sea-level rise in 1990, and (b) 1 m sea-level rise in 1990. (Modified from Nicholls et al. 1999)

The expansion of the flood plain due to a 1 m sea-level rise will increase PaR to 60 million people/year based on 1990 population. Allowing for the additional factor of increased flood frequency within the existing coastal flood plain, PaR doubles to 120 million people/year (Baarse 1995). Upgrading coastal protection infrastructure against a 1 m rise in sea level as outlined above could collectively cost US \$1,000 billion (1990 dollars), or 5.6% of the 1990 Global World Income. In this case, PaR is reduced from 60 to 7 million people/year, so there would be substantial benefits to coastal inhabitants. As these cost estimates assume an instantaneous rather than a progressive response and do not consider erosion in non-tourist areas or the costs of water management and drainage, they are more useful as a relative cost rather than an absolute adaptation cost.

Coastal wetlands are already declining at 1%/year to indirect and direct human activities. They would decline further

Global Vulnerability Analysis,
Fig. 4 The 20 regions used in the
 GVA. (From Hoozemans et al.
 1993)



due to a 1 m rise in sea level: more than half of the world's coastal wetlands could be lost.

The coastal regions defined in Fig. 4 have different problems in the GVA1. Six regions are vulnerable to the loss of coastal wetlands: North America; Central America; South America Atlantic Coast; north and west Europe; northern Mediterranean; and Pacific large islands (GVA regions 1, 2, 4, 7, 9, and 18). Nine regions were considered to be vulnerable with respect to both flood impacts/response costs and loss of coastal wetlands: the Caribbean islands; west Africa; the Indian Ocean small islands; east Asia; the Pacific small islands; the southern Mediterranean; south Asia; southeast Asia; and east Africa (GVA regions 3, 11, 15, 17, 19, 10, 14, 16, and 12). Relative flood impacts are significant for small island settings, but the absolute impacts are highest in south, southeast, and east Asia.

The improved and validated approach of GVA2 suggests that under the 1990 situation, PaR is 10 million people/year. Even without sea-level rise, PaR is likely to increase to the 2050s due to increasing coastal populations. A 1 m rise in sea level produces a 14-fold increase in PaR given the 1990 world, rather than a threefold increase as found by Baarse (1995). Therefore, in the absence of adaptation, the impacts of sea-level rise on flooding are much more dramatic than previously realized. Evolving protection reduces the magnitude of flood impacts, but the relative increase in people flooded given sea-level rise is still dramatic. Under a lower sea-level rise scenario of 38 cm by the 2080s, the global increase in flooding will be seven-fold compared with the situation without sea-level rise. Most of these people will be flooded so

frequently that some response seems inevitable. The most vulnerable regions are similar to GVA1, comprising large relative increases in the small island regions of the Caribbean, Indian Ocean, and Pacific Ocean small islands (GVA regions 3, 15, and 19), and large absolute increases in the southern Mediterranean, west Africa, east Africa, south Asia, and southeast Asia (GVA regions 10, 11, 12, 14, and 16).

Wetland losses given a 1 m rise in sea level could approach 46% of the present stock. Taking a 38 cm global scenario by the 2080s, between 6% and 22% of the world's wetlands could be lost due to sea-level rise. When added to existing trends of indirect and direct human destruction, the net effect could be the loss of 36–70% of the world's coastal wetlands, or an area of up to 210,000 km². Therefore, sea-level rise is a significant additional stress which makes the prognosis for wetlands even more adverse than existing trends. Regional losses would be most severe on the Atlantic coast of North and Central America, the Caribbean, the Mediterranean, and the Baltic. While there is no data, by implication, all small island regions are also threatened due to their low tidal range.

The major change in flood impacts from GVA1 to GVA2 reflects the more realistic assumptions in GVA2 concerning the present protection status, and its degradation as sea-level rises. The most important effect on PaR is the increased risk of flooding in the existing flood plain, rather than the expansion in the size of the flood plain as sea levels rise. Wetland losses in GVA1 and GVA2 are difficult to compare, but results for the common 1 m scenario are similar. The main benefit of GVA2 is its more flexible form.

What Have We Learned/Next Steps?

The analyses described here have proven the concept and utility of global vulnerability assessment for policy analysis. The results confirm that global sea-level rise could have a range of serious impacts on the world's coasts if we fail to plan for these changes. Further, these impacts will be greater in some regions than others with parts of Asia, Africa, and small island regions most adversely impacted. This is an important result to be considered by the UNFCCC policy process. However, what to do is not evaluated by the existing analyses.

In scientific terms, the rigor of developing such models gives improved insights into the functioning of the integrated (i.e., coupled natural-human) coastal system from the local to the global scales. For instance, this work explicitly considers the relationship between local wetland accretion and elevation change rates and potential regional and global wetland losses. It also explores how increased protection of human activities in low-lying areas may exacerbate wetland vulnerability under a scenario of rising sea level. Continued research efforts will provide important insights that will be useful to research programs such as the International Geosphere–Biosphere Programme, Land–Ocean Interactions in the Coastal Zone Project (IGBP-LOICZ) (Holligan and de Boois 1993) and more generally improve the scientific basis of longterm coastal management.

Policy analysis for climate change requires flexible tools, which can link different emission scenarios all the way to potential impacts and adaptation potential. This will allow exploration of a wide range of sea-level rise (and other climate change) and socioeconomic scenarios, including different mixtures of mitigation and adaptation options. The experience with developing GVA1, and its improvement to GVA2 provide the basis to develop such tools. Important needs for future models include operation at a finer resolution, a better description of impact processes, and the facility to include different response and adaptation pathways, among other improvements. A European Union research project called Dynamic and Interactive Assessment of National, Regional, and Global Vulnerability of Coastal Zones to Climate Change and Sea-level rise (or DINAS-COAST) is exploring these issues and will report in 2004.

Cross-References

- [Changing Sea Levels](#)
- [Coastal Subsidence](#)
- [Deltas](#)
- [Demography of Coastal Populations](#)
- [Dikes](#)
- [Holocene Coastal Geomorphology](#)

- [Natural Hazards](#)
- [Sea-Level and Climate Change](#)
- [Small Islands](#)
- [Wetlands](#)

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Gravel Barriers

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Definition

Gravel barriers are morpho-sedimentary units comprising beach face and rear landward-sloping unit whose surface sediment size $\geq 3\phi$. Surface sediment may be swash sorted by particle size and shape to form zoned facies. Internal structure shows matrix-supported as well as open-worked clast-supported stratification. Thus fine sediment ($\leq 0\phi$) is an element of most gravel barriers. The type concept of sand, mixed sand and gravel, and pure gravel barriers (Kirk 1980) is simplistic. Gravel barriers comprise sediment mixtures, the surface is but one coarse-clastic dominated facies set, as a function of incident wave energy, swash dynamics, and tidal range. Carter and Orford (1993) consider "coarse-clastic" a more appropriate descriptive term than "gravel."

Location and Type

Most gravel (including shingle) literature concerns mid-to high-latitude barriers (mainly northern latitudes) associated with paraglacial coasts. These barriers tend to be transgressive and sediment depleted (Orford et al. 1991). There are also gravel barriers, predominantly regressive and sediment abundant, in southern latitudes formed from mega-river supply or from reworked fluvial gravel fans associated with mountains. Gravel accumulations constructed from storm-generated coral reef debris can occur on tropical atolls.

Morphology

Transgressive barriers show a classic barrier morphology: narrow (<100 m), elongated longshore, and comprising one ridge crest up to 14 m high above mean sea-level. Ridge height is a function of sediment shape and size and breaking wave height. Breaches in transgressive barriers are usually initiated by cross-barrier hydraulic gradients, forced either landward by storm surges or seaward by terrestrial flood impoundment. A reduction in the fine-sediment component is likely in transgressive barriers. Regressive barriers show multiple beach ridges often with seaward trends in ridge elevation. Internal sediment facies variation reflects the importance of cusate and berm

morphology superimposed on prograded ridge sedimentation. What controls multiple ridge development is still widely debated: extreme storms, storminess phases, extreme tidal elevation, sediment supply and textural mix, and stochastic processes.

Essential Concepts

Relative sea-level (RSL) change sets paraglacial barrier development tempo. Forbes et al. (1995) argued that longshore supply is controlled by RSL rise. Jennings et al. (1998) suggest that RSL rise controls types of gravel barrier development: spatial stability exists between 2 and 8 m a⁻¹ RSL rise. Below 2 mm a⁻¹, RSL rise is too slow for sediment replenishment and allows longshore reworking of a barrier. Above 8 mm a⁻¹, the high-tide shoreline is moving too fast for barrier concentration.

Orford et al. (1995) identified barrier retreat rate as 1 m a for each millimeter per annum RSL rise.

Barriers are defined as drift or swash-aligned, based on sediment supply. Paraglacial barriers related to till-covered basements are often fed from supply-limited sources of sediment (e.g., drumlins). Early to middle stages of drumlin erosion (high longshore supply) lead to associated longshore drift-aligned barriers (forced regression). As sediment supply reduces, the barrier becomes swash-aligned with both longshore cannibalization of shoreline barrier sediment and onshore movement of remnant sediment by overtop and overwash. Spatial fluctuations in sediment supply are not as important for barriers fed from unlimited supply sources.

Gravel barrier development is related more to storm wave than fair-weather conditions. Beach crest height is proportional to past breaker heights recorded at high water, hence the storm connection. Extreme runoff may flow over the crest top, transporting gravel to the crest (=over-top). Overtopping builds up the crest elevation, reducing the rate of subsequent overtopping and preventing barrier retreat. Extreme volumes of overtopping water can wash sediment down the barrier back-slope (=overwash). This lowers the crest height and allows smaller breaking waves to overwash. Consistent overwashing leads to landward migration of the gravel barrier. Quiescence and storminess periods may regulate short-term barrier retreat rates, but RSL rise controls long-term retreat.

Basement structure defines the barrier's accommodation space. When sediment supply is scarce then a crenulate coastal configuration is essential to define both shoreline hinge points and offshore bathymetric variation that causes refraction of incident wave by which barriers can accumulate sediment. Crenulate coastlines help transform drift-aligned gravel spits into swash-aligned gravel barriers as supply dwindles.

Conclusion

Swash-aligned barriers cannot rollover indefinitely so barrier breakdown will occur (Orford et al. 1996). Sediment may come from rollover incorporation but this is outweighed by sediment loss to the barrier's extending length. Where sea level is rising the reducing return period associated with extreme water levels means that the barrier is more liable to overwashing and multiple phases of barrier-breaching. Wave refraction around the breaches forces the barrier remnants to fall back on themselves, with sediment moving onshore along transverse shoal lines to act as a source for new landward barrier construction. It is possible that swash-aligned barriers in rollover mode may be "scalped" in extreme storms (=ver-stepped). The upper barrier is swept onshore to form a new barrier leaving the old basement as a relict feature on the foreshore (Forbes et al. 1991). The twentieth century has seen a trend for fixing gravel barriers by artificially maintaining the crests, allowing urban development in their lee. The coastal squeeze is very pronounced behind such modified barriers with no space for rollover. The high likelihood of catastrophic gravel barrier failure is a salient problem for twenty-first century coastal zone management.

Cross-References

- Beach Sediment Characteristics
- Changing Sea Levels
- Drift and Swash Alignments
- Gravel Beaches
- Paraglacial Coasts
- Reflective Beaches
- Sediment Budget

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Gravel Beaches

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Definition

Beach is the sediment surface that extends seaward to the zone of intense wave disturbance of sediment particles and landward to the zone of change in physiographic form or start of vegetation.

Exfiltration is the short-term drainage of beach groundwater that occurs following wave run-up on the beach face and before the succeeding wave advance causes recharge of the beach groundwater.

Gravel is a size designation of rock fragments that ranges from 5 to 75 mm, according to a commonly accepted classification system. The size is measured along the particle's intermediate axis.

Iribarren number, also known as surf similarity parameter or breaker parameter, is a numerical ratio of beach face steepness to wave steepness. The value of the ratio is related to wave behavior upon encountering the beach, such as the type of breaking, the extent of run-up, and the amount of reflection.

Morphodynamics is the study of changes to a landscape (in the present case, the nearshore) caused by externally applied forces (waves and currents) and through mutual adjustment among features of the nearshore system.

Overwash is the transport of water contained in a broken wave over the crest of the beach or barrier and is the agent for transporting sediment and drift debris onto the beach crest, onto the upland, or into a lagoon behind the barrier.

Shingle is beach material in the size range of gravel, with all particles having smooth sides and rounded edges. The interstitial spaces of shingle are not filled with finer sediments.

Wave swash/backwash is the flow of water up and down a beach face, respectively, resulting from wave breaking. Each wave crest breaking at the shore launches a pulse of water up the beach face. The water returns down the beach face under the force of gravity between arriving wave crests.

Introduction

Gravel beaches are accumulations of shore material formed into distinctive shapes by waves and currents which contain lithic particles in the gravel or shingle size range (a median diameter between 5 and 75 mm). The beach form is generally a seaward-sloping boundary between mobile sediment and a water body and contains a flat or landward-sloping surface at the upper limit of the beach. The term gravel beach typically includes shingle beach (nearly uni-size sediment) and may be used to refer to mixed sand and gravel (MSG or mixed) beaches as well. Gravel beaches are common in high latitudes where coastal sediment has a glacial transport history. They also occur at other latitudes, especially near eroding cliffs and mouths of steep river drainages. The value of gravel beaches is increasingly recognized as substrate for certain biology and as environmentally compliant shore protection by shoreline managers.

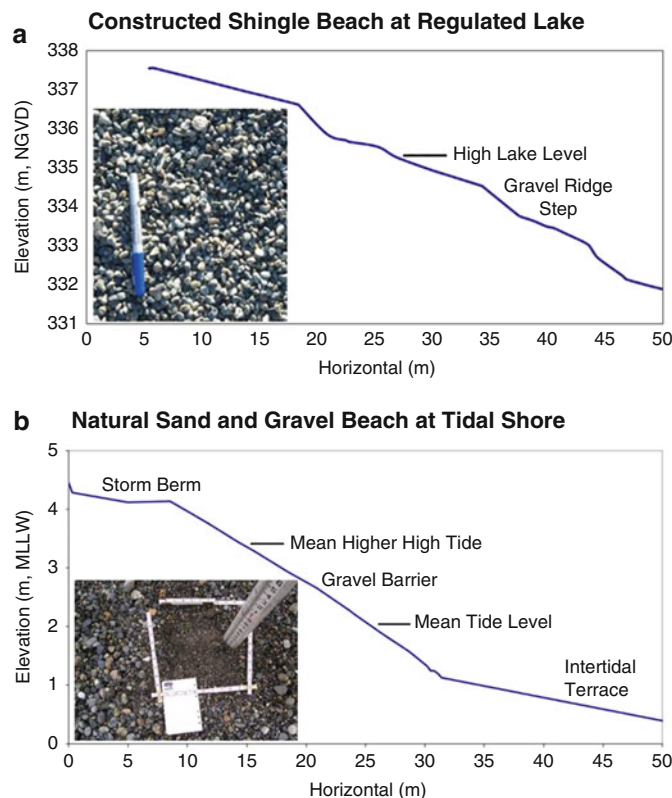
Characteristics

The gravel beach shape in vertical profile is influenced by relative amounts of gravel and finer material supplied to the nearshore and reworked by beach processes. Figure 1 illustrates beach features to be discussed. The height of the storm berm crest is roughly the upper limit of wave run-up in storms

with frequent occurrence. The berm crest height is approximately 0.5 m above mean higher high water in moderate wave energy exposures and could be higher in more energetic settings. One or more gravel ridges may exist in the subaerial profile of a shingle beach. A bar is rarely present on the submerged profile owing to weak offshore directed coarse sediment transport.

The composition of the gravelly surface results from winnowing of finer sediments from the original material, which can occur along the entire transport path. A shingle beach contains no or negligible amounts of sand and finer material. Chesil Beach in Dorset, UK, stands out as a shingle beach with a long history of study of processes relating wave energy to particle sorting and profile characteristics.

The relative amount of fines and the degree of wave sorting divide MSG beaches into two categories. One is a broadly homogeneous mixture of sand and gravel, with some grading across shore and alongshore. The distinctly sandy region, if present, is usually exposed only during low spring tides. Frequently, sandy materials underlie surficial coarse sediments. The second category of mixed beach is a composite beach, characterized by a wide, sandy, intertidal terrace flanked by a gravel ridge, sometimes called the high tide beach. In this second beach category, an abrupt change in beach face slope marks the division between the two sediment types. Slope of the gravel ridge commonly ranges from 6 to



Gravel Beaches, Fig. 1 Idealized beach profile features for (a) shingle beach and (b) mixed sand and gravel beach

10°. Both types of mixed beaches may have complex stratigraphy. The surface layer of pebbles and cobbles is often underlain by coarse particles of mixed sizes supported in a sandy matrix.

The mixture of sediment sizes determines the beach permeability to wave swash and thus the asymmetry between wave uprush and backwash and the instantaneous water table within the beach face. As water from the advancing wave percolates into the beach face, the water table locally is raised and the volume of water available for transporting material seaward during backwash is diminished.

The variance in sediment size with depth has great influence on swash infiltration. This in turn controls swash hydrodynamics and groundwater flows and the beach profile response to waves. At the point where the volume of sand as a percent of total beach sediment exceeds a threshold, porosity, reflectivity, and wave run-up relationships to incident waves (among other parameters) change in a nonlinear manner in response to each other. Experiments with beaches composed of different sizes and size distributions of sediment show that permeability is controlled by the finest 10% of sediment within the mix and that permeability is the dominant factor in profile evolution under wave attack.

Shingle beach profiles develop an abrupt step located just seaward of the shoreline near the wave break point. The step is observed in the laboratory and the field. Bores develop and collapse immediately after waves break over the shallow face of the step. The step's initiation and maintenance are related to wave height and period and so are thought to force wave breaking. Asymmetrical incident bores interact with supercritical backwash flow to create a vortex, which could be responsible for forming the step. The vortex is most energetic when wave period is resonant with swash duration. The coincidence of sediment mobilization by turbulence generated through wave breaking and sediment advection up the slope by the collapsing bore is responsible for distributing and maintaining sediment sizes across the swash zone. Very coarse grains are found at the step, and sizes are selectively moved upslope to maintain the characteristic steep profile.

Beach reflectivity changes rapidly as the water level changes around the elevation of the break in slope between the intertidal terrace and gravel ridge of mixed beaches. Greater energy reflection diminishes the energy available to create high run-up. Wave energy reflected by a shingle beach is nearly constant for all values of wave steepness (ratio of wave height to deep-water wave length) greater than 0.02, but reflection increases rapidly as wave steepness decreases below 0.02. Wave breaking at a shingle beach and gravel terrace of the mixed beach is the plunging breaker type, and breaking occurs closer to the shoreline than for the sand beach case. Plunging breakers at the shore confine longshore transport of gravel to the swash zone, as opposed to the entire nearshore zone of sand beaches.

The term “morpho-sedimentary dynamics” is defined as the mutual association and feedback system operating among hydrodynamics, hydraulics, sediment transport, beach morphology, and textural mosaics of the beach surface sediment deposits. This approach marks a fundamental difference with studies and simulations of sand beach processes and morphology change. A morpho-sedimentary-dynamic approach treats sediments and the patterns created with them as both an expression and control on gravel beach behavior.

Buscombe and Masselink (2006) suggest an approach for better understanding of dynamic processes involving field observations of phenomena and the study of the underlying physics through either behavioral modeling or probabilistic analysis. The phenomenological approach adopted in behavioral models attempts to replicate observations of system behavior with numerical expressions without detailing the underlying physical processes of every conceivable variable. A probabilistic analysis might prove fruitful in helping define the statistical properties of a morpho-sedimentary-dynamic system using techniques that focus on system “memory” of the antecedent state. This approach could help describe a beach state where process interactions and feedbacks make unclear how to define independent and dependent variables beyond boundary conditions.

Physical Monitoring

Large-Scale Laboratory

Scaling constraints require mixed sand and gravel physical models to be constructed at prototype scale. The large wave flume (GWK) at the Coastal Research Center at the University of Hannover is one facility capable of such prototype scale experiments. Lopez de San-Roman Blanco et al. (2006) conducted experiments using only gravel and mixtures of sand and gravel as part of the European Union project “Large Scale Modeling of Coarse-Grain Beaches.” Tests generated both random and regular waves, and total test duration was sufficient to produce equilibrium profiles.

Instrumentation measured the sediment surface profile, water surface elevations along the flume, internal pressure in the beach material in the swash zone, water table gradient in the beach, wave run-up, velocities in the surf zone, and sediment distributions along the beach profile. Similar measurements were made at three field sites having sand, gravel, and mixed sediment beaches. Initial to final beach profile change and sediment balance (in the cross-shore orientation) were investigated.

The study produced a conceptual model of physical processes that drive morphological response of beaches. Data analysis validated empirical predictive formulas for profile change and beach crest and step elevations of coarse-grain beaches and examined the qualitative applicability of those

tools to mixed beaches. Blanco's synthesis of observations emphasized that beach material permeability to swash during uprush and the groundwater circulation during backwash were paramount in influencing all other parameters, particularly bottom shear stress and boundary layer formation, that define beach behavior, sediment transport, and morphology.

The barrier dynamic experiment, known as BARDEX (Williams et al. 2012), measured hydrodynamics and morphology change of a prototype scale gravel barrier in the Delta flume (operated by Deltares in the Netherlands) to document dynamic responses to both waves and tides, including varying water levels on the barrier's lagoon side. The project's aims were to better understand processes that form, maintain, and erode gravel barriers and beaches, due to their value in mitigating threats of flooding and erosion, and to refine numerical models as planning and management tools. Analysis results showed that lowering the beach water table promoted net onshore sediment transport in the swash zone and berm development. Elevating the beach water table produced opposite results, which were most pronounced during tidal simulations in the flume.

Field Measurements

Horn and Li (2006), recognizing the important role that asymmetric uprush and backwash velocities due to swash infiltration/exfiltration plays in profile formation, conducted field measurements on a gravel beach (mean particle size 9 mm during tests) at Slapton Sands, Devon, UK. They measured surface pressure under uprush and backwash, subsurface pore water pressures, swash velocities, and bed elevations at eight cross-shore positions during rising and falling tide. They analyzed the field data by using the numerical BeachWin model developed by Li et al. (2002) and compared simulated beach geometry with surveyed profiles. Varying model parameters and observing the resulting computed profile relative to the measured profile showed that major factors are hydraulic conductivity, ratio of sediment transport rate during uprush to that during backwash, and hydraulic friction at the sediment boundary.

Simulating Beach Response

Modeling approaches used to understand and predict the behavior of gravel beaches broadly comprise conceptual models and numerical models. Numerical approaches can be further subdivided into empirical parametric models, based on field and/or laboratory data, and process-based models that can simulate the main physical processes active across the foreshore.

Conceptual models are concerned primarily with large-scale processes such as beach crest overtopping, overwash,

and barrier rollover. In these models, the morphodynamic response of gravel barriers is related to the difference between wave run-up levels and the height of the barrier crest and might include factors of wave steepness and the ratio between the wave period and the swash period. Conceptual models are valuable as a catalyst for developing new empirical and numerical models and thus for advancing understanding of gravel beaches.

Empirical

The bulk of research effort has utilized laboratory-derived data to develop empirical description of gravel beach behavior and has focused primarily on wave run-up and cross-shore beach profiles. Researchers have also developed expressions for beach and barrier stability and for longshore transport of gravel.

Wave run-up. Because the elevation of wave run-up is essential to coastal science and engineering, early researchers developed empirical run-up relationships. The first of these used Iribarren-type parameters (involving profile slope and wave steepness). Stockdon et al. (2006) developed an empirical run-up equation for general applications using the sum of a dimensional Iribarren-based parameter for wave setup at the shoreline, plus total swash excursion, which has separately parameterized dimensional incident and infragravity swash components.

Cross-shore profiles. Detailed scaled laboratory flume tests have provided data for the development of parametric gravel beach profile models. SHINGLE is perhaps the best known and describes the beach profile as a set of three curves defined by the imposed wave conditions and the sediment properties. This model was found to underperform for cases of very coarse sediment or swell or bimodal seas, but GWK experiments reported by Román-Blanco (2006) demonstrated that for some situations SHINGLE reproduced the observed profile response well. Polidoro et al. (2017) presented more recent applications of the model.

HR Wallingford (2016) has studied gravel beach profile response to bimodal sea in the laboratory in a series of scaled laboratory tests which represented four distinct mixes of sediment and bimodal sea states in the UK. The resulting SHINGLE-B model (<http://www.chanelcoast.org/shingleb>) employs four curves (between the prescribed limits) to describe the beach profile. A regression model defines empirical relationships between parameters defining the shape of each curve and bimodal waves. Comparisons between observed and predicted pre- and post-storm gravel beach profiles from SHINGLE-B at three sites in the UK have demonstrated reasonable agreement.

Beach and barrier stability. The response of a gravel barrier crest to waves defines the degree of protection from coastal flooding the barrier can provide to the hinterland. Analysis of laboratory experiments identified the following beach crest

responses: (a) elevation by overtopping, (b) lowered by undermining without overtopping, (c) elevation by overwashing with rollback, (d) lowered by overwashing with rollback, and (e) elevation maintained with the profile contained seaward of the barrier crest.

The barrier inertia model (BIM) relates incident wave steepness to a dimensionless barrier inertia parameter. Local calibration of the BIM is necessary to simulate the appropriate beach crest response. The model may under predict the potential for overwash under conditions of bimodal waves, high wave steepness, and barriers with geometry differing from the original studies. Although subject to these limitations, BIM is a tool widely used by practitioners.

Longshore transport. Van Rijn (2014) described the application of the detailed model CROSMOR for cross-shore and longshore gravel transport. Results from the model compare favorably with laboratory and field data and outperform the widely used CERC formula.

The STRAND model transforms waves between offshore and the start of the swash in a wave-averaged manner. Swash-zone dynamics are explicitly estimated using a ballistic-type approach. The resulting predictions of longshore sediment transport on gravel beaches are shown to correspond well with laboratory data. Work to incorporate longshore models such as STRAND in cross-shore profile models is currently being undertaken and is expected to improve the predictions of profile change attributable to longshore sediment transport gradients.

Process Based

Until recently, numerical models of cross-shore transport have been constrained by an inability to model the processes. The BeachWin model (Li et al. 2002) computes depth-average surface water motions, groundwater flow, and sediment transport and can simulate berm growth relatively well. However, the model is very sensitive to the imposed friction factor and less able to predict cross-shore profile change lower on the beach. The ratio of sediment transport rate during uprush to that during backwash is a key parameter. A major factor in beach profile development was shown to be hydraulic conductivity (swash infiltration), especially in forming the dimensions and location of the berm on the profile. More recent studies of gravel beach profiles demonstrated that friction and sediment transport efficiency are the primary morphological controls, and both these factors vary through time as profiles adjust to the imposed waves and to moderation of wave uprush by infiltration.

In studies to simulate the morphodynamic response of gravel beaches during overwash and berm-building conditions, the wave-averaged XBeach model has been modified to include effects of infiltration, to represent the incident-band swash, and to reduce the onshore-directed transport of sediment. While reproducing observed gravel beach behavior

reasonably well, these studies highlight the need for local calibration when using this kind of model.

McCall et al. (2015) report the development of the wave-resolving version of the XBeach model (XBeach-G) to simulate the morphodynamics of gravel beaches during storms. XBeach-G is a non-hydrostatic extension to the XBeach model that allows solution of intra-wave flow and surface elevation variations due to short waves in intermediate and shallow water depths. XBeach-G also computes the exchange between groundwater and surface water enabling representation of the upper swash infiltration losses and exfiltration effects on lower swash hydrodynamics that are characteristics of gravel beaches with large hydraulic conductivity. Modeling studies of gravel beach storm responses demonstrate XBeach-G performs well.

A simplified version of the 1D model is provided as a GUI (<https://oss.deltares.nl/web/xbeach/xbeach-og>) which allows the user to control the initial profile shape, grain size, and forcing conditions including tidal modulation and bimodal seas with defined spectral characteristics. The model is flexible with regard to defining forcing conditions and outside the GUI XBeach-G can be run in 2D with oblique waves. At present the XBeach-G model has not been fully validated for morphodynamic updating. Model calibration by the user is required. Typically, this is based on pre- and post-storm profiles.

Summary

Gravel beaches are categorized by their sediment makeup as shingle or mixed sand and gravel. Gravel beach processes and morphology are important subjects to coastal science and engineering because maintaining and even constructing these features are valuable for environmental and flood protection reasons. A gap remains between the present understanding of the physical processes controlling gravel beach morphology and the integration of detailed processes in models. In almost all cases, empirical and process-based models have been developed and validated using data from physical model experiments focused on the beach face processes. Increasingly it is demonstrated that process-based models provide more reliable prediction of gravel beach behavior, especially during extreme events. Developing and disseminating user-friendly versions of these models would benefit practical coastal management applications concerned with storm impacts and flood safety.

Cross-References

- [Beach Sediment Characteristics](#)
- [Cross-Shore Sediment Transport](#)

- [Dynamic Equilibrium of Beaches](#)
- [Gravel Barriers](#)
- [Longshore Sediment Transport](#)

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Gross Transport

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Gross transport refers to the total movement of sand approximately parallel to the shoreline over a specified period, often 1 year (see entries on “► [Cross-shore Sediment Transport](#)” and “► [Longshore Sediment Transport](#)”). Waves approaching the shore at an oblique angle, such that the crests of the breakers are at an angle to the shoreline, generate longshore currents that convey the mobilized sand along the shore. In many locales, the wave approach direction varies over time such that this transport can be in either direction relative to the

shoreline (see entry on “► [Waves](#)”). Consequently, sand may move upcoast for an interval and then, as wave conditions change or the lesser effects of tidal or wind-driven currents intervene, reverse direction and move downcoast. Gross transport amounts or rates account for all transport regardless of direction. The units of gross transport are typically volumetric rates (cubic meters per year).

There are difficulties associated with estimating gross transport. First, an estimate requires comprehensive sets of data collected over long periods. These would typically include the height and the direction of breaking waves measured for long enough intervals to average out the randomness and often enough to include the real variation in waves brought about by changing atmospheric forcing. No autonomous method for measuring either the height or the angle of breakers has been developed. However, a technique for making autonomous measurements outside the surf zone and converting them to equivalent values for calculating longshore transport has been developed (Seymour and Higgins 1978). Finally, the models for longshore transport are rudimentary and can result in estimates with very large errors, even when quality wave data are available. Only occasionally, special conditions exist that will allow direct measurement of gross transport (Seymour et al. 1981). One such condition is the construction of a jettied entrance that will function for a sufficiently long period as an effective trap for longshore transport in both directions.

The significance of gross transport to coastal engineers is related to understanding the fate of beach sand in the presence of some mechanism for sequestering or otherwise removing sand from the system. In other instances, the coastal engineer is usually concerned with the difference between the upcoast and downcoast transport (see entry on “► [Net Transport](#)”). Entrances to tidally influenced bays or lagoons along sandy coasts will typically form shoals offshore and inshore of the entrance channel. These shoals can impound very large quantities of sand. If a new entrance is constructed (see entry on “► [Navigation Structures](#)”) coastal engineers must be concerned with the impact on the adjacent beaches during the formation of these shoals. The availability of sand to form shoals is a function of the local gross transport so that this value is useful in determining strategies for remediation. Further, dredging of shoals in natural or constructed inlets may be required before they reach their equilibrium extents in order to maintain safe navigation channels. Again, the rates at which these channels will refill are a function of the gross transport. These data are useful in estimating the frequency and cost of dredging operations.

Narrow continental shelves often contain submarine canyons (see entry on “► [Continental Shelves](#)”). These underwater valleys often extend from close to the shore to great depths beyond the shelf. It is generally agreed that these canyons have been formed over geological time by erosion

from episodic sand-laden gravity-driven flows (Heezen and Ewing 1952). The presence of huge depositional fans at the canyon outlets indicates that these features can function as significant loss mechanisms for beach sand. The canyon head, near the shoreline, fills with sand delivered by longshore transport. Typically, the canyon will fill during strong longshore transport events regardless of their direction. Thus, gross transport rates provide an indication of the potential for losses at submarine canyons, which could have significant effects on the condition of associated beaches.

Cross-References

- Continental Shelves
- Cross-Shore Sediment Transport
- Energy and Sediment Budgets of the Global Coastal Zone
- Longshore Sediment Transport
- Navigation Structures
- Net Transport
- Sediment Budget

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Ground-Penetrating Radar

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Definition

Geophysical technique that uses radar pulses for subsurface imaging.

Ground-penetrating radar (GPR or georadar) is a subsurface imaging technique that has revolutionized coastal research. Originally designed for the purpose of

environmental and geotechnical investigation of the shallow subsurface, this high-resolution geophysical method is being successfully used in stratigraphic research of sedimentary sequences of various age and composition. GPR was instrumental in imaging the internal architecture of a variety of recent geological settings: fluvial (Leclerc and Hickin 1997; Roberts et al. 1997), glaciofluvial (Beres et al. 1995, 1999), periglacial (Jol et al. 1996a; Busby and Merritt 1999), eolian (Schenk et al. 1993; Harari 1996; Jol et al. 1998), and lake deltas (Jol and Smith 1991; Smith and Jol 1997). Its application in coastal research has led to significant advances in our understanding of mesoscale (centimeters to 10s of meters) stratigraphy of marginal marine sequences and has already resulted in several radar facies models (Baker 1991; FitzGerald et al. 1992, 2000; Jol et al. 1996b; van Heteren et al. 1996, 1998; Smith et al. 1999; Buynevich et al. 2009; Hein et al. 2016). Knowledge of the principles and limitations of the GPR technique in coastal settings is essential if accurate and complete models of large-scale coastal development are attempted. A general introduction to the GPR method and discussion of its applications to coastal stratigraphic research are presented here.

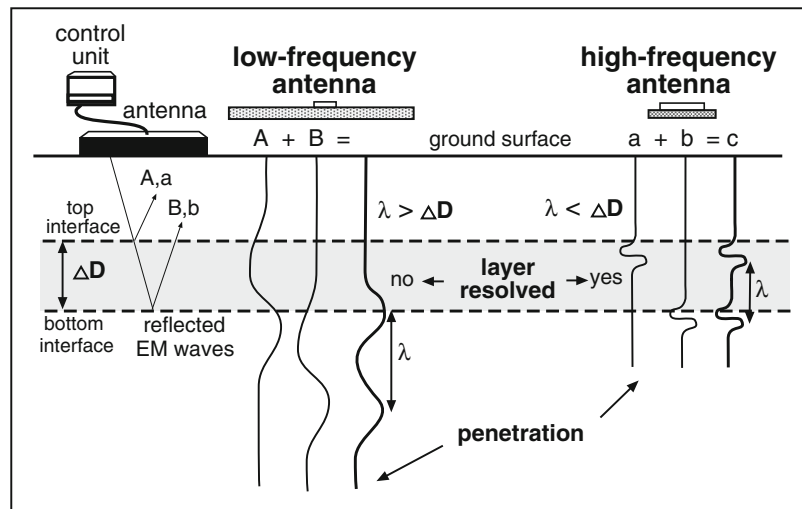
Theoretical Background and Methodology

Georadar method is in many ways similar to seismic reflection profiling, except that it uses electromagnetic (EM), rather than acoustic, waves with the frequency range of 10–2300 MHz. The behavior of the EM waves, governed by Maxwell's equations, is controlled by many factors, including electric conductivity, magnetic permeability, and dielectric permittivity. The first two parameters control the attenuation and ultimately the penetration depth of the EM signal. Highly conductive (less dielectric) materials result in signal dissipation and loss. Sensitivity and frequency of the antenna, as well as the gain setting on the control unit, also control the penetration and resolution of the GPR record (Fig. 1; Davis and Annan 1989; Conyers and Goodman 1997; van Heteren et al. 1998).

Dielectric permittivity is the ability of a material to become polarized and respond to incoming EM waves (von Hippel 1954). Relative dielectric permittivity (ϵ_r) is a dimensionless constant, which is obtained by comparing the permittivity of a particular material with that of a vacuum (which is one). This parameter controls the velocity of EM waves and is inversely related to it:

$$v = \frac{c}{\sqrt{\epsilon_r}} \quad (1)$$

where v is the velocity measured in nanoseconds ($1 \text{ ns} = 10^{-9} \text{ s}$) and c is the speed of light (0.2998 m/ns). Typical



Ground-Penetrating Radar, Fig. 1 Generalized diagram depicting the differences in resolution and penetration depth for the low- and high-frequency antennae. Low-frequency wave (A) defines the top interface, but the distance between the interfaces (ΔD) is less than the wavelength (B). The relatively long wavelength (λ) of the resulting EM signal (C)

does not resolve both interfaces. A high-frequency EM wave defines both the top (a) and bottom (b) interfaces. The resulting wave (c) is shorter than the distance between the two interfaces. The penetration in this case, however, is less than with a low-frequency antenna (Modified after Conyers and Goodman 1997)

values of relative dielectric permittivities for common geological materials are presented in Table 1. The magnitude of a particular subsurface reflection is directly related to the difference in dielectric properties of the two subsurface materials. From Eq. 1, the velocity of EM waves ranges from about 3 cm/ns in water to 30 cm/ns in air. Therefore, the groundwater table commonly appears as a strong subhorizontal reflector on the radar trace. Above it, the unsaturated portion of the sedimentary sequence with lower permittivity values will be penetrated more readily by the radar signal. Consequently, a greater thickness of dry sediments can be represented on a particular record, compared with the part of the sequence below the water table, and will have a lower vertical exaggeration than the saturated portion. Once the velocities of EM waves or dielectric constants of specific lithologies (both saturated and unsaturated) are established and the time (t , ns) is obtained from the record, the approximate depth (D) to a particular target layer may be calculated using the two-way travel time between the ground surface and the layer in question, or between two layers as:

$$D = \frac{vt}{2} = \frac{ct}{2\sqrt{\epsilon_r}} \quad (2)$$

Sediment characteristics that affect the subsurface behavior of EM waves include pore-water and clay mineral content and their chemical composition, as well as magnetic properties of constituent minerals (Topp et al. 1980; van Heteren et al. 1998). High concentrations of heavy minerals (e.g., magnetite, ilmenite, garnet, epidote) form as storm-lag deposits on

Ground-Penetrating Radar, Table 1 Relative dielectric permittivity (ϵ_r) of common earth materials

Material	ϵ_r
Air	1
Ice	3–4
Freshwater	80
Seawater	81–87
Sand	
Unsaturated	3–7
Saturated	20–32
Unsaturated, with gravel	3.5–6.5
Saturated, with gravel	15.5–17.5
Silt	
Unsaturated	2.5–5.0
Saturated	20–30
Clay	
Unsaturated	2.5–5.0
Saturated	15–40
Bedrock	4–15

Source: Davis and Annan (1989), Conyers and Goodman (1997), van Heteren et al. (1998)

beaches or deflation lag in dunes, often producing prominent reflectors in radargrams (Buynevich et al. 2009). In some cases these sedimentological anomalies, as well as thick iron-stained horizons, may have high-enough magnetic permeability values such that they attenuate the magnetic portion of the EM signal and preclude further penetration (Topp et al. 1980).

Ground-penetrating radar profiles taken over areas with substantial relief need to be topographically corrected (surface normalized). This arises from the fact that the ground

surface on the time-record output is represented as the horizontal surface. For example, on a profile taken over a dune, the dune surface will be depicted as a horizontal reflector, and (sub)horizontal layers of the underlying sequence (e.g., dune/beach contact, washover deposit, peat horizon) will appear as concave-upward reflections (i.e., the mirror images of dune topography). For short segments of the trace, groundwater table, if present, may be assumed (or determined) to be nearly horizontal and thus aid in correcting the record.

The radar data are commonly collected in either step mode or continuous mode. In the step mode, the source of the signal (transmitter) and the receiver are placed on the ground and, after the reading is obtained, moved to the next location. Continuous mode involves the unidirectional movement of both transmitter and receiver (usually housed within one antenna box). The cart-mounted GPR system allows rapid collection of continuous traces over large areas and has proven to be an effective method of subsurface data collection. In recent years, the advances in radar technology and software have improved the resolution and processing of the field data. Using a series of parallel, closely spaced transects, a three-dimensional image of the subsurface, as well as plan-view time slices that correspond to various depths, can be obtained (Beres et al. 1995, 1999; Conyers and Goodman 1997). Most recently, a combination of time-domain reflectometry and sediment peels has been used by Van Dam and Schlager (2000) to identify the reflector-producing horizons and construct synthetic radar traces analogous to synthetic seismograms. Because the description and analysis of sedimentary facies on GPR records are similar in many ways to seismic-stratigraphic attributes (e.g., reflector configuration, frequency, continuity, etc.), the onshore records of the barrier lithosome can be integrated with the available nearshore and offshore seismic reflection profiles (van Heteren et al. 1998).

Limitations of the Techniques

The most obvious limitations of a specific GPR system setup are the penetration depth and resolution, which are traded off based on antenna center frequency. Both of these attributes depend on the choice of the antenna, which is in turn governed by the research objectives (Smith and Jol 1995). Figure 1 shows that the top and bottom interfaces of a subsurface layer can only be resolved if these are separated by at least one wavelength of the radar signal (cf. Davis and Annan 1989). High-frequency antennae (500–2000 MHz) are smaller in size and provide high resolution at the expense of relatively limited penetration (typically less than 8–10 m). Antennae with frequency range of 12.5–50 MHz have poor resolution but allow for a maximum probable penetration of 45–65 m (Smith and Jol 1995). The optimal antenna

frequencies for stratigraphic research are 100 MHz or 120 MHz, which allow penetration of 15–20 m (deeper in unsaturated sequences), while still providing high-resolution images.

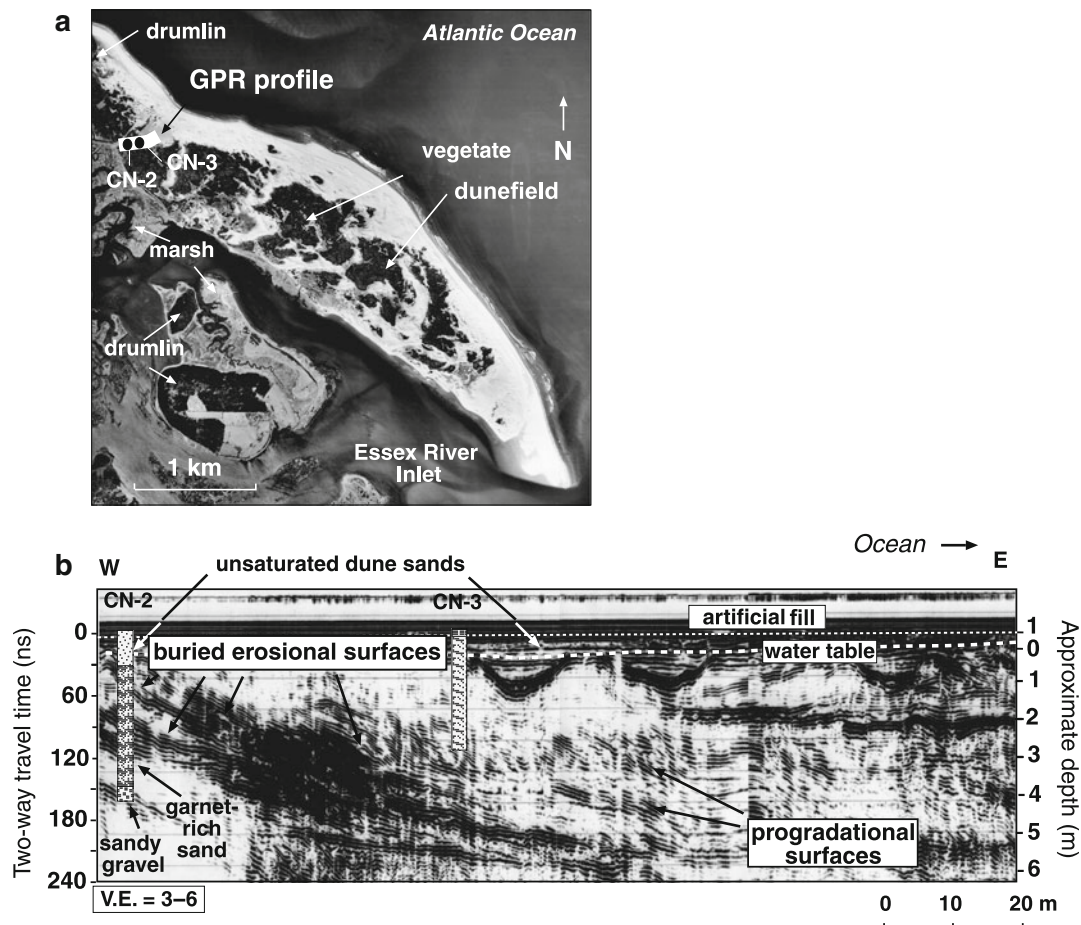
One of the important considerations in GPR research is the attenuation of the EM signal by brackish to saline groundwater due to its high conductivity. As a result, sections of many profiles adjacent to a beach or backbarrier margin, or deeper portions of profiles affected by saltwater intrusion, are often reflection-free (van Heteren et al. 1998; Buynevich et al. 2009). In some locations, the barrier is wide enough such that most of the stratigraphy can be imaged by GPR without attenuation. Besides seawater, increase in clay content drastically increases conductivity thereby reducing or precluding signal penetration. The top of the clay unit itself, however, often appears as a strong reflector on the radar trace. As mentioned above, sedimentary deposits with high magnetic permeability may also limit signal penetration while themselves providing strong reflectors. Aside from sedimentological characteristics, the distortion and amplitude change of the EM waves as they propagate into the earth may reduce the accuracy of the final depth/distance calculations.

Due to inherent differences in sediment properties from site to site, the velocity of EM waves may also vary (Davis and Annan 1989; van Heteren et al. 1996). The velocity values for a particular site may be calibrated through common midpoint analysis (Jol and Smith 1991) or by measuring the depth to specific marker horizons in sediment cores and calculating velocity values from Eq. 2. In general, unless the geology of a study site is well known, sediment cores should be taken along the GPR profile to interpret the major reflectors. In turn, the radar images of dipping subsurface horizons (e.g., buried marsh layer, sloping bedrock surface, tidal inlet channel) may be used to maximize the coring effort by planning a core site in the area where the depth to a target reflector is minimal.

Examples of GPR Images from Coastal Environments

Styles of Barrier Progradation

Many coastal areas have experienced progradation (seaward) growth. Geomorphic expression of this process, such as a series of beach ridges, is often used to determine the origin, magnitude, orientation, and chronology of barrier growth (Tanner 1995). However, in many cases dense vegetation, parabolic dune migration, or human development have modified or obscured the surface expression of barrier growth (Fig. 2a). In such areas, subsurface records, complemented with sediment cores, may provide the only means of analyzing the erosional-depositional history of a barrier. In many instances the progradation of a barrier may be punctuated by



Ground-Penetrating Radar, Fig. 2 (a) Vertical aerial photograph of Castle Neck Barrier, Massachusetts, dominated by vegetated parabolic dunes. Note the absence of beach ridges. (b) Shore-normal GPR transect taken across a parking lot reveals a series of strong seaward-dipping reflectors in the landward segment of the GPR trace giving way to a

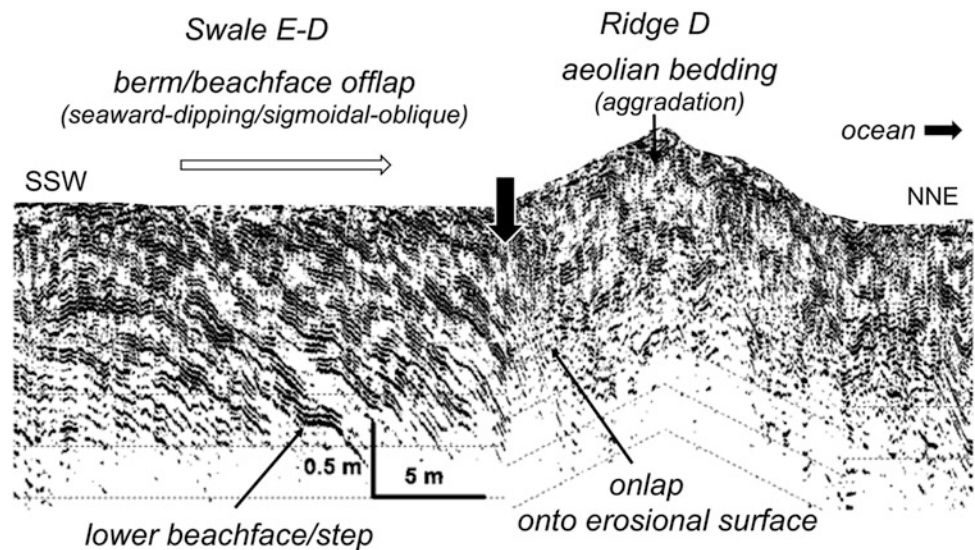
sequence of less prominent, nearly uniformly spaced reflectors in a seaward direction. Sediment core CN-2 penetrated several layers of concentrated garnet-magnetite sands interbedded with quartz-rich units. All records are taken as a continuous trace with a 120 MHz antenna. See text for discussion

erosion and shore retreat. At these sites, there may be no distinct morphological evidence of an erosional event, except occasional washovers or dune and berm scarps confined to the youngest portion of the barrier. These features, particularly in the earlier constructional history of the barrier, are often preserved as buried accumulations of coarse-grained sediments or heavy-mineral concentrations (HMCs) that are rarely detected in the field (Buynevich et al. 2009). Such lithological anomalies may be observed in sediment cores, but their geometry and continuity can only be confirmed in geophysical records (Fig. 2b). For example, the GPR profile in Fig. 2b illustrates a series of prominent tangential-oblique reflectors that represent buried erosional beach face and berm scarps. They grade into uniformly spaced sigmoidal-oblique reflectors in a seaward direction, which mark a period of increased sediment supply (Buynevich et al. 2009). Similar records were recently obtained in carbonate settings, showing the potential role of storm scarps (truncation) as nuclei for beach/dune ridges in Bahamian strandplains (Fig. 3).

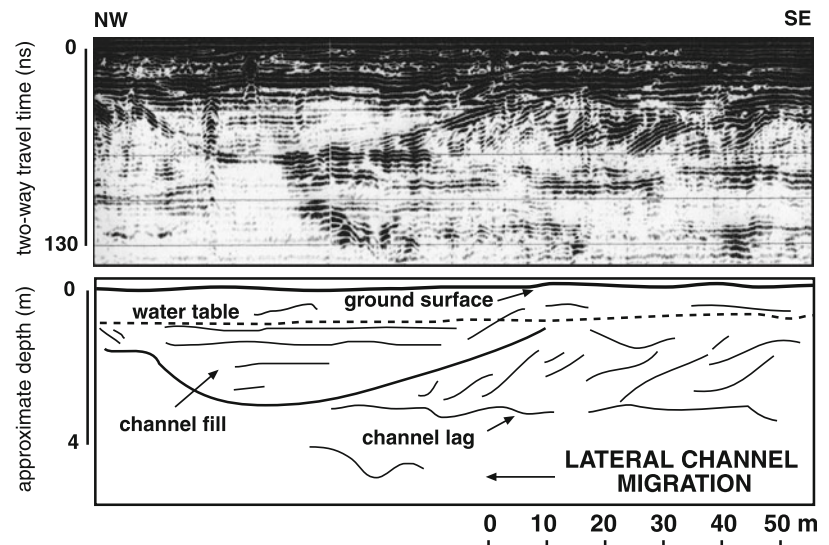
Geometry of Tidal Inlet Paleo-channels

Channel-fill sequences of tidal inlets may comprise a significant portion of the barrier lithosome, and, in some instances, the locations and dimensions of former inlet channels can be detected with GPR. Mixed sediment barriers are ideal for the recognition of inlet-fill structures. Due to large contrast between the coarse-grained channel lag and finer-grained channel-fill deposits, the outline of the channel often appears as a prominent concave-upward reflector. Figure 4 shows a paleo-inlet channel that has migrated along a retrograding, sand-and-gravel barrier as evidenced by a series of northward-dipping reflectors. Eventually, the inlet stabilized in one location and was infilled by sediment from a seaward source recorded as subhorizontal reflectors within the paleo-channel. Using GPR profiling, the locations of the former inlets can be mapped and compared with historical maps, where available. At least 18 historical inlets were mapped along Duxbury Beach, Massachusetts, where none exist today (FitzGerald et al. 2000). In addition, such elements of inlet

Ground-Penetrating Radar,
Fig. 3 Shore-normal radargram across Moriah Cay strandplain, Little Exuma, the Bahamas. Note the truncation of paleo-beach face surfaces (*large arrow*) overlain by a dune ridge, suggesting a genetic relationship



Ground-Penetrating Radar,
Fig. 4 Shore-parallel profile and interpretation of a buried tidal inlet paleo-channel at Duxbury Beach barrier, Massachusetts. A series of dipping reflectors indicate the migration of the channel in a northward direction. A prominent concave reflector on the *left* represents the final position of the paleo-channel that was subsequently filled



channel geometry as depth, width, and approximate length (using a series of records) can be determined. The elevation of the paleo-inlet channel relative to present sea level can also be estimated.

Stratigraphy of Coastal Lakes

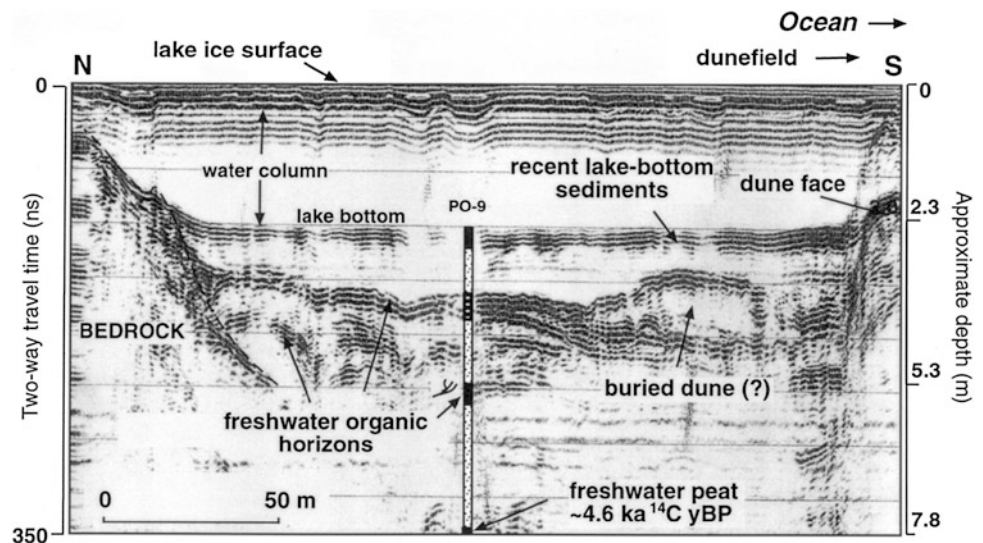
Freshwater lakes and ponds of various origins (closed lagoons, glacial depressions, dune swales, deflation basins, etc.) are common features along many coasts. Their sedimentary fills serve as archives of depositional events that result from climatic, oceanographic, and geomorphic changes in coastal regions. Shifts from organic- to clastic-dominated deposition result in a sequence of interbedded layers with distinct lithological and dielectric properties, making these systems suitable for GPR research. Figure 5 shows a shore-normal survey across an ice-covered coastal lake. A steeply dipping reflector representing the bedrock surface can be traced to a depth of

over 7 m below the lake surface. The seaward portion of the trace shows the margin of the barrier dune field as a landward-dipping surface with several basinward-dipping internal reflections. Below the flat lake-bottom reflector, a sequence of wavy reflectors can be traced across the profile. These represent muddy, organic-rich lake bottom sediments interbedded with eolian sands. A prominent convex-up reflector with a transparent core is indicative of a buried dune. This interpretation is based on similar reflector configuration observed on the lake bottom adjacent to a recently migrating dune.

Summary

Ground-penetrating radar is continuing to serve a valuable tool for high-resolution imaging of antecedent geology, stratigraphy, and hydrogeology of coastal systems. Although

Ground-Penetrating Radar,
Fig. 5 Radar transect taken over an ice-covered surface of Silver Lake, Maine. The undulating subhorizontal reflectors are organic-rich lake-bottom deposits. Note: The convex-upward reflection within the lake sequence interpreted as a buried dune



saltwater attenuation presents a significant limitation in coastal lowlands, areas with moderate to high rainfall and relatively good sediment permeability often contain considerable freshwater lenses (5–20 m), which ensure good penetration of EM signal. Varying degrees of textural and compositional heterogeneity of sediments in many coastal sequences produce the lithological contrast necessary to generate subsurface reflections. These systems provide excellent natural laboratories for effective and detailed stratigraphic analysis using GPR profiling supplemented with sediment cores. Such studies have already significantly improved our knowledge of coastal development over a wide range of temporal (years to millennia) and spatial (centimeters to 10s of kilometers) scales and served to emphasize the complexity of coastal processes and resulting stratigraphic records.

Cross-References

- Beach Stratigraphy
- Beach and Nearshore Instrumentation
- Coastal Sedimentary Facies
- Hydrology of the Coastal Zone
- Monitoring Coastal Geomorphology
- Paleocoastlines
- Sequence Stratigraphy

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Gulf Shorelines, Last Eustatic Cycle

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Introduction

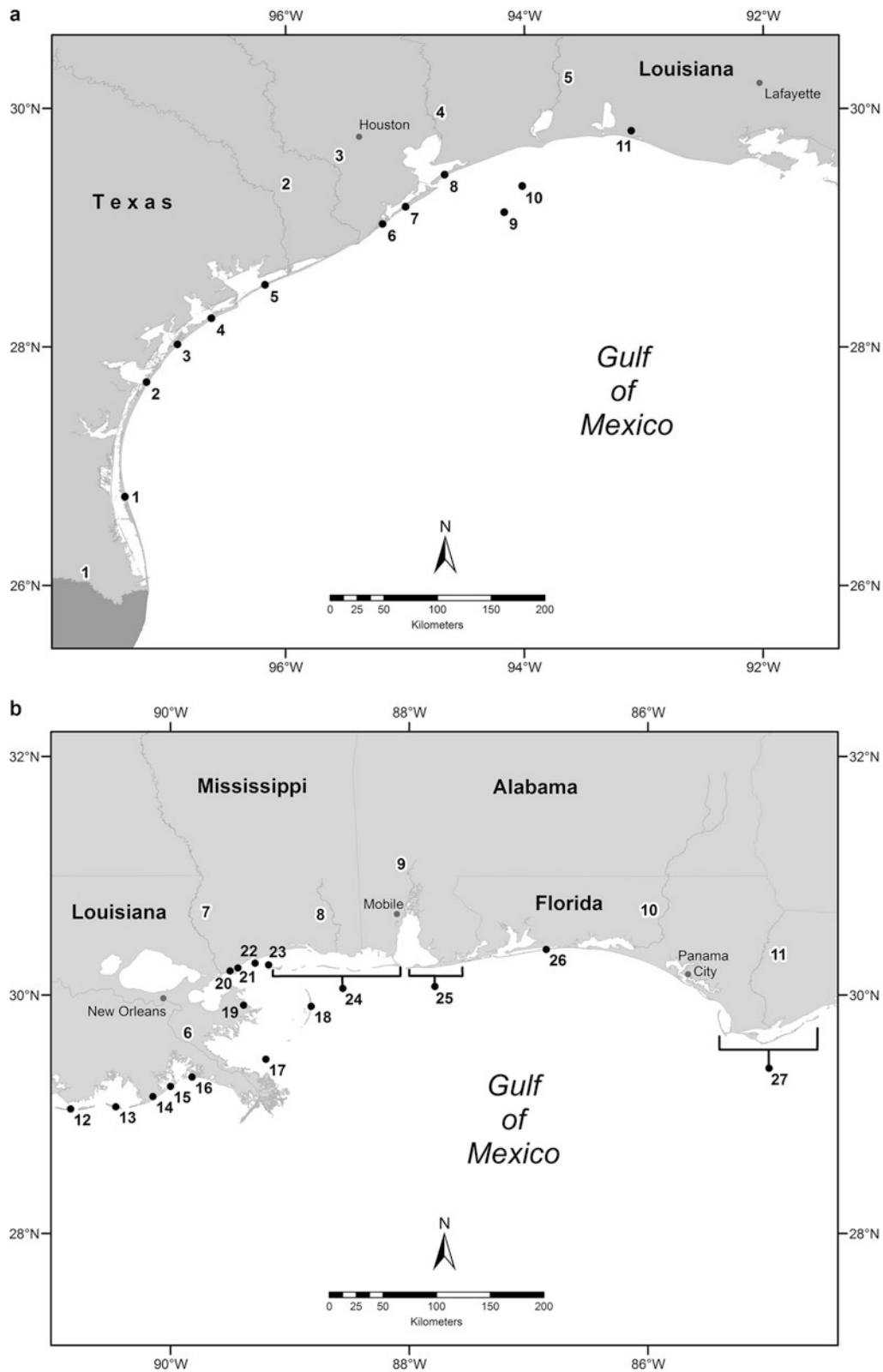
As the result of detailed geophysical, coastal, and geological surveys, including drillcore and seismic studies conducted in increasing detail, the Gulf of Mexico is among the best explored marine basins, in terms of its late Quaternary history as well. Conditions throughout the Last Eustatic Cycle (LEC) provide insight into potential future coastal and nearshore changes. This summary broadly outlines northern and eastern Gulf coastal changes between the Last Interglacial (LIG) climate optimum (thermal maximum) and the present highstand. Landform development during marine highstand,

lowstand, and intermediate phases; their influence on Gulfwide sedimentation patterns following the Last Glacial Maximum (LGM) indicate the overwhelming role both of relict and modern sediment sources in the construction modes and dimensions of coastal barriers, deltaic and other lithosomes and their highly varied sediment sources. Coastal evolution reflected significant regional differences between the NW, NE, and central (Mississippi Delta) sectors in terms of barrier categories, the development of barrier chains and other coastal, nearshore, and shelf landforms, as well as climate fluctuations. These conditions involved both wetter and more arid phases, with consequent changes in fluvial discharge volumes.

Mainland Sediment Sources: Northwest vs. Northeast

Direct discharge and secondary reworking of Mississippi River shelf deltas and drowned coastline lithosomes provided the lion's share of sediments in the NW Gulf region. Before man-induced major hydrological changes in its watersheds prior to ~1900 AD, the annual sediment discharge of the Mississippi to the Gulf was ~400 million metric tons. A substantial portion consisted of coarser suspended load. Sand bedload is routinely estimated at <5% of the total (Meade and Moody 2010). Among lesser NW coastal plain streams, contribution by the defunct "Western Louisiana River" (probably, the early Red River) and the integrated Brazos-Trinity-Sabine stream network was the most significant. Colorado and Rio Grande (Fig. 1) discharge volumes followed in importance. Several streams exceeded their current volume during wetter climate phases (Anderson et al. 2014, 2016). Despite reduced stream runoff, erosion-driven diminished vegetation cover, caused by recurring Late Quaternary dry episodes, may have also increased sediment yields (Otvos 2004).

In addition to the large Mobile and Apalachicola (Fig. 1) sediment supply, sediment delivery from mid- and outer-shelf relict deltas related to the Choctawhatchee, Perdido, Escambia, Blackwater, and Yellow rivers also had an impact on the relatively narrow NE shelf (Bart and Anderson 2004). In contrast, sediments from primary fluvial, estuarine, and reworked secondary (relict) deposits account for the wider distribution and greater dimensions of the dominant relict and modern deltas and shore barrier sources. In the NW, 174-km-long Padre Island, Texas (Fig. 1), which is among the longest barrier islands globally. Barrier spits in the NW are 3–8 km wide and exceed 32 km in length. In comparison, only three barriers exceed 24 km length on the NE Gulf. Few barriers reach 1.0–1.5 km width; most remain in the 200–800 m range. The extensive interglacial Gulfport barrier ridge system and its Wisconsinan-to-early Holocene eolian sand cover



Gulf Shorelines, Last Eustatic Cycle, Fig. 1 (a and b) Location index. Rivers: 1. Rio Grande; 2. Colorado; 3. Brazos; 4. Trinity; 5. Sabine; 6. Mississippi; 7. Pearl; 8. Pascagoula; 9. Mobile-Tensaw; 10. Choctawhatchee; 11. Apalachicola. Islands and other sites: 1. Padre Is.; 2. Mustang Is.; 3. St. Joseph Is.; 4. Matagorda Is.; 5. Matagorda barrier spit; 6. Follets Is.; 7. Galveston Is.; 8. Bolivar barrier spit; 9. Heald Bank; 10. Sabine Bank; 11. Center of SW Louisiana Chenier plain;

12. Isles Dernieres (Last Islands); 13. Timbalier Is.; 14. Moreau-Caminada ridge plain; 15. Grand Isle; 16. Cheniere Ronquille; 17. Breton Is.; 18. Chandeleur Is.; 19. St. Bernard subdelta plain; 20. Pt. Clear Is.; 21. Campbell Is.; 22. Square Handkerchief Shoal; 23. Cat Island Shoal; 24. Mississippi-Alabama barrier chain; 25. Morgan-Perdido strand plain belt (Figs. 3 and 4); 26. Santa Rosa Is.; 27. Apalachicola Coast

represented major sand sources for mid-to-late Holocene barriers along the NE mainland shore (q.v., ► [“Last Interglacial”](#) encyclopedia entry).

Secondary Sediment Sources

Deposits reworked during transgressive and regressive phases of the eustatic cycle include relict shelf-margin, mid-shelf delta deposits, and associated river and tidal channel, fluvial, deltaic, and marine bar sands. Relict Mississippi deltas represent the largest submerged lithosomes. Combined with other shelf deposits eroded during periods of gradual or intermittent transgression, deep transgressive ravinement erosion produced large volumes of reworked sediments, subsequently transferred landward by various means of cross-shelf transport. While extensive outer and mid-shelf deltas occupied the northwestern continental shelf, only the combined Mobile-Pascagoula-Pearl system, source of the large relict “Lagniappe” delta, and, to lesser extent, the Apalachicola River created sizable shelf-margin deltas in the northeast (Kindinger et al. 1994; Roberts 1997). The best preserved assumed paleoshorelines, those of Substage MIS 5a and Stage MIS 3, were also important sediment contributors in the NW (Anderson et al. 2014), while only fewer and smaller linear barriers, relict shoreline lithosomes, have been identified in the presently submerged NE and E Gulf shelf areas.

Interglacial Highstand (~130–112 ka) MIS 5e

Surface erosion during the preceding low sea-level phases, bolstered by uplift of the coastal hinterland that is capped by the Pliocene Citronelle Formation removed all pre-MIS 5e shore lithosomes from the inner shelf and the coastal plain. Citronelle-correlative, alluvial, in part fossiliferous Late Pliocene paralic, and nearshore deposits directly underlie the Late Pleistocene in the Mississippi-Alabama coast and nearshore (Otvos 1997; 2001; USGS 1996, 2001). Older Pleistocene estuarine and marine units are absent. The LIG highstand shore during the LIG climate optimum is represented by the 1–2-km-wide, ~5–10-m-high Gulfport barrier chain often topped by substantial Wisconsinian eolian sand beds. From trace fossil content, sedimentary structures, and ridge morphology of the Gulfport barrier, the relative sea level (RSL) was estimated at ~+3 m to +5 m (q.v., ► [“Last Interglacial”](#) entry, present volume). This semicontinuous ridge trend flanks the present mainland shore between SE Louisiana and Florida Peninsula (Otvos 2001, 2004; Fig. 1 in Otvos 2005). In contrast, while not exposed to the open Gulf, only along the west-central Texas coast is the correlative Ingleside barrier trend continuous. A 20-km-long shore-parallel escarpment inland, mistaken earlier for an older Pleistocene marine shoreline, separates the OSL-dated penultimate interglacial

(MIS 7) alluvial Montgomery coastal terrace with its significant flora assemblage from the seaward adjacent younger LIG deposits. The latter units include the late Pleistocene nearshore-coastal Gulfport, Biloxi, and alluvial Prairie Formations and related land surfaces on the Mississippi coast. A recent seismic survey (Bullock et al. 2017) confirmed the fault line, instead of the earlier assumed “shoreline-scarp” origin of Big Ridge Scarp in Mississippi (Otvos 1981a). The single Pleistocene marine–estuarine complex along the Gulf coastal plain stands in sharp contrast to the multiple marine coastal terrace stairsteps of the U.S. south Atlantic coast (Otvos 2018).

Major Sea-Level Decline (~112–59 ka) MIS 5d–MIS4

The early preglacial Wisconsinian MIS 5a paleoshoreline trend is well established on the NW Gulf shelf. Its eroded deposits also nourished younger shoreline lithosomes (Berryhill and Suter 1986). A rapid ~60-m sea-level drop steepened stream gradients and enlarged drainage basins across the newly exposed former shelf area. Substantial erosion accompanied valley extension and downcutting. Incised river channels terminated in numerous mid-shelf deltas (Anderson et al. 2014, 2016).

Fluctuating, Subsequently Declining Sea Level (~59–28 ka) MIS 3

From its Stage MIS 4 lowstand at ~–100 m, sea level rose to between –30 m and –60 m. This rise was followed by decline and brief stillstand. High-frequency, short-term oscillations were superimposed on the sea-level changes. In addition to the Mississippi, Brazos, Colorado, the long-defunct “Western Louisiana River,” and the Rio Grande (Fig. 1) also constructed sizable deltas along the western shelf margin. Accompanying further shifts of the regressing shoreline, a large mid-shelf ridge plain also came into being (Berryhill and Suter 1986; Anderson et al. 2014, 2016). Only the combined Mobile-Pascagoula stream network, possibly including the Pearl, as well as the Apalachicola River formed sizable coeval deltas on the northeastern shelf. At least 12 Stage MIS 3 deltas and delta clusters that were located at –55 m to –85 m depths in the mid- and outer NW Florida-Alabama shelf multibeam sonar survey also revealed numerous small relict island lithosomes, associated with relict deltas. Lithified by carbonate cementation, during the subsequent marine immersion, “hardground” surfaces protected submerged barriers against erosion (Gardner et al. 2005, 2007). A 38-km-long drowned shelf-margin barrier of 8-m surface relief and ridge-and-swale topography was discovered at –65 m to –69 m depth on the Alabama-SE Mississippi shelf (Gardner et al. 2001).

Lowstand (~28–18 ka) MIS 2

Northwestern Gulf

Continuous sea-level decline from ~–80 m to full-glacial (LGM) –120 m low valley entrenchment and channel incision accompanied rapid stream extension seaward. Large shelf-margin deltas formed along and gulfward of the lowstand shoreline. Sediment fill choked several incised stream valleys (Anderson et al. 2016). Holmes (2011) assigned ~14 ka age to a seismically identified prominent submerged barrier located SW of the large Colorado-Brazos shelf-edge delta complex. The island's –75 m crest and –120 m base values, however, suggest LGM lowstand age, instead.

Northeastern Gulf

Few large and several smaller river deltas prograded on the shelf margin along the ~–120 m lowstand shoreline between ~21 ka and 18 ka. In addition to the sizable “Lagniappe Delta,” constructed by the combined Mobile-Pascagoula (-Pearl?) river network (Kindinger 1988; Kindinger et al. 1994), Bart and Anderson (2004) also recognized shelf-edge deltas related to the MIS 2 lowstand.

Rapid Late to Postglacial Transgression (18–4 ka) MIS 1

Intensive transgressive ravinement erosion of relict and contemporary seafloor deposits marked rapid transgression, driven by wholesale global ice melt. Landforms that marked successive, often stepwise sets of higher shorelines at best were only partially preserved. Variable rates of deglacial sea-level rise, fluvial discharge volumes, onshore-driven cross-shelf sediment transport, and major differences in antecedent topography controlled modes and rates of landform evolution. These factors impacted barrier islands, shoals, bays, back barrier, lagoon development, and shore configuration. Barrier overstepping events followed transgressive island drowning that created new shoals and emerging barrier islands landward of extinguished ones (q.v., ► [“Barrier Island Formation and Development Modes”](#) entry). Delta shorelines underwent major landward shifts across the mid- and inner shelf. Influenced by deglacial meltwater pulses, sea-level rise prior to ~8 ka frequently followed a stepwise pattern. Sediment infill choked several incised valleys. Subrecent barrier development and mainland coast progradation began by ~5–6 ka during significant deceleration of the sea-level rise (Anderson et al. 2014).

Northwestern Gulf

Stepped sea-level rise episodes driven by meltwater events resumed between ~10 ka and 7 ka (Anderson et al. 2016).

Following transgressive overstep by early Holocene Brazos-Colorado deltas, aggradational Mustang Island, the assumed oldest modern Gulf barrier was established between ~9.5 ka and 7.5 ka in the flooded, deep Nueces Valley (Shideler 1986; Anderson et al. 2014). In contrast to the smooth outline of the central Texas shore, an irregular shoreline characterized the coast near the incised Trinity Valley of E Texas. Subbottom data from E Texas inner shelf, landward of the present Heald and Sabine Banks, revealed barrier island-fronted early and mid-Holocene estuaries, dated ~8.4–~7.7 ka and ~7.4–4.7 ka, respectively. The widths of mid- and late Holocene lagoons in the rear of these barriers compared favorably with the <45 km width of modern Pamlico Sound, North Carolina. “Sabine Bank Island,” in existence between ~5.3 ka and 4.7 ka, and present Galveston Island (initiated ~5.5 ka) became links in the active barrier chains between ~7.7 ka and 4.7 ka. Shoal aggradation was likely to have initiated Matagorda Island ~6.7 ka. Transgressive deposits buried estuarine beds before gulfward barrier progradation (Wilkinson 1975). Barrier overstepping, emergence from newly established marine sand shoals landward of the previous barrier sites followed island decay (q.v., ► [“Barrier Island Formation and Development Modes”](#) entry). After ~7.5 ka B.P. and ~4.7 ka, respectively, pulses of rapid mid-Holocene sea-level rise led to transgressive overstepping of Heald, respectively, Sabine Bank islands (Anderson et al. 2014, 2016); (q.v., ► [“Barrier Island Formation and Development Modes”](#) and ► [“Coastal Barrier Preservation and Destruction”](#) entries). North Padre, initiated shortly after the emergence of Galveston Island, both by shoal aggradations, also underwent extensive progradational expansion (Bernard et al. 1970; Fisk 1959). Shallow lagoonal basins prevailed landward of transgressing barriers. Only after erosional elimination of presumed preexisting transgressive barriers and marine sediment deposition over brackish (lagoonal) beds did new islands emerge from the seafloor. Before annihilation of preexisting transgressive barriers, their eroded remnants may have become cores of new islands. Should the existence of antecedent transgressive islands be established by reduced-salinity nearshore and lagoonal deposits landward, this may reveal a transgressive phase in the prehistory of Matagorda and Mustang islands.

In contrast with the fewer and smaller NE barrier island and spits, the alongshore continuity and greater dimensions of the NW barriers indicate existence of much larger sediment sources, a variety of active and relict mainland, estuarine, and seafloor lithosomes. Ravinement, cyclonic storm erosion, and landward transport of reworked shelf-margin, mid-shelf relict delta, and shoreline deposits were key to barrier nourishment and maintenance. Fluvial sediment discharge would have been bolstered by runoff-enhancing cool and wet conditions inland during the early Holocene. Semiarid climate episodes resulting in sparser vegetation cover would have marked late

Pleistocene Wisconsinan and certain early-to-mid-Holocene time intervals (Otvos 2004; Anderson et al. 2016). NE and E Gulf shoreline lithosomes, linked to Stages MIS 5a and MIS 3, are widespread in the sediment-enriched NW shelf region. In contrast, only few comparable features were recognized in the NE. The 100-km-long, 5-m-high, 2–6-km-wide shore-parallel South Perdido Shoal, submerged in 35 m deep water near De Soto Canyon, may be also an ancient shore relict (Hine and Locker 2011). Two ~15-km-long, shore-normal shoal trends, located at 8–24 m depth, represent cape-retreat massifs (Swift 1975). They suggest the proximity of mid-Holocene relict headlands. The shoal massifs may indicate successive positions of landward receding Apalachicola Delta shorelines (Flocks et al. 2011).

Brief flooding-backstepping episodes between ~14.5 and 13.8 ¹⁴C ka interrupted smooth sea-level rise during deglacial transgression. By 6.6–5.8 ka, the delta retreated to the present Apalachicola Bay area (Osterman et al. 2009). A few large submerged barrier islands mark the carbonate-dominated eastern Gulf shelf. The 30-km-long, 5-km-wide Pulley Ridge barrier W of the thick Florida carbonate platform occurs at ~58 m depth (Fig. 7.20 in Hine and Locker 2011). Stepped relict outlier reefs located off SW Florida and the coeval beaches may mark marine stillstands at ~–80 m to –60 m (Locker et al. 1996). By ~6.5 ka, sea-level reached and then gradually submerged the highest areas on the NE shelf, including those of present Mississippi Sound and the Louisiana nearshore. Evacuated by ebb-tidal currents and occasional storm tides from the Mobile-Tensaw bayhead delta and Mobile Bay and delivered by westward (longshore) transport along the NW Florida-Alabama mainland shore, the combined sand volumes established a 12-km-wide ebb-tidal delta off Mobile Bay entrance pass (Fig. 1; ► [“Barrier Island Formation and Development Modes”](#) entry, present volume). A 17-m-high, ~11-km-long relict barrier body, isolated member of the LIG Gulfport shore ridge trend, between ~5.5 ka and 4.5 ka became the focal site in initiating the Alabama-Louisiana barrier chain (Otvos 2004, 2018; q.v., ► [“Last Interglacial”](#) entry, present volume). Surrounded by the fast-rising Gulf, the ridge forms the core of present eastern Dauphin Island. The western flank of the Mobile Bay tidal delta was attached to the island’s elevated Pleistocene core area. Due to significantly slower sea-level rise, the rate of westward transmitted littoral drift volumes increased along the new island.

A long, narrow subtidal-intertidal sand platform that covered open marine sandy mud deposits was attached to the downdrift end of eastern Dauphin (vis. ► [“Barrier Island Formation and Development Modes”](#) entry, q.v., present volume). Isolating Mississippi Sound as a lagoon by ~5.2 ka, it added the long central and western sectors to the island, approximately parallel with the Mississippi-Alabama mainland shoreline. The platform reached as far west as the New

Orleans area in SE Louisiana (Otvos 1981b). Flocks et al. (2011) suggested that westward continuation of late Pleistocene scarps beneath east Dauphin had defined other barrier island locations downdrift, as well. However, a substantial open marine and shoal interval separates island deposits from the buried late Pleistocene land surface (see Otvos 1981b, this entry). By the time the islands emerged, such scarps would have been buried under the thick Holocene sediment interval. Instead, in addition to eastern Dauphin’s location, seafloor bathymetry and surf zone position were the key factors that determined future sites of shoal zone and island development during decelerating sea-level rise.

Holocene “Highstand” (Reduced Sea-Level Rise Rate After ~4 ka), MIS 1

Genetic barrier island categories during current “highstand” phase Otvos (2018):

- (1) *Shoal aggradation*. First introduced by de Beaumont (1845) and supported by historical records and drill data. Numerous islands in the Chandeleur chain, Louisiana, Mississippi Coast, Mobile ebb-delta, and W Florida Peninsula serve as examples. Overseas: North Sea, Black Sea.
- (2) *Sand spit segmentation*. In addition to assuming island emergence from breaker bars, Penck (1894, p. 560) also attributed island development to barrier spit fragmentation and consequent inlet formation by subsidence. Hoyt (1967) and Penland et al. (1985) promoted variants of this idea. However, Gilbert (1885) is widely but mistakenly cited as having proposed this formation mode.
- (3) *Isolation of tall late Holocene-Recent shore dunes or other postglacial ridges during transgression* (North Sea, Haage 1899 in Ehlers (1988, p. 23); SE Canada, Ganong 1908; Hoyt 1967). Credible evidence is absent from the Gulf Coast (Otvos, ► [“Barrier Island Formation and Development Modes”](#), encyclopedia article, q.v., this volume). McGee (1890) was misidentified as proponent of this category.
- (4) *Composite islands*. Relict Neogene and Pleistocene topographic highs acting as cores, surrounded by prograding sandy late Holocene shore deposits (e.g., Dauphin Island, AL, several Florida Peninsula and Georgia islands (Davis 1994; Otvos 1981b)).

Northwestern Gulf

Formerly jagged, the NE Texas shoreline became smoother as Holocene sea-level rise decelerated. Already established coastal barriers underwent progradational, aggradational, or transgressive development. Matagorda and Bolivar barrier spits and St. Joseph Island emerged after 4 ka, possibly earlier.

So did Follets and South Padre islands; the only transgressive barriers on the present Texas coast. Due to converging littoral drift, longshore currents, and presence of the oldest modern Texas barriers, three-quarters of the total barrier volume accumulated in central members of the NW Gulf coast barrier chain (Anderson et al. 2014, 2016).

Northeastern Gulf

Alabama-Louisiana Barrier Island Chain

The long offshore sand platform, topped by a sand shoal and barrier island belt, isolated Mississippi Sound as a shallow lagoon. Low-salinity sediments settled on the lagoon's shallowest, mainland-flanking bottoms, flooded only in the final stage of the transgression. Gradually separated from marine waters and exposed to mainland freshwater runoff, salinities steadily declined upward in the lagoonal sediment sequence (Otvos 1981b). The aggradational-progradational development of the Mississippi barrier islands is indicated by high salinity open marine microfaunas in the sandy mud beneath the island-platform lithosomes, even within short distances landward from the island chain. The absence of low salinity lagoonal deposits directly beneath and landward of the barriers refutes suggestions of their transgressive origins. Ancestors of the present barriers likely included narrow island sectors, similar to those attached downdrift to the western ends of Dauphin, Petit Bois, and East Ship. Seaward prograded wider and higher barrier ridge plain sectors, developed over broader platform areas, may have been as prominent as at present. The earliest credible sand shoal dates, 5.3–4.6 ka, obtained from eastern New Orleans sand pits, Point Clear, and Cat Island drill holes (Otvos 1981b, 2005; Otvos and Giardino 2004; Miselis et al. 2014), predate Mississippi-St. Bernard delta growth; encroachment and burial of the western one-quarter of the island chain less than a millennium later.

Eastward prograding delta lobes covered several members of the SE Louisiana island chain between ~4.0 ka and ~2.9 ka. An oyster reef, enclosed in prograded brackish Pearl Delta deposits that surrounded and partially buried relict Campbell and Point Clear barrier islands, dated 3.1–2.9 ka (Otvos 1997; Otvos and Giardino 2004). Terminating westward littoral drift from Ship Island by shoal development that reduced wave energy, the advancing delta lobe stopped sand transmission to the west Mississippi barriers, including Cat, Point Clear, Campbell islands, and few smaller nearby islets in the chain. Present-day Square Handkerchief and Cat Island shoals (Fig. 1) according to drill data represent defunct remnants of such islands (Otvos and Giardino 2004). Tropical storms often severed and eliminated long, narrow, low-lying island sectors in the island chain. The “Twin Mobile hurricanes” of 1740 permanently isolated Petit Bois from W Dauphin. Cyclonic storms

repeatedly dissected Ship Island until Hurricanes Camille (1969) and Katrina (2005) rendered separation of the two Ship islands apparently permanent. Immediate, if limited, post-hurricane recovery, nourished by locally concentrated sand resources, tends to be rapid (Fig. 2), while long-term island shrinkage in E. Ship reached critical proportions (Otvos and Carter 2013; Eisemann et al. 2018). Unlike conditions in the sand-rich NW Gulf nearshore, cross-shelf sediment transfer from reworked relict delta and shoreline lithosomes, such as the “Lagniappe” Delta, apparently failed to mitigate storm reduction and restore island growth to previous levels.

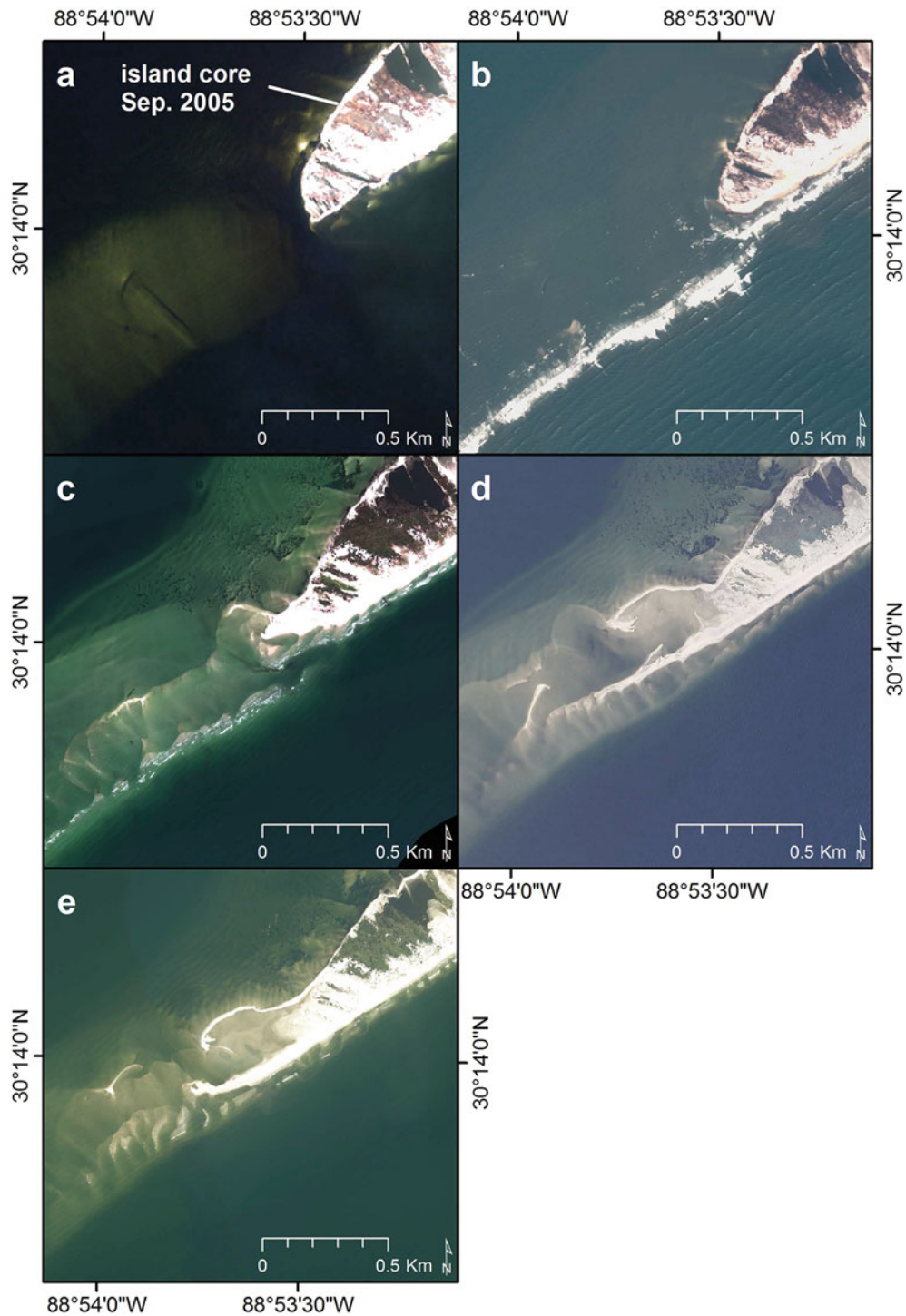
Mainland Ridge Plain Development, NE Gulf Coast

The Alabama-NW Florida shore sector between the Mobile Pass-Fort Morgan and Gulf Beach-Big Lagoon (Fig. 1b) is a unique example of significant late Holocene mainland plain progradation. Sediment supply-driven “normal regression” created a ~64-km-long, ~2–3-km-wide strand plain complex (Fig. 3). Gulfward progradation of strand plain ridge sets started from the late Holocene mainland shoreline at the seaward toe of the shore-parallel LIG Gulfport barrier plain. Erosional boundaries outline and subdivide several beach ridge sets (Fig. 4). Critical evaluation of OSL dates (Otvos 1981b, 2010, 2018; Blum et al. 2003; Rodriguez and Meyer 2006; Donnelly, written com. 2016) suggests that the Holocene ridge plain belt formed between ~5.4 ka and 2.1 ka. A similar late Holocene addition to the mainland coastal is *SW Louisiana's Chenier Plain* (q.v., ► “Cheniers” entry) that started ~3 ka (Owen 2008). Marking an abrupt local change from depositional to an erosional regime, a straight, sharp line truncates the ridge plain behind the present beach (Figs. 3 and 4).

Santa Rosa, in appearance similar to the very narrow, spit-like Alabama-Mississippi barrier sectors, experienced no long-term progradation, aggradation, or shoreline erosion-related permanent land loss. Its interior dune fields severely impacted by storms, the island outline did not change significantly (Stone et al. 2004; Wang et al. 2006). Despite its vulnerability, this 64-km-long, mostly 300–500-m-wide, low island remained stable over its recorded history. Except in its prograded, curved long western tip, it lacks strand plain topography. An unusually narrow lagoon backs the island (Otvos 1982).

Apalachicola Delta Coast Barriers, Northwest Florida

In sharp contrast to often ephemeral, rapidly disintegrating, landward retreating delta barriers, formed from fine sands of the overwhelmingly muddy Mississippi River deposits, sand-enriched Apalachicola River sediments nourished more stable, if fewer, prograding barriers. Ravinement erosion and onshore sediment transport provided additional sand sources. Submerged cape-retreat shoal ridges and late Pleistocene

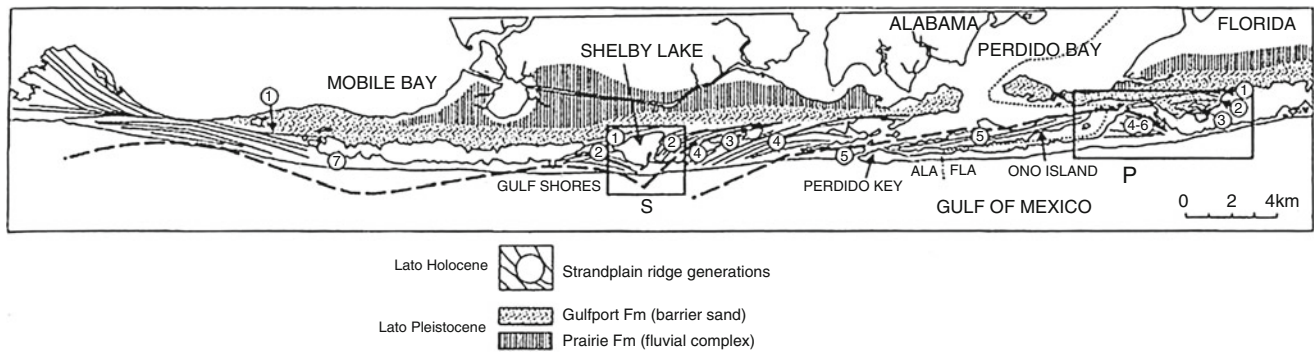


Gulf Shorelines, Last Eustatic Cycle, Fig. 2 Aggradational post-Katrina recovery on shoal platform, SW East Ship Island. Recovery and island reemergence mimics vertical aggradation origins and subsequent lateral growth of numerous barrier islands. Note stages of berm

basin formation and sand infill. (a) September 13, 2005; (b) February 2, 2006; (c) November 13, 2007; (d) August 24, 2009; (e) July 28, 2010. IKONOS and Quickbird satellite images (Otvos and Carter 2013)

coastal sand bluffs, composed of interglacial and glacial-stage sand deposits in mainland barrier ridges, represented abundant sediment sources for the Holocene barriers. Two large and two smaller barriers form the group (Fig. 5). They include

St. Vincent Island, the most massive barrier islands on the NE Gulf. Erosional truncation lines outline individual ridge sets (Fig. 6). From ~2.7 ka until recently, island progradation was intermittent.



Gulf Shorelines, Last Eustatic Cycle, Fig. 3 Holocene mainland strand plain complex, SE Alabama-NW Florida (Otvos 1997)

Minuscule “Richardson Hammock” strand plain island emerged between 3.1 ka and 2.9 ka and is the oldest component of St. Joseph barrier.

Prograding northward, “St. Joseph Island” gradually absorbed additional islets. By 1.0 ka, the island was ~14 km long. After a short, wide sand spit linked St. Joseph Island to the mainland shore ~600 yr. ago, this slender island was converted to a barrier spit (Otvos 1992, 2005, 2012, 2018; Rink and López 2010).

Mississippi Delta Complex Barrier Islands

The Holocene Mississippi Delta Complex in southern Louisiana formed numerous subdeltas since ~6–7 ka. Following avulsion of the trunk channels, each subdelta created several delta lobes (Coleman et al. 1998; Roberts 1997). After river discharge ceased, channel switching resulted in abandoned delta lobes. Steady retreat of the Gulf shoreline and expansion of delta plain lakes during late Holocene transgression, caused by wave erosion, tectonic, and compactional subsidence (Wolstencroft et al. 2014), had accelerated due to the low resistance of unconsolidated mud, peat, and fine-to-very-fine-grained sand deposits to wave erosion. Three major island groups have evolved.

South Louisiana Barrier Islands In the course of their historically documented landward progradation, Timbalier and several other south Louisiana barrier islands lost most of their area to wave erosion and delta subsidence (Williams et al. 1992; q.v. ▶ “Barrier Island Formation and Development Modes”; ▶ “Coastal Barrier Preservation and Destruction”; ▶ “Eustasy and Sea Level” entries, this volume). Two islands, Grand Terre and Grand Isle, attained 11.5–14 km lengths and are 1.3–2.5 km wide. Islands formed at sites where the muddy river discharge already ceased or did not interfere with wave energy, littoral drift, and progradation. Cheniere Ronquille and E. Timbalier islands (Ritchie et al. 1992, pp. 13–14; Penland et al. 2003; Kindinger et al. 2013; Flocks and Terrano 2016),

and certain members of the Isles Dernieres (Last Islands) chain, as well as Timbalier Island, Grand Terre, and Grand Isle display ridge plain morphology that prove their aggradational-progradational evolution. These processes and landforms defined true barriers, not marsh island remnants.

The Last Islands (Isles Dernieres) chain (Fig. 1) that fronted the Lafourche-Grand Caillou delta headland includes several irregularly shaped marsh island remnants (Whiskey Island). Relict tidal creek channels preserved in their morphology were inherited from the original delta plain. The low islands represent ever-diminishing, beach sand-veneered frontal fragments of Isle Derniere, originally a ~50-km-long marshy delta plain island. Due to sediment compaction and shore erosion, by 1853 Isle Derniere marsh island was separated from the mainland delta plain by coalescing delta lakes, bays, during regional subsidence that significantly raised relative sea-level in southern Louisiana. Before 1853, Derniere was separated from the mainland delta plain by the erosion of coalescing delta lakes, by bays in its rear, and by subsidence due to sediment compaction subsidence (Fig. 1; Williams et al. 1992, p. 39). A few sectors that display progradational ridge topography (e.g., parts of Trinity Island) may have emerged as barrier islands off the shores of the fragmented delta marshland from reworked shoal sands or prograded from the shores of fractured marsh islands. Western, respectively eastern sand spits of original large Derniere (Last) marsh island were underlain by somewhat thicker barrier sand intervals. By 2010, Wine became a sand shoal. Greatly diminished, breached, Raccoon is expected to follow soon. Drastically reduced in past decades, the remaining islands also face disintegration. An ambitious effort by the Coastal Protection and Restoration Authority (CPRA 2017), the planned restoration of marsh and dune vegetation, as well as island rebuilding by sand nourishment and grass planting is underway in an attempt to reverse island decay.

Four barriers flank the Moreau-Caminada headland (Fig. 1). SW-ward diverging littoral drift nourished Timbalier

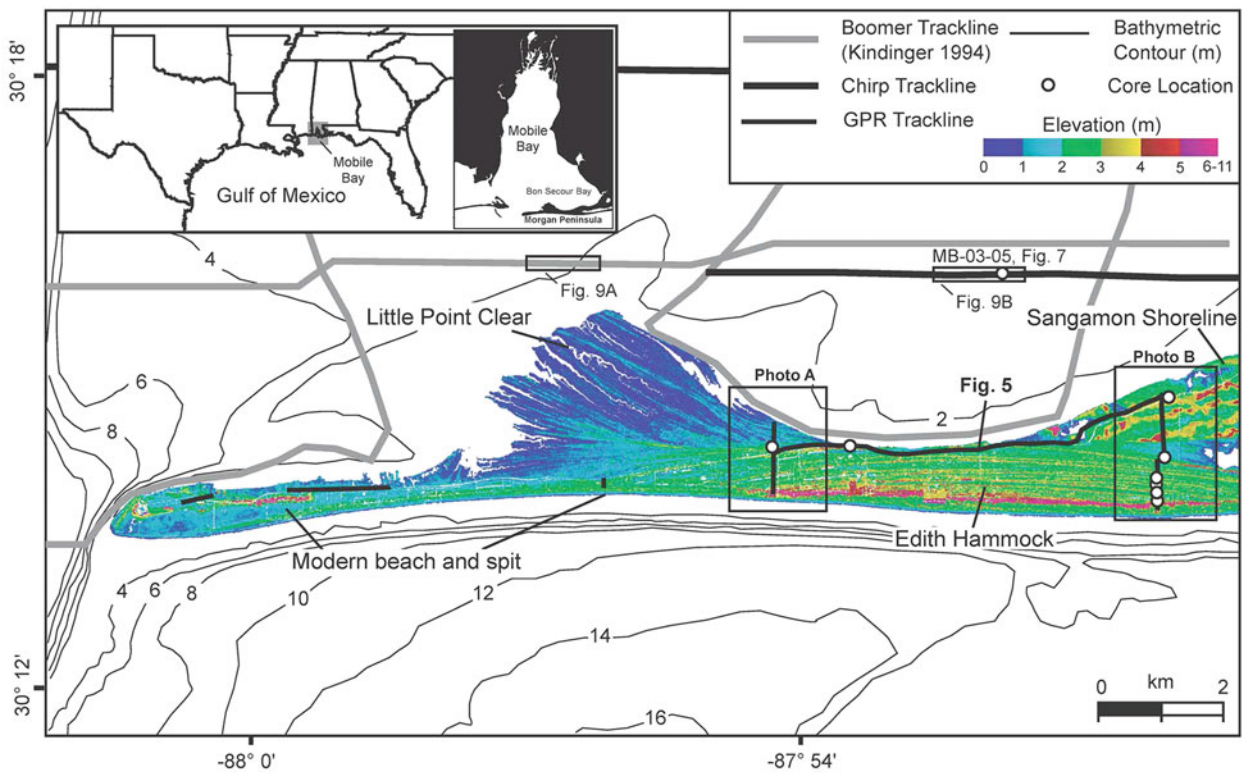
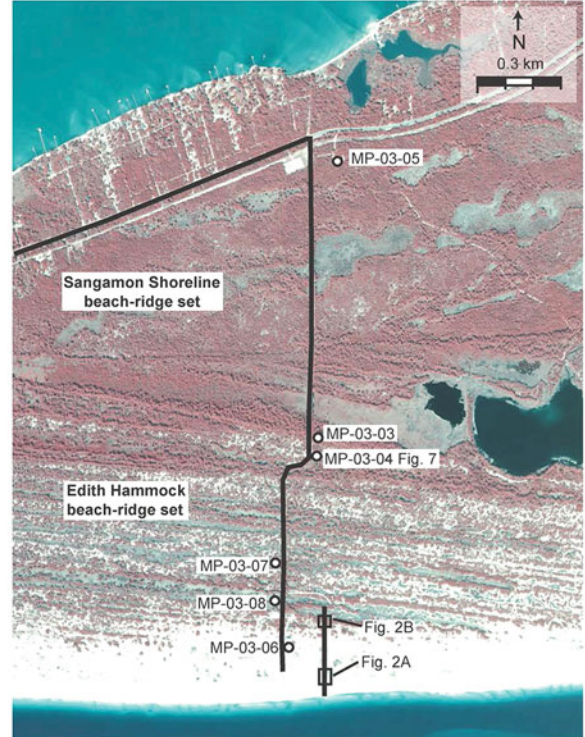
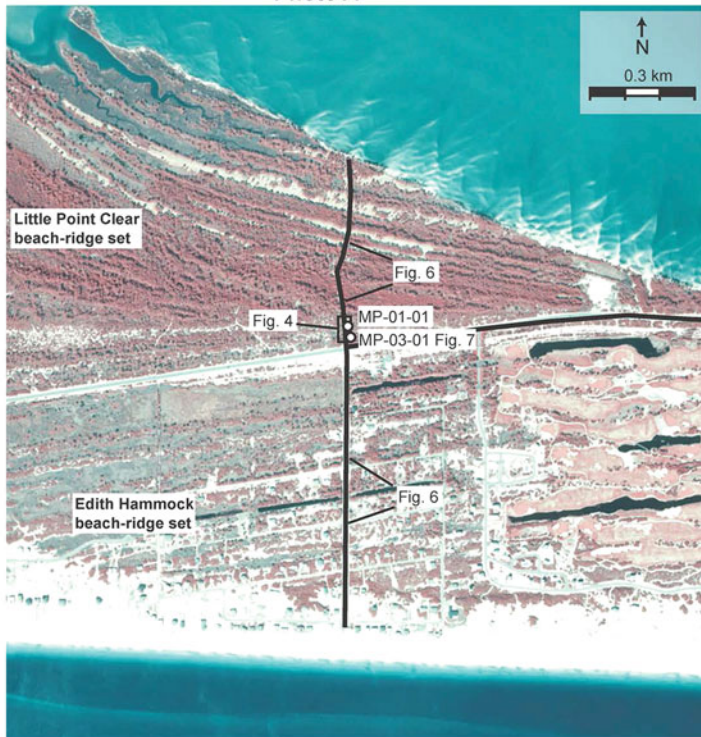


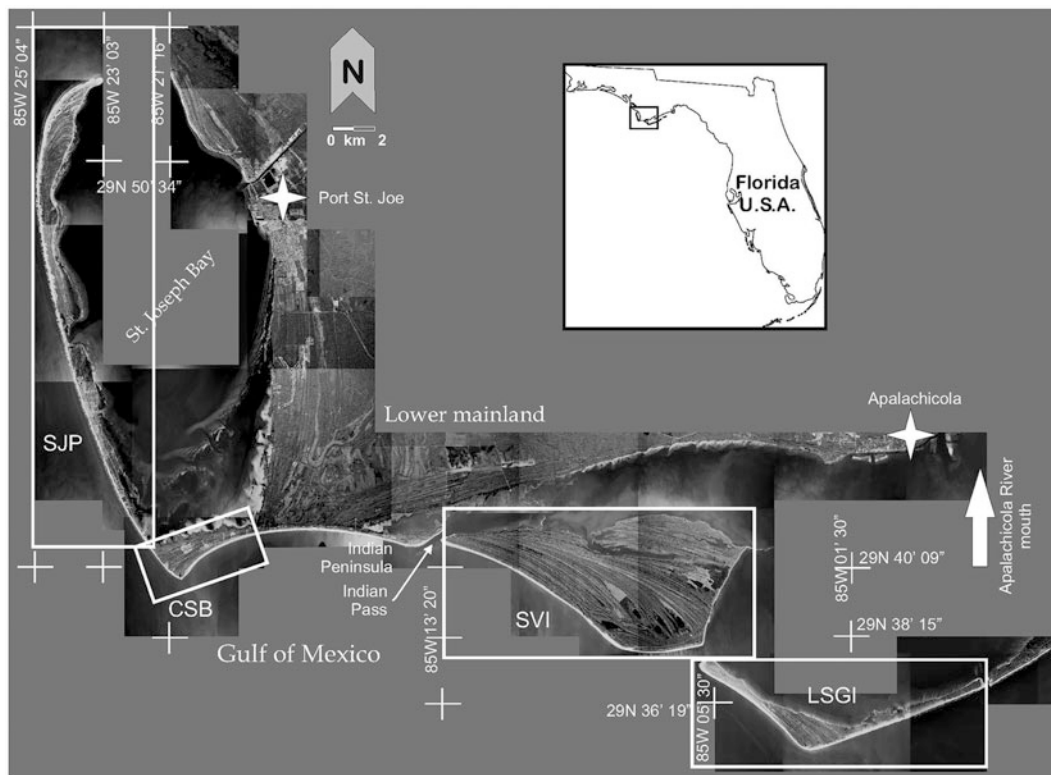
Photo A

Photo B



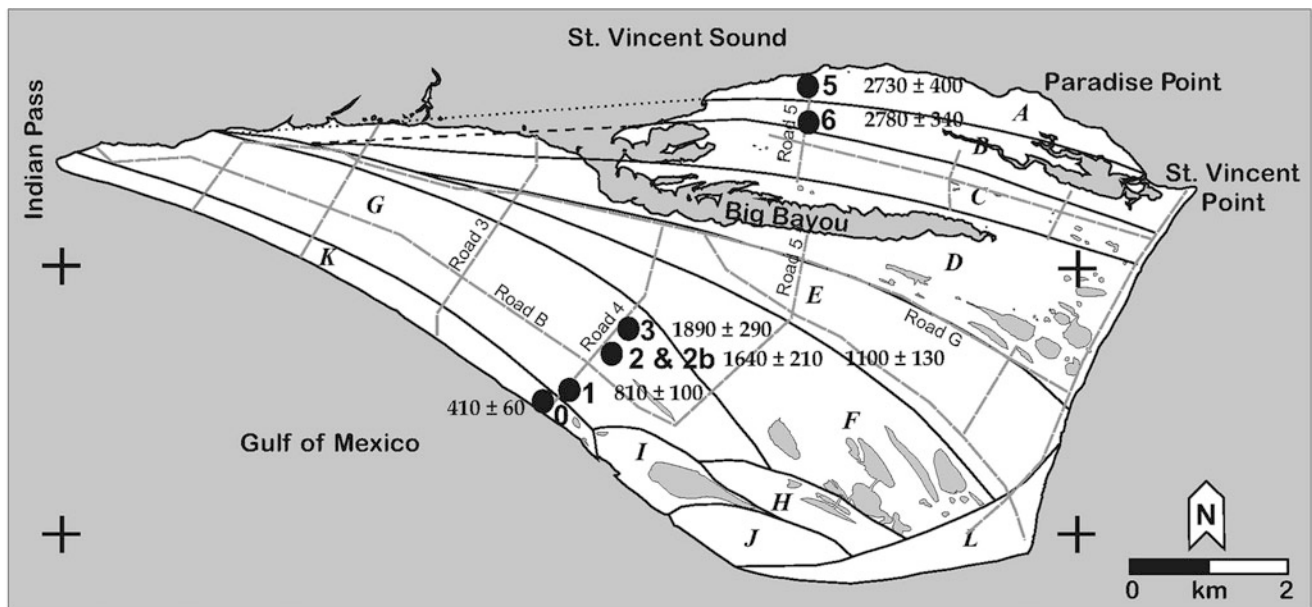
Gulf Shorelines, Last Eustatic Cycle, Fig. 4 Middle and late Holocene Little Point Clear and Edith Hammock strand plain sets, southeast Alabama. Sharp separation between Last Interglacial strand plain (top; Gulfport Formation ridge set; “Sangamon Shoreline beach ridge-set”)

and middle to late Holocene ridge sets. Mobile Bay: top left (Courtesy of Dr. A. Rodriguez, see Rodriguez and Meyer (2006)). (See, Otvos 2018)



Gulf Shorelines, Last Eustatic Cycle, Fig. 5 Apalachicola Coast barriers. *SJP* St. Joseph barrier spit; *CSB* Cape San Blas relict E-W barrier spit; *SVI* St. Vincent Island; *LSGI* – Little St. George and

St. George islands. “Lower mainland”: Last Interglacial Gulfport Barrier strand plain (Courtesy Rink and López (2010))



Gulf Shorelines, Last Eustatic Cycle, Fig. 6 OSL-dated St. Vincent Island strand plain sets, Apalachicola Coast, NW Florida ridge sets A-through-L. Progradational strand plain generations, separated by erosional lines of truncation. Extensive swale pond development

characterizes the north due to slight compactional subsidence and minor eustatic sea-level rise during the last three millennia (Courtesy Rink and López (2010))

and E. Timbalier islands, while the NE-directed, probably larger drift volumes nourished the somewhat sturdier, relatively less vulnerable Grand Terre and Grand Isle barrier islands. Most of the recently heavily eroding headland consists of strand plain ridges that formed between ~ 0.7 ka and ~ 0.4 ka. Progradation of this sizable headland strand plain continued until ~ 0.3 ka (Gerdes 1982). Allochthonous dark “coffee-ground” beds (Coleman 1966), composed of grains of carbonized plant detritus, interlayered with fine, very well-sorted fine sand and silty-sandy strata, uniquely characterize a beach ridge plain in the peat-enriched Mississippi delta complex. Reworked organic matter originated in much older St. Bernard and other delta lobes of the interior.

Concentrated and deposited on the foreshore of a Gulf beach, alternating layers of clean sand, muddy sand, and coffee-ground layers were incorporated into the lithologically unique sediment ridge plain sequence. Indicating reworking and retransportation, the detrital organic layers dated two-to-three millennia older than the enclosing fine sand beds. One swash-concentrated layer composed of carbonized grains that covered a modern Gulf beach foreshore yielded a 250 ± 60 -year date (Otvos 1969; Gerdes 1982). Lafourche delta sediments accumulated in the western headland as late as nearly a century ago. Intensive mainland shore retreat of up to 2.5 km took place in the past century. The four flanking strand plain barriers, including Grand Isle (Conatser 1971; US Army 1972), and much more vulnerable Grand Terre apparently also formed by nearshore shoal aggradation off the Lafourche delta lobe. No indications of the existence of previous spit-connections to the mainland and other island shores are known here either.

Chandeleur-Breton Island Chain At its record extent, the eastern limit of St. Bernard delta complex (“subdelta”) was located ~ 72 km east of its first accurately mapped mid-nineteenth-century position in SE Louisiana. The ancestor of the ~ 75 – 80 -km-long Chandeleur chain may have formed offshore from the retreating subdelta shoreline by reworking of sandy fluvial and tidal channel fill, sand bar, shoal, and marine nearshore deposits. Fine sands from these facies were concentrated and incorporated into newly emerging shoals that gave birth to a series of emerging barrier islands. The islands underwent westward-directed transgressive migration at the heels of the rapidly retreating, decaying subaerial delta plain shoreline, gradually replaced by landward expanding wide lagoons (Otvos and Giardino 2004). The thickness of the subtidal sand shoal deposits that transgressed landward over lagoonal muds varies between 0, where lagoonal muds outcrop in the beach face (Ritchie et al. 1992), and 5–10 m, dominantly in the north and center (Frazier 1967; Penland et al. 1985; Otvos 1986). Degradation of the Chandeleur chain (Figs. 1 and 7) progressed faster in the south, endowed by lesser delta and modern offshore sand resources.

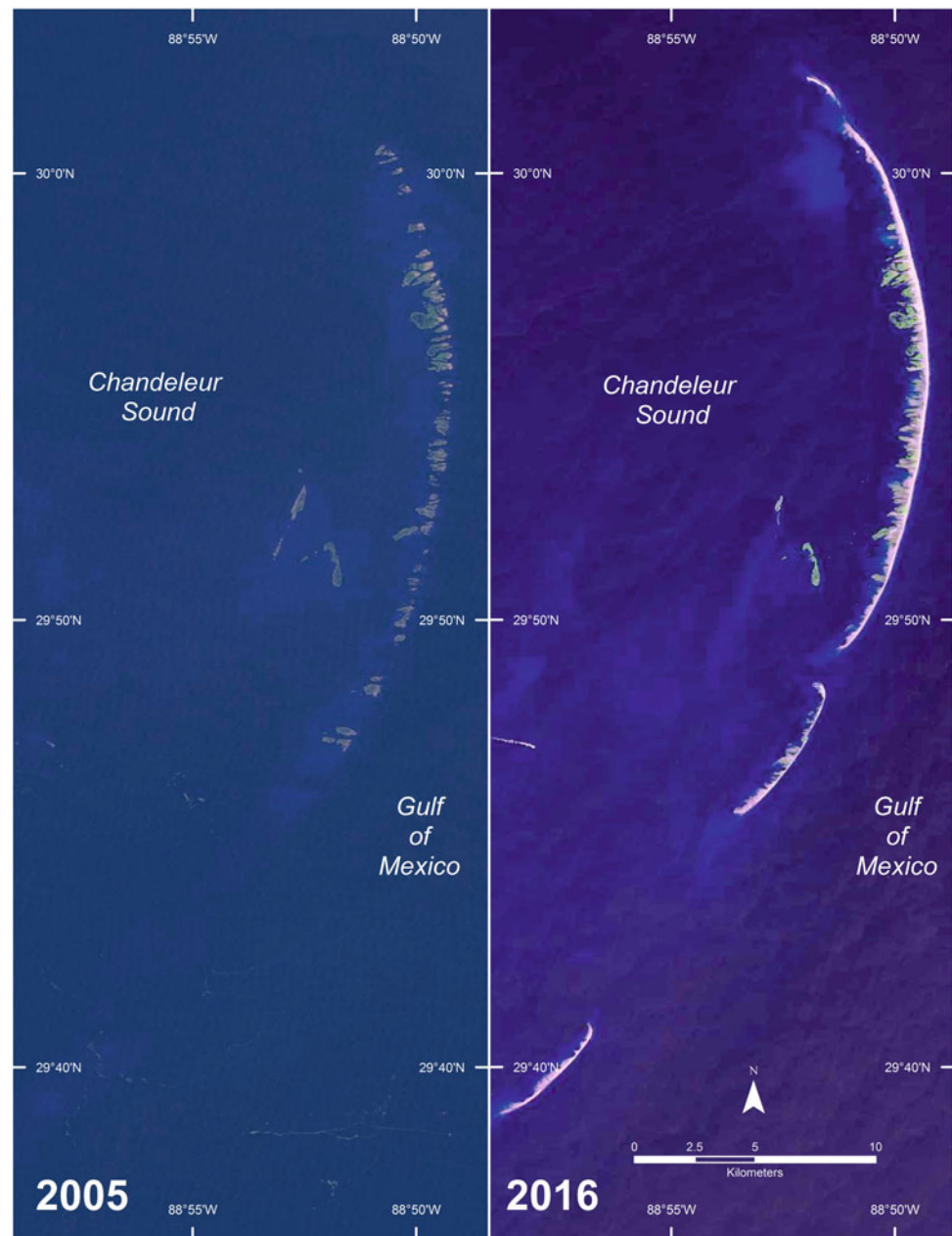
Transgressive retreat of the delta front island chain occurred during periods of storm destruction that alternated with calmer weather intervals of barrier reemergence. Episodes of frequent and intensive tropical storms, followed by calmer intervals, have repeatedly reduced islands to subtidal shoals, respectively, converted shoals to islands (e.g., Grand Gosier, Stake, Curlew islands). Until early in the twentieth century, storm overwash drove “tank tread-rollover” style landward island retreat. Islands often reemerged landward of their pre-storm positions (Otvos 1981, 1986; Williams et al. 1992; Otvos and Giardino 2004; Fearnley et al. 2009; Flocks et al. 2009; Otvos and Carter 2013; Otvos 2018). This process somewhat resembled transgressive overstepping in non-deltaic islands. Between 1855 and 1989, Gulf-shore recession along the northern and central islands ranged from 22 m to 2.2 km. In the small southern islands, total retreat distance averaged ~ 1.3 km. Concurrently, the lagoonal shores of the island chain receded by 264 m to 2.0 km (Ritchie et al. 1992; Williams et al. 1992).

Between 1855 and Hurricane Katrina (2005), the combined northern and central island area declined from 44.5 km^2 to 4.7 km^2 (Fearnley et al. 2009; Flocks et al. 2009; Rogers et al. 2009). Following unprecedented area reduction during and briefly after the 2005 and 2008 hurricanes, by 2016–2017 the central and N island sectors returned to a high level of lateral continuity (Fig. 7), comparable with the storm-free years between 1998 and 2001.

Island Evolution Model, Mississippi Delta Complex

An often-cited three-stage island evolution scheme (Penland et al. 1985) assumes the formation of shore-parallel barrier spits from eroding delta shore headlands. Wave-breaching would convert barrier spits into island chains. In the proposed field example, the two opposite flanks of the disintegrating *Last Islands* marsh headland grew narrow terminal barrier spits that later split into a series of islands (Fig. 1). Originally spits grew from the opposite flanks of the Derniere headland. Raccoon Island, in the W, was ~ 7 km and Wine Island, on the E, ~ 4 km long. The spits never extended beyond the headland area (Williams et al. 1992). As Penland et al. also recognized, instead of expansion, the assumed step toward the “island arc” stage, the islands are fast disintegrating. By 2012, Wine became a sand shoal, and re-breached Raccoon was reduced to but a fraction of its original size. Instead of spit growth, expected to result in a barrier arc, both the prograded barriers and the intermixed marsh island fragments in the chain face rapid elimination. Deposits in the low delta plain headlands usually contain limited volumes of sand, mostly in relict stream, tidal channel, and sand bar lithosomes. Peaty-muddy deposits predominate. Sand volumes, freed by headland erosion and onshore supply, were insufficient for the construction of sturdy, long barrier spits.

Gulf Shorelines, Last Eustatic Cycle, Fig. 7 Comparison images of Chandeleur Island-Shoal Chain Post-Hurricane Katrina Recovery. 2005 (Sep. 16): immediate post-Hurricane Katrina; (Landsat Thematic Mapper), subdued gray areas are shoals with few surviving brighter-toned islets. 2016 (Sep. 30): island reemergence following extended fair-weather interval (Landsat 8 Operational Land Imager), images of Northern and Central Chandeleur Chain



No historical documentation or other evidence, only speculative assumptions, exists in the nearby *Moreau-Caminada headland* area (Fig. 1) for the existence of headland-attached barrier spits that, following storm breach, would have given birth to four flanking barriers (Otvos 1986, 2018). Equally important and refuting the applicability of the island evolution model (Kulp et al. 2016; Kramer 2016), the headland is coeval with and postdates its flanking islands. Lafourche subdelta growth and strand plain ridge progradation that formed the headland continued until recent reversal to a robust erosional regime. Before Bayou Lafourche's discharge was blocked in 1904, the western headland area continued to expand (Gerdes 1982). A simpler and more plausible genetic

alternative invokes longshore and onshore sand supply along the mainland shore, mediated by tidal deltas across intervening inlets and passes. The resulting islands also emerged from sand shoals and bars. Sufficient sand volumes from updrift and offshore sources enabled progradational expansion of all four ridge plain islands.

The *Chandeleur island chain* (Figs. 1 and 7) was the preferred example for the second stage in the island evolution model, conversion of headland-flanking barriers into island arc. During erosional retreat of the mainland shore, aided by subsidence, the distance between barrier spit-derived barrier islands and the disintegrating, sinking delta plain increased; the developing new lagoon (Chandeleur-Breton Sound)

steadily widened. The earliest Chandeleur generation may have formed off the easternmost delta plain shoreline during the greatest gulfward expansion of the active delta lobe. However, based on documented island history, one may assume that the chain periodically was reduced to a series of submarine sand shoals in the course of its retreat. Subsequent aggradation during fair-weather intervals raised the islands above sea-level. As the mainland shore and the transgressive islands retreated toward the Louisiana mainland, the chain underwent punctuated episodes of storm overwash-driven landward shift. Only the deepest shoal lithosomes, “island roots” (Penland et al. 1985), may have survived migration and massive reworking of the subbottom deposits. The combination of deep ravinement and seafloor erosion by intensive tropical storms eliminated all tangible details of island prehistory; the higher shoal and barrier intervals fell victim to prolonged erosion (Otvos 1986).

The gently arcuate island chain outline is unrelated to late Holocene delta shoreline, including headland positions. The shapes of the island chains result from the interaction between seafloor bathymetry and refracted wave set characteristics, as well as island modification by storm erosion and overwash such transgressive episodes that drove the barrier sands over island and lagoon deposits into Chandeleur-Breton Sound.

Summary

Differences in sediment sources, volumes, and erosion mechanisms, variable off- and onshore transport, and deposition processes explain major contrasts in landform development and history between the NW and NE Gulf shores during five phases of the Last Eustatic Cycle. Direct means of sediment delivery included river discharge, ebb-tidal evacuation from estuaries, and bluff erosion along the mainland shore. Sediment reworked in and transported landward from relict shelf deltas, ancient shore lithosomes, and stream valley fill was the major indirect source of new coastal lithosomes. Vertical and horizontal shoreline fluctuations driven by sea-level and sediment supply variations, changes in climate, fluvial, and estuarine processes and the dimensions of associated drainage basins played key roles. The pronounced NE vs. NW asymmetry in terms of barrier dimensions and development, closely linked to the volumes of direct and secondary onshore and offshore sediment supply, is the direct result of the far greater dimensions of paleodeltas and relict shoreline lithosomes that formed on the NW Gulf shelf throughout LEC. Only few major submerged coastal landforms, including barriers and paleodeltas, have been conclusively identified in the NE region. Several modern shore barriers predate the current highstand stage of slowed sea-level rise, while most NE islands and several large barrier spits, including St. Joseph and Morgan peninsulas, formed since ~4 ka. The

terms *transgressive* and *regressive*, as well as their relationship to island “rollover” and progradation require clarification in the context of island migration modes.

Landward-shifting elongated transgressive delta barriers were intermittently driven over their lagoons by storm overwash (e.g., North Chandeleur Island, LA). The regressive strand plain (beach ridge plain) barriers, nourished by abundant littoral drift, prograde seaward or parallel with the mainland shore. They display ridge-and-swale topography. Building wide ridge plains, St. Vincent, Cat, and Galveston barrier islands, Bolivar, and Matagorda barrier spits underwent extensive seaward progradation. In addition to transgressive islands, regressive barriers may also actively advance toward the retreating, slowly transgressed mainland shore. Thus, St. Joseph barrier spit, NW Florida (Figs. 1 and 5), and the Mississippi-SE Louisiana barrier chain, its evolution greatly impacted by fluvial and tidal delta development, prograded at right angle toward the mainland shorelines. Strand plain islands along the southern fringe of Lafourche subdelta shifted at an oblique angle toward the mainland shoreline (Williams et al. 1992).

While the initial formation mode of islands that subsequently became transgressive is unknowable, apparently most modern barriers have aggraded from submarine sand shoals. Among late Holocene barriers, landward prograded regressive, strand plain islands dominate over transgressive ones and over rare exclusively vertically aggraded barriers. Several Gulf islands, established around and over pre-Holocene, even pre-Quaternary island cores (Dauphin, several W Florida Peninsula barriers) belong to the composite barrier category (Otvos 1981b; Otvos and Carter 2013; Davis et al. 2003). Originally ~140 km long, this longest barrier chain on the E Gulf Coast was initiated on a narrow, long submarine shoal platform, formed after ~5.2 ka over open marine sandy muds. Reduced to ~110 km length by eastward-advancing Mississippi delta lobes, western members of the chain soon were rendered inactive by shoaling. Several islands were partially or completely buried under delta deposits, others reduced to shoals. In part also nourished by erosion of relict shelf delta lithosomes, the modern Apalachicola Coast barriers underwent only progradation. None experienced transgressive rollover. Somewhat similar in nature to the subrecently prograded, mud- and seashell-dominated Louisiana Chenier plain, the SE Alabama-NW Florida strand plain belt reflects the complex evolution of a sizable mid-to-late Holocene coastal plain area.

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Headland-Bay Beach

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A headland is defined in common language as: (1) a point of usually high land jutting out into a body of water: promontory; (2) high point of land or rock projecting into a body of water. Therefore, a headland-bay beach is a beach whose shape is mainly conformed by the fact that it is located between such headlands, or at least adjacent to one. Some of the synonymous terms that can be found in literature to describe a headland-bay beach are: bay-shaped beach, pocket beach, zeta bay, bow-shaped bay, and half-heart bay. This type of feature shows a gradually changing curvature which Krumbein (1944) noted resembled that of a logarithmic spiral curve.

Johnson (1919) gave an incisive description of wave refraction caused by headlands along an embayed coastline, and Krumbein (1944) showed a simplified diagram of wave refraction into a bay lying to the lee of a headland. According to Yasso (1965), a headland was considered to be any natural or artificial obstruction that extended seaward from the coastline and caused a change in some element of the coastal wave pattern because of its presence.

Historical Development

Observations of headland beach morphology in nature Krumbein (1944) was the first author to describe a beach planshape—Halfmoon Bay, California (USA)—as being similar to the increasing radius of curvature found in the logarithmic spiral, although the restricted classification was removed some years later and the paper was reprinted

in 1947. Yasso (1965) selected four US beaches for testing goodness of fit to the logarithmic spiral approximation. Poles for three of the best-fitting logarithmic spirals were located in close proximity to the seaward end of each headland. The spiral angle α was found to range between 41.26° and 85.64° . Berenguer and Enriquez (1988) reviewed data from 24 beaches around Spain and derived empirical correlations between geometrical characteristics of the layout of the offshore breakwaters and beach planshape. Moreno and Kraus (1999) performed fittings of analytical planshapes to a data set of 46 beaches in Spain and the United States and derived preliminary engineering design guidance including the proposal of a new functional shape.

Observations of headland beach morphology in the laboratory Silvester (1960) tracked the time evolution of a beach in a physical model and observed that the beach between headlands tended to reach an equilibrium shape—logarithmic spiral shape—in response to persistent swell directions under a certain wave angle of attack. The model coastline was allowed to erode without replenishment of sand at the updrift end. Yasso (1965) pointed out that the conditions of Silvester's physical-model test (lack of continuous sediment supply to the updrift end of the model and close spacing of the headlands) suggested that the equilibrium form achieved in the model may not be identical to that achieved under natural conditions of sediment supply and wider separation of headlands.

Silvester (1970) performed additional model tests in which three different wave conditions were generated at three different angles of incidence to the alignment of a headland on an initially straight sandy beach. It was observed that the coastline developed three distinct curvature zones: first, a near circular section in the lee of the upcoast headland; second, a logarithmic spiral; and finally, a segment tangential to the downcoast headland. The time evolution of the spiral constant angle was plotted for the three incident wave angles tested, but only one wave condition was run long-enough as to see an asymptotic trend. A graphical linear relationship between the

logarithmic spiral constant angle α and the wave angle β was provided based on the three angles tested.

Spătaru (1990) studied Romanian Black Sea beaches subjected to normal wave incidence by means of physical models. Equilibrium beach planshapes were considered to be described by arcs of circumferences, and provided design guidance based on geometry.

Numerical Model Approaches

Mashima (1961) constructed wave-energy diagrams from wind-rose diagrams and studied the configuration of stable coastlines based on energy considerations. When the wave energy is a semi-ellipse, the configuration of the stable coastline is approximately parabolic on which the tangent direction at the apex of the parabola coincides with the direction of the major axis of the wave-energy diagram.

LeBlond (1972) attempted to study how wave-induced longshore currents in the presence of a headland could erode a linear beach by developing a numerical model. LeBlond (1972) stated that if there existed a planimetric shape which the headland beach asymptotically approached, it must have the following properties: (1) it should be concave outwards, near the headland, and then convex outwards. (2) the sand transport should increase monotonically along it. (3) erosion, by causing the beach to be displaced normal to itself, should not qualitatively change the shape of the beach. LeBlond also pointed out that the logarithmic spiral did not satisfy the first condition because it is always concave outwards, and that there may be other curves satisfying all of the above three conditions, and one could not decide “a priori” which one will be the equilibrium one.

The main modification implemented by Rea and Komar (1975) with respect to previous numerical modeling efforts, was the combination of two orthogonal one-dimensional grids to simulate beach configurations, so that beach erosion could proceed in two directions without the necessity of a full two-dimensional array. Testing of the model in a hooked beach coastline configuration indicated that the coastline would always attempt to achieve an equilibrium configuration governed by the pattern of offshore wave refraction and diffraction and the distribution of wave-energy flux.

Walton (1977) presented an analytical model to describe the equilibrium shape of a coastline sheltered by a headland using a continuous wave-energy diagram consisting of representative offshore ship wave height and direction observations. It was found to produce coastline shapes similar to the logarithmic spiral shape for sheltered beaches in Florida. The model worked by establishing that the coastline orientation at a certain point was normal to the average direction of the so-called energy of normalized wave attack—that is,

the energy which is allowed by the headland to reach the shore.

Yamashita and Tsuchiya (1992) constructed a numerical model for three-dimensional beach change prediction to simulate a pocket beach formation. The model consisted of three modules to calculate waves, currents, and sediment transport and beach change. The wave transformation module was based on the mild-slope equation of hyperbolic type; the current module was horizontally two-dimensional with direct interaction with sea-bottom change, which was evaluated by the sediment transport model formulated by Bailard (1982) in the third module.

In a theoretical work on the subject of headland-bay beaches, Wind (1994) presented an analytical model of beach development, where the shape of the headland-bay beach remained constant with time and expanded at a rate according to a time function. Wind's (1994) conceptual framework was based on knowledge of the existence of a headland-bay beach shape centered around a pole and that evolved in time in a more or less constant shape. If the position of the coastline is described by the radius r , the angle δ , and time t , the evolution of the coastline with a constant shape implies that the coastline might be described as

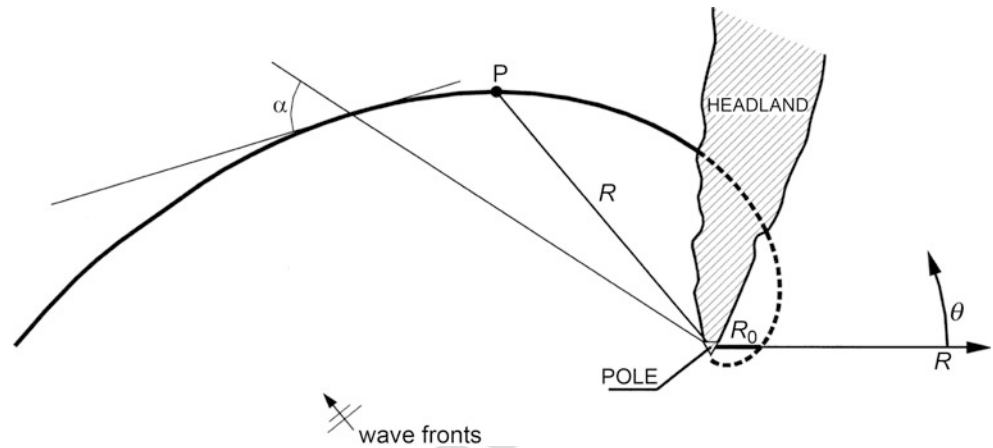
$$r(\delta, t) = r_0 f(\delta) e(t) \quad (1)$$

where r_0 is the constant, $f(\delta)$ is the shape function, and $e(t)$ the evolution function of the coastline in time. With respect to the time function, it was shown that in the diffraction zone it should follow a $t^{1/3}$ law, whereas for the refraction zone a $t^{1/2}$ law was found. This implied that the evolution of a headland-bay beach in the diffraction zone should initially be faster and on the long term, slower than the evolution in the refraction zones. With respect to the shape, the function shape $f(\delta)$ is expressed in terms of functions representing the diffracted wave field. The logarithmic spiral is obtained by taking the functions for the group velocity and the geometrical part for the driving force as constants.

Equilibrium Planshapes of Headland-Bay Beaches

Three functional shapes have been proposed to describe the equilibrium planshape of headland-bay beaches, namely the logarithmic spiral shape, the parabolic shape, and the hyperbolic tangent shape.

The *logarithmic spiral* (also named equiangular, or logistic spiral), first described by Descartes, was described as the curve that cuts radii vectors from a fixed point O under a constant angle α (Fig. 1). The equation of the logarithmic spiral can be written in polar coordinates as

Headland-Bay Beach,**Fig. 1** Definition sketch of the log spiral planshape

$$R = R_0 e^{\theta \cot \alpha} \quad (2)$$

where R is the length of the radius vector for a point P measured from the pole O , θ is the angle from an arbitrary origin of angle measurement to the radius vector of the point P , R_0 is the length of radius to arbitrary origin of angle measurement, and α the characteristic constant angle between the tangent to the curve and radius at any point along the spiral. The pole of the spiral is identified as the diffraction point (Silvester 1960; Yasso 1965), and the characteristic angle of the spiral is a function of the incident wave angle with respect to a reference line. For headlands of irregular shape and for those with submerged sections, the diffraction point cannot be specified unambiguously, a problem entering specification of all equilibrium shapes. The reference line extends from the approximate location of the diffraction point to a downdrift headland. This shape is extremely sensitive to variations in the characteristic angle α because the angle enters the argument of an exponential function (Moreno and Kraus 1999), being the practical consequence that α has to be accurately defined. For engineering application four unknowns must be found: location of pole (two coordinates), characteristic angle α , and scale parameter R_0 . The shape of the log spiral is controlled only by α , with the parameter R_0 determining the scale of the shape. In fact, the functioning of R_0 is equivalent to setting a different origin of measurement of the angle θ . In other words, graphically the log spiral may be scaled up or down by turning the shape around its pole.

Values of α for headland-bay beaches reported in the literature range from about 45° to 75° . As α becomes smaller, the log spiral becomes wider or more open. There are two singular values for α : if $\alpha = 90^\circ$, the log spiral becomes a circle, and if $\alpha = 0^\circ$, the log spiral becomes a straight line.

Various authors have noted that fitting of the log-spiral shape is difficult in the downdrift section of the beach. It is a particular concern in attempting to fit to long beaches or

to beaches with one headland. However, even in these situations, a good fit could be achieved for the stretch near the headland (Moreno and Kraus 1999).

The *parabolic shape* of a headland-bay beach was proposed by Hsu et al. (1987) and is expressed mathematically in polar coordinates by Eq. 3 for the curved section of the beach and by Eq. 4 for the straight downdrift section of the beach (Moreno and Kraus 1999),

$$\frac{R}{R_0} = C_0 + C_1 \left(\frac{\beta}{\theta} \right) + C_2 \left(\frac{\beta}{\theta} \right)^2 \quad \text{for } \theta \geq \beta \quad (3)$$

$$\frac{R}{R_0} = \frac{\sin \beta}{\sin \theta} \quad \text{for } \theta \leq \beta \quad (4)$$

where R is the radius to a point P along the curve at an angle θ , R_0 is the radius to the control point at angle β to the predominant wave-front direction, β is the angle defining the parabolic shape, θ is the angle between line from the focus to a point P along the curve and predominant wave-front direction; and C_0 , C_1 , and C_2 are the coefficients determined as functions of β —the coefficients C_0 , C_1 , and C_2 are listed in Silvester and Hsu (1993) form from $\beta = 20^\circ$ to 80° at a 2° -interval (Fig. 2).

For this parabolic shape, the focus of the parabola is taken to be the diffraction point. The three coefficients needed to define the shape (see Silvester and Hsu 1991, 1993) are functions of the predominant wave angle with respect to a control line. The control line is defined similarly to the case of the log-spiral shape as the line that extends from the diffraction point to a reference point, at an angle β between the control line and the predominant wave crest orientation. Downdrift of the reference point, the coastline is assumed to be aligned parallel to the incident wave crests. This shape pertains to that of a long straight beach with shape controlled by one headland. Values of β ranged in prototype beaches from 22.5° to 72.0° , whereas the variation in model beaches was from 30° to 72° .

Sensitivity tests were performed (Moreno and Kraus 1999), where the response of the parabolic shape to a change in the value of the characteristic angle β and of R_0 was analyzed. The results proved that R_0 is a scaling parameter—the length of the control line, the alongshore extent of the shape decreases as β increases because the parabolic shape is defined only for $\theta \geq \beta$. In summary, the angle β controls the shape of the parabola, and R_0 controls its size. Because the control line intersects the beach at the point where the curved section meets the straight section of the beach, the sensitivity of the parabolic shape to errors in the estimation of the control point was examined. This was done by jointly changing R_0 and β while keeping the distance from the headland to the straight coastline constant. This observation means that the control point is not well-defined, that is uncertainty in selection of the control point and hence the corresponding joint combination the radius R_0 and β has little influence on the final result.

According to Moreno and Kraus (1999), the parabolic shape provides good fits for beaches with a single headland, because they consist of a curved section (well described by the portion of the beach protected by the headland) and a straight section (well describes the down-drift section).

González (1995) provided an improvement in the lack of definition of the location of the control point on the

formulation of the parabolic shape by developing a relationship between β and the geometry of the beach and the incident wave climate according to the following equation:

$$\beta = \frac{\pi}{2} - \tan^{-1} \left[\frac{(0.1 + 0.63y/L)^{1/2}}{y/L} \right], \quad (5)$$

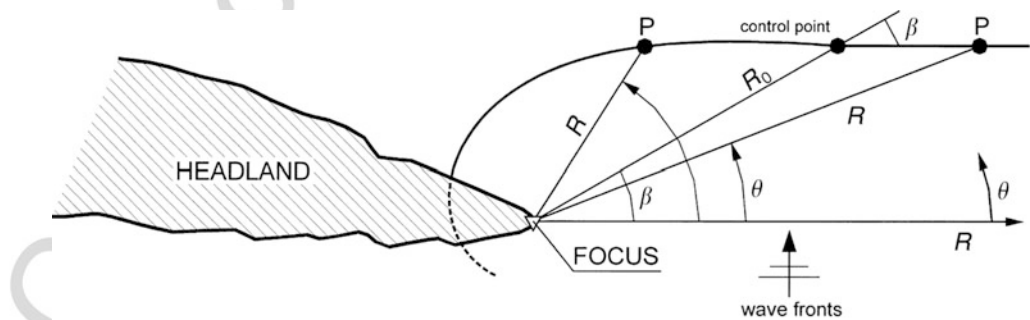
where y is the offshore distance from diffraction point to coastline, L the average wavelength in the lee area (between coastline and diffraction point).

For practical use of the parabolic shape, five unknowns need to be solved for: location of focus (two coordinates), characteristic angle β , scaling parameter R_0 , and the orientation of the entire parabolic shape in plan view.

The *hyperbolic tangent shape* was developed by Moreno (1997) and proposed for engineering design of equilibrium shapes of headland-bay beaches by Moreno and Kraus (1999) to simplify the fitting procedure and to reduce ambiguity in arriving at an equilibrium coastline shape as controlled by a single headland (Fig. 3). As mentioned above, it can be difficult to specify the location of the pole or focus, and the characteristic angle (angle between predominant wave crests and the control line) for developing a log-spiral shape or a parabolic shape. In addition, the log-spiral shape does not

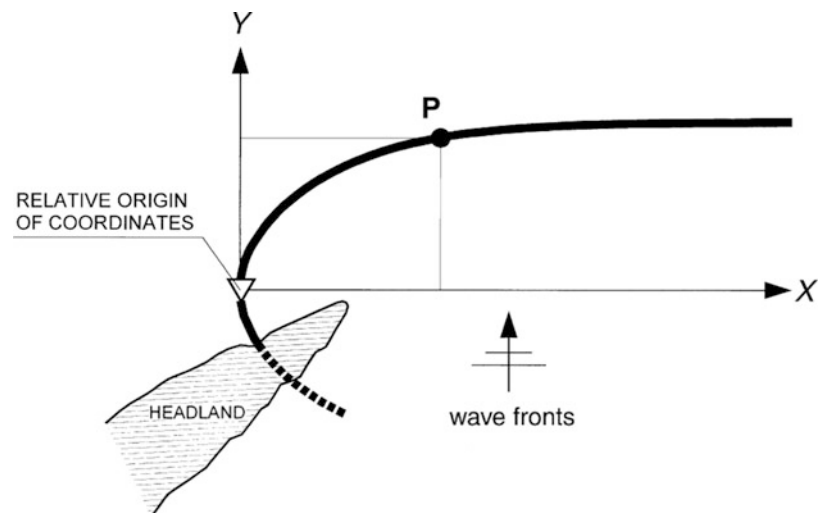
Headland-Bay Beach,

Fig. 2 Definition sketch of the parabolic planshape



Headland-Bay Beach,

Fig. 3 Definition sketch of the hyperbolic tangent planshape



describe an exposed (straight) beach located far downdrift from the headland, so that another shape must be applied.

The hyperbolic tangent functional shape is defined in a relative Cartesian coordinate system as:

$$y = \pm a \tanh^m(bx) \quad (6)$$

where y is the distance across shore, x is the distance alongshore; and a (units of length), b (units of 1/length), and m (dimensionless) are empirically determined coefficients.

This shape has three useful engineering properties. First, the curve is symmetric with respect to the x -axis. Second, the values $y = \pm a$ define two asymptotes; in particular of interest here is the value $y = a$ giving the position of the downdrift coastline beyond the influence of the headland. Third, the slope dy/dx at $x = 0$ is determined by the parameter m , and the slope is infinite if $m < 1$. This restriction on slope indicates m to be in the range $m < 1$.

According to these three properties, the relative coordinate system should be established such that the x -axis is parallel to the general trend of the coastline with the y -axis pointing onshore. Also, the relative origin of coordinates should be placed at a point where the local tangent to the beach is perpendicular to the general trend of the coastline. These intuitive properties make fitting of the hyperbolic-tangent shape relatively straightforward as compared to the log-spiral and parabolic shapes, making it convenient in design applications.

Sensitivity testing of the hyperbolic tangent shape was performed (Moreno and Kraus 1999) to characterize its functional behavior and assign physical significance to its three empirical coefficients: a controls the magnitude of the asymptote (distance between the relative origin of coordinates and the location of the straight coastline); b is a scaling factor controlling the approach to the asymptotic limit; and m controls the curvature of the shape, which can vary between a square and an S-curve. Larger values of m ($m \geq 1$) produce a more rectangular and somewhat unrealistic shape, whereas smaller values produce more rounded, natural shapes.

To fit the hyperbolic tangent shape to a given coastline, we must solve for six unknowns: the location of the relative origin of coordinates (two coordinates), the coefficients a , b , and m , and the rotation of the relative coordinate system with respect to the absolute coordinate system. Because of the clear physical meaning of the parameters, fitting of this shape can be readily done through trial and error.

Moreno and Kraus (1999) found the hyperbolic-tangent shape to be a relatively stable and easy to fit, especially for one-headland bay beaches. According to their work, the following simple relationships were obtained for reconnaissance-level guidance:

$$ab \cong 1.2 \quad (7)$$

$$m \cong 0.5 \quad (8)$$

The physical meaning of Eq. 6 is interpreted that the asymptotic location of the downdrift shoreline increases with the distance between the coastline and the diffracting headland. Equations 6 and 7 are equivalent to selecting one family of such hyperbolic tangent functions for describing headland-bay beaches, and these values are convenient for reconnaissance studies prior to detailed analysis. Equation 6 could be more precisely written as:

$$a^{0.9124} b = 0.6060 \quad (9)$$

Headland-Control Concept of Shore Protection

The headland-control concept of shore protection was first proposed by Silvester (1976) and further discussed by Silvester and Ho (1972), and was described as a combination of groins and offshore breakwaters at alongshore and seaward spacings such to create long lengths of equilibrium-bay beaches (Silvester and Hsu 1993). The structural dimensions in proportion to beach length are much smaller, and headlands are spaced much farther apart than offshore breakwaters. Therefore, it is intended to be a "regional" means of shore protection. Because headlands form pocket beaches, they might best be applied in a sediment-deficient area or for stabilizing an entire littoral reach of coast.

Headland beaches compartmentalize the coastline and reorient it in the local compartments to be parallel to the wavecrests of the predominant wave direction. If a coast has a substantial change in wave direction annually, the headland-bay beach might not be as stable as beaches behind traditional detached breakwaters, or shorter headland-bay beach compartments would be required.

A headland-bay beach design requires that a tombolo forms or be created (as by beach fill) behind the anchoring headland. If this connection is lost, the pocket beach is destroyed, and sediment can move alongshore, between adjacent compartments. Headland-bay beaches function and have their main attribute in creating independent pocket beaches, for which there is little or no communication of sand alongshore. Therefore, a headland beach presents a total barrier to littoral drift and can only be considered as a shore-protection alternative if such a barrier would not pose a problem to adjacent beaches.

The assumption that there is a single predominant wave direction which controls the final coastline shape is questionable where long distances are involved. If the design goal is to stabilize a regional extent by multiple pocket beaches, the headland-control concept might be appropriate.

Cross-References

- Bay Beaches
- Dynamic Equilibrium of Beaches
- Engineering Applications of Coastal Geomorphology
- Headland-Bay Beach
- Longshore Sediment Transport
- Shore Protection Structures
- Wave Refraction Diagrams

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Health Benefits

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Introduction

Alternative, or parallel, medicine has been steadily gaining followers and its merits have been discussed in conferences held at traditional colleges of medicine. Osteopathy, chiropractic, naturopathy, naprapathy, acupuncture, Chinese traditional medicine, aromatherapy, thermalism, thalassotherapy, and others have entered the common vocabulary for some time, some celebrating their 100th anniversary on the scene. The latter two make use of ocean algae in packs, powders, and other forms.

Thalassotherapy and thermalism are medical approaches known for thousands of years and have gained access to hospitals, private practitioners' offices, and the hallowed halls of some universities' faculties of medicine. Sea products are also used in balneotherapy and aromatherapy. The first is a treatment technique based on bathing and used in both thalassotherapy and thermalism in which plant extracts, oils, marine products (mainly salts and algae) are added to the water. Aromatherapy is more controversial; it is based on the use of organic substances (essences) with massages or in bath and steam treatments; centers are not numerous but it is seemingly *à la mode* in Great Britain where it is offered in 24 sites; a volume handling the topic in America has been published in 1999. Thonon-les-Bains is the only French center with a comprehensive program. (Valnet 1995).

Thalassotherapeutics [θαλασσα, sea; θεραπεια, care, treatment] and thermalism [θερμος, warm, heat] got a boost in the nineteenth century when tourism lifted out of its infancy. The enthusiasm displayed by prestigious visitors to

“health” centers—particularly of thermalism at the time—made it possible for “cure” stations to associate cultural, leisure, and even gambling activities with a treatment program. Napoleon III and his empress Eugénie contributed much, albeit mostly unintentionally, to the new fashion. A journey to a “cure” station took on a double aspect: a visit to a place to improve or restore one’s health, the avowed aim of a trip, and of tourism and discovery, the hidden aim of the voyage, because a voyage it often was: roughly 7 h by railway from Paris to Plombières or to Deauville. Today the *curist* and *tourist* are commonly both one and the same person.

Tourism has become a full-fledged activity. Both the “treatment” and the desire to know a region and its cultural traits motivate the visitor. And for a modern person in search of his “better-being” simultaneously with his “well-being” marketing specialists play it from both sides: tourism uses thalasso/thermalism as a travel theme, while the cure-station uses tourism to diversify its offer. The aim is to “twin,” to link, the medical quality of “cures” to the tourists’ and curists’ quality of life. An effort that has also to be directed toward children who may find relief, cure, or improvement of ailments related to breathing, skin, and developmental troubles, such as enuresis, growth, and fracture healing. *Mens sana in corpore sano*: the body will heal the better if the mind is kept away from the problem: tourism may thus in some cases become a component of the curative process.

It has been often said, lately, that the tourist is increasingly in search of a “return to the sources”; this is not a legend, but fact, and water, pure clean water that is, is commonly equated with health and wellbeing. A concern that has not escaped the attention of the European Union’s Commission (Anonymous 1996; Charlier 1999). Hence an increased interest in “water centers” whether Tatabanya, Baden-Baden, Mangalia, Varna, Djerba, or Le Touquet-Paris-Plage. The blossoming of the “social tourism” as a consequence of paid leave, the explosion of mass tourism and a trend to spend holidays at considerable distance from one’s home-base are powerful factors in making “exotic” destinations popular and sites, such as the Black Sea, both easily conceivable in vacation planning, and also very accessible. In Brussels for instance separate offices offer exclusively “health and tourism” products: *Thalasso* for one, *Thermalism* for another, are thriving on the fashionable Avenue Louise yet are patronized by a clientele from every social and income level.

The Concept of Health Tourism

Efforts have been made for decades in Romania to develop tourism and seawater therapeutics along the Black Sea coast. Other coastal countries in the area—Bulgaria, Russia, Ukraine—have not been lagging either. The success, in France for

instance, of thalassotherapy centers has been heralded by economists, tourism officials, and many health practitioners.

Thalassotherapy, and also thermalism, is thus certainly not new. What can be new is the updating of the facilities and the introduction of new technological advances. Health tourism is a powerful development tool and is undergoing an in-depth change. Where balneotherapy and exposure to coastal climate have been essentially a re-adaptive and convalescence treatment, a constantly growing segment of the clientele is made up of younger- and middle aged-people who seek an effective approach to reshaping, the *remise en forme*, encompassing not only physical reshaping but also a health restorative process.

This involves demands on computers, equipment, personnel qualification, research, thus also retooling and refurbishing, and a fresh outlook on approach and problem-solving. Marine muds have an economic potential in muds export to other countries, for example, the Italians provide muds to Bourbonne-les-Bains, France and the Israelis ship their Dead Sea muds to Plombières-les-Bains, France.

Competitive, Concurrent or Concomitant?

Thermalism, thalassotherapy, and lately aromatherapy, all claim scientifically proven curative effects. An abundant literature has been published recently, for instance in France, but often strongly publicity-slanted (François 1999). Which does not mean that several treaties have not been written long before (Jacob 1570; Russell 1720). Boulangé (1995a, b 1997a, b); Boulangé et al. 1989) of the University of Nancy, and others, like Collin (1995) and Constant et al. (1995), have made important contributions to the field (Larivière 1958; Lance 1988; Hérisson 1989; Valnet 1995). Both thermalism and thalassotherapy use waters, muds, and thermal gases.

If thalassotherapy is sole in using algae (but not algal mud), seawater and its “aeration,” techniques, however, appear to be quite similar in thermalism and thalassotherapy. They associated at the *Thermalies* of the 17th *Salon de la Santé* at Versailles in March 1999, a congress twinned with MEDEC *Salon de la Médecine*. As for aromatherapy it has not really wrought its place in the panoply of alternative medicines; it uses a variety of oils, extracted from plants; care with aromatherapy is not thus far eligible for benefits from the social security system in France, though included in “putting back into shape” treatments in some centers, including well-known ones such as for instance Thonon-les-Bains. With the notorious exception of Belgium, thermal and thalassotherapy cures are reimbursable medical expenses in most countries of the European Union and in some others as well.

There are more than a hundred thermal cure centers in France alone not counting those of overseas departments (La Réunion and Guadeloupe); thalassotherapy centers are at least a good fifty.

France with 50 centers, leads in numbers, with Germany in second place. Other stations surround the Black Sea, and are also located in Denmark, The Netherlands, and Belgium (Ostend, Knokke). Belgium (Spa, Limelette), and Luxembourg (Mondorf-les-Bains) have famed thermal centers and it is Spa that gave its name to the cure centers. The practice of “taking the waters” is far less common in the United States where it rather faded away during the in-between World Wars period, though Saratoga Spa is still “on the map.”

Historical Perspective

The map of Europe is literally strewn with places where the Romans tapped thermal waters and built their famous *thermae* from the *Baths of Caracalla* to those of Trier (Trêves), Lindesina (today Bourbonne-les-Bains), or Plombières dating from before the year 100. Roman Emperor Augustus was there, and so were scores of other leading figures of history such as writers Diderot, Chateaubriand, and rulers or their relatives like Laetitia Bonaparte (Napoleon I's mother), Napoléon III, Louis XV of France, German Emperor Wilhelm II to name but a very few.

Buchet (1985), describing medicine and surgery during the first century in Gaul, focuses on the role of thermal waters, but use of seawater for therapeutic aims was known in what are contemporary Egypt, France, Italy, and Greece as far back as 3000 BC. Nor was thalassotherapy a stranger in the medical arsenal of classical times. The knowledge and practice was spread by Celts, Gauls, and especially Romans (Grenier 1960; Kretzschmer 1966; Rameau 1980; Buchet 1985; Anonymous 1991; Malissard 1994). Bath, in Britain (the Romans' Aqua Sulis) got its name from the Romans' custom. Ancient Greeks placed considerable faith in the healing power of the sea. Greek poet Euripides (480–460 bc) wrote “The sea restores man's health,” Greek philosopher Plato (428–437) “The sea washes all man's ailments,” and 20 centuries later historian Jules Michelet (1798–1874) opined “*La terre vous supplie de vivre; elle vous offre ce qu'elle a de meilleur, la mer, pour vous relever . . .*” But thalassotherapy faded away in the Western world imbued with Aristotelian logic, later nurtured by Gallileo (1568–1642) and Descartes (1596–1650), more recently by Pasteur (1822–95) and physiologist Claude Bernard (1813–78) (Larivière 1958).

Springer (1935) traces it back, in modern times, to the Margate Royal Sea-Bathing Hospital, and famed Blackpool, England, has its spot in seawater therapy history (Charlier 1975). Russell (1720) of England, Barelli of Italy, Perochaud (and closer to us Rivière) of France, Benecke of Germany are credited as founders of the contemporary seawater therapy while Boulangé and his coauthors act as contemporary

spokesmen for French thermal-therapy (1995a, b). Thalassotherapy is, however, no longer limited to utilization of the maritime climate, but involves administration of seawater orally and by injection, use of the spray of water, of the pounding effects of waves, of heated seawater baths, and such even newer approaches as combining electroacupuncture and sea-water therapy. The 1935 four hundred seashore sanatoria and preventoria have multiplied during the second half of this century.

While centers of marine cures are numerous in France, Germany, Belgium, Russia, Ukraine, Romania (Eforie, Mangalia), Bulgaria (Varna), Israel, little or no interest has been shown for decades in the United States and is at best stagnant in Great Britain. In intense use before the 1920s, seawater injections though credited with healing nervous and blood diseases in children, fell in disuse (Larivière 1958). In the late 1930s near miraculous cures of nervitis, lumbagos, cellulitis, and obesity focused anew attention on the healing effects of sea-air and seawater (Gruber 1968). Today injections and oral administration of seawater, even in minute quantities (in wise opposition with the Russell prescriptions of large quantities!) can claim serious therapeutic effects.

Rebirth

The resurgence of interest is coupled with new concepts. One hundred fifty years ago a report ventured that “therapeutics draw good results, every day, from the use of seawater and from salt springs; and although its use in baths and tubs is often not as advantageous as when taken in the surf, when the mechanical action of the fluid is added to its chemical action, one can still expect much of this (thalassic) therapeutic application . . .” (Translation 1856).

In the twentieth century, under the influence of Freudian writings and the psychosomatic philosophical movement, medical thinking split into a traditional scientific approach and “anthropological medicine” which includes acupuncture and thalassotherapy (Range 1935). The coincidence of timing between the renewed interest in marine cures and the growing disenchantment with current ways of life in an ever increasingly technological society may be underscored. The return to the sources' desire is, of course, part of that trend.

Thalassotherapy is neither limited to nor solely based upon use of seawater: part of the treatment is the change of lifestyle, the new surroundings, the freeing of the individual from modern life's stresses, embodying aerosol- and heliotherapy. The tiny salt particles contained in sea-air (aerosols) work their way into the deepest parts of pulmonary alveoles and settle on their walls with a probably not negligible physiological effect (Woodcock and Blanchard 1957). The high proportion of ultraviolet seaside sun rays influences favorably calcium metabolism. That natural oligo-element

and others such as magnesium, manganese and cobalt, which buttress the organism's natural defenses, are also absorbed through warm seawater baths. Its biochemical properties make it a successful side-effect-free substitute for comfort medications.

Heated seawater causes a dilatation of cutaneous vessels and under water jet streams has the same beneficial effects as the pounding of the waves against the body and its spraying by sea foam. The initial shock of cold water in swimming pools has been looked at as a potential negative factor, particularly for older persons. However, most centers swimming pools are now adequately heated and wave machines provide the beneficial pounding.

Physiological effects of marine climates are reflected in a slowing down of the rates of breathing and heartbeat. The amplitude of the respiratory movement and pulmonary ventilation are increased, and so are the hematites in the blood and hemoglobin ratio, while heart contraction is reinforced; the body is thus better prepared for the beneficial impact of sea-water baths due to an increase of cutaneous exchanges. Many physicians recognize such additional symptoms as neuroendocrinic and growth stimulation, and an increase of diuresis and of basic metabolism.

Showers prior and after baths, overall or localized, exert a dual thermal and mechanical action on vessels and nerve endings; alternating of short cold- and warm-water sprays may well have the same tonifying effect as the Finnish sauna. Gynecological irrigations favor seawater's hypertonic action upon mucous tissues and penetrative ability. Nasal irrigations, aerosols, gargles help with sinus problems, ear-, nose-, and throat-ailments (Arehart 1969).

The medical and pharmaceutical value of marine "products" has of course been proven. Didemnin-B, diazonide-A, dolastin-10, and discodermolide are all potential cancer-fighting compounds derived from marine organisms, dwellers of the coral reefs ecosystems. Marine organisms produce chemical compounds and over 6000 unique compounds have been isolated with hundreds providing "drug leads"-with antiviral, antibacterial, and antifungal properties. The bryozoan *Bugula neritina* produces bryostatin-1, a potential anticancer agent, *Pseudopterogorgia elisabethae*, a soft-bodied coral known popularly as the Caribbean sea whip, produces anti-inflammatory pseudopterosins.

French and German pharmaceutical firms market vials of seawater tapped at 50 km offshore at depths of up to 20 m. It is claimed that such waters when purified provide cures for gastric troubles. With a reduced salinity the water remains nevertheless rich in magnesium and other oligo-elements, and free of chlorides, closely resembles blood plasma. Bread, crackers, and pasta made using seawater are marketed in France, Germany, and The Netherlands. Seaweeds play a significant role in cosmetology (De Roeck-Holtzhaver 1991).

Algae

Algae have been alternatively, and concurrently, praised and damned along coasts (Charlier 1990; Charlier and Lonhienne 1996) including the Black Sea (Bologa 1985/1986; Petrova-Karadjova 1990; Bologa et al. 1999). The European Union under its COST-48 program has encouraged research into their use and their eradication (Morand et al. 1990; Guiry and Blunden 1991; Schramm and Nienhuis 1996). They have a role among others in food and feed, in cosmetology, methane- and fertilizer-production, and in therapy (Găstescu 1963; Pricajan and Opran 1970; Cotet 1970).

The passage of algae components such as iron and cobalt through the skin is controversial for several physiologists. On the other hand, biomedical applications of *Lyngbya majuscula* are recognized by oncologists; this reef dwelling blue-green alga produces curacin-A which functions as an anti-proliferative that inhibits cell division, the mechanism by which cancer grows and spreads. Algae powder has been added to seawater baths and algae creams are sold in pharmacies and cosmetology stores.

They are present in some marine muds, and in some treatment centers their proportion is increased. German physicians had already in 1929 collected muds in a remote corner of Wilhelmshafen harbor. Romania has advertised them widely. French "marine cures" use an alga jelly mixed with wet sand heated in a double boiler. The mixture applied to the body slowly releases its heat. The ionic displacement of marine electrolytes and algal constituents through the skin is however, not universally agreed upon. Challenged 15 years ago, the practice is continued both in centers and on shores. German centers provide *pelotherapy* using silt packs rich in vegetal and mineral substances, rather similar to the moor-silt.

A mineral spring discovered in the royal residence of Ostend (Belgium) launched a thermalism and thalassotherapy center (1856) at one time catering to as many as 80,000 "curists." The vegetal marine mud, also in use here, is principally made up of compacted peat carried at strong high tides to the beach area. Dried, it is turned into a powder and mixed with a marine clay powder. For use in local applications the peloid is mixed with seawater and heated in a double boiler.

Some of the hyper-saline lakes of the Romanian Black Sea coast, according to the season have temperatures that may reach 27 °C with an alkaline pH. The microfauna is abundant and at least 30 species of algae have been identified. The water level may fall to 14 cm of the adjoining Black Sea. The bottom muds, rich in amino acids and carcinoids, have a high rate of natural radioactivity. Some muds are sapropelic, with phytoremaines, particularly algae which are putrefied in an anaerobic environment. Lake Techirghiol, once a bay of 1170 ha, now separated by a sand bar, is the source of the mud but irrigation of surrounding land caused a salinity decrease with resulting ecosystem changes.

Black Sea centers ring its shores; originally catering especially to their own nationals, they have increasingly drawn foreign visitors. Mangalia, southernmost resort, has attracted seamen since classical times; it forms with Eforie and Neptun, artificial creations, the Romanian cure complex. Blessed with a balmy climate, the center offers a therapy based on seawater and sapropelic mud use, sulfurous mesothermal springs; mud baths and application of mud poultices, it acquired some international reputation. The black, pasty, sapropelic mud comes from Lake Techirghiol, with a mineralization of 80 g/L. Concentrated mud extracts have been shipped to distant locations. At 150 m from the sea, the beach facilities follow the Egyptian method of open-air treatment. An air rich in iodine, magnesium, bromine, and sodium chloride creates an ambience particularly favorable for aeroionization and insolation.

As in Germany, pelotherapy is also practiced with peat mud found in Lake Mangalia. Bicarbonated, hypotonic, mesothermal (26 °C), radioactive, sulfurous water sprouts from springs on or near the Mangalia and Neptun beaches. The sapropelic mud rich in carbonaceous or bituminous matter has a plasticity value of 250 g, a thermal metric capacity index of 20.99. It is enhanced by bacteriostatic, bactericidal, and antiallergic qualities due to its high vitamin (C, E, B2 and B12), nicotinic acid, hormones, and organic content.

Economics

Setting aside the savings aspects in hospital days and pharmaceuticals consumption which benefit state and private insurance systems, and the individual, and considering the tourism aspects, it appears that thalassotherapy and thermalism are large earners and big employers. Taking, for example, the sole thermalism in France, in 1998 centers hosted 548,003 curists representing 9,864,054 “visitors” days for insured parties and 527,629 days for other curists, for a year’s total of 10,391,683. Many did not come by themselves but were accompanied by noncurists, representing an additional 300,000 persons.

A low-priced cure costs an average FRF 1000 (€ 150) for a 6-day stay, with a per person FRF 2200 (€ 320) tab for food and accommodations (thus exclusive of such additional expenses as beverages, entertainment, sundry purchases). The income for the French centers, besides payment for medical services, exceeds thus by far 13,400,000 “days” $\times$ FRF 2200 = FRF 29,480,000,000 or approximately US\$5,781,000,000, about EUR 4,494,000,000 [Exchange rates used in the calculations are FRF6.3 = US\$1; Euro 1 FR = F 6.56]. It is furthermore estimated that the 100-plus thermal centers provide employment to at least 100,000 people.

Examining these numbers for a developing economy as for instance in Romania, one one-hundredth of this is 1000 jobs

and an income of US\$46,740,000 for a single low-priced center. *Thermae* income amounts to, using the same formula, FRF 104,000,000 or about US\$16,508,000 or EUR 15,857,000 [Exchange rates used in the calculations are FRF6.3 = US\$1; EURO1 = FRF6.56]. An adjustment factor is naturally needed as wages and prices are clearly higher in France than in Romania. Nevertheless, thalassotherapy (and of course thermalism) are not to be looked at as an insignificant economic player (Guillemin et al. 1994; Boulangé 1995a, b; Anonymous 1997; Collin 1997).

Revenues are generated by use of facilities, hotels, restaurants, but also by pharmaceuticals and cosmetics: the University of California has received in royalties for patented pseudopterosins over US\$1,200,000, and the cosmetics firms have collected several millions more. Algae and muds can thus be “earners.”

Treatment Centers

Thalassotherapy stations have a long history with the largest number of stations in Germany and France. Most of the 22 German stations grew in importance during the last half century. They have attracted a large clientele and contributed substantially to the growth and expansion of touristic sites. France remains the leader with the largest number of stations, some catering to the well-to-do. Of over 40 hydro-linked health resorts found in Great Britain, the birthplace of sea-bathing hospitals, only Springs Hydro at Packington, Ashby de la Zouch in Leicestershire propose thalassotherapy; algae baths or seaweed body wraps are offered in Scotland (St. Andrews, Kingdom of Fife) and in London and balneotherapy at Newport Pagell (Buckinghamshire). Thalassotherapeutic facilities sprung up principally around the Mediterranean in Spain, Monaco, Greece, and Tunisia, and in Israel. Black Sea facilities acquired a solid reputation over the last decades. The development of a therapy *de pointe*, free from extravagant claims in up-to-date facilities can nurture a sustainable and rational growth of Romanian Black Sea resorts.

One may recall the medications developed some decades ago-and still in use-by Romanian Dr. Aslan (Gerovital, Aslanvital) against antioxidant (ao) ageing. Her work remained highly controversial to the point that some considered her therapy charlatanism. Is a similar risk conceivable with thalassotherapy and its sister cure thermalism? Indeed development must be considered on a long-range scale.

The introduction of new technologies such as the combination of electro-acupuncture with thalassotherapy is due to win over a new clientele. Thalasso-electro-acupuncture brought back from China relatively recently, has rapidly gained *droit de cité* and garnered enthusiasts in French centers.

Comparisons have been made among European centers and a critical analysis of the recent relevant bibliography published (Collin 1995; Constant et al. 1995; Graber-Duvernay 1999). A study conducted by the French National Health Insurance System (*Caisse Nationale d'Assurance Maladie*) observed 3000 persons who took a thermal cure during a span of 3 years and found a health improvement dealing with various ailments, for example, rheumatism, back pain, arterial problems (Morand et al. 1990; Guiry and Blunden 1991), in two-thirds of the group and a concomitant decrease in the length and frequency of hospital stays. Furthermore, the use of medicines dropped or was cut for 72.4% of the patients while the disbursements of the Health System for thermal care represent barely 0.22–0.43% and “cures,” stays at thermal centers, represent 0%, 89–1% of the total medical “consumption.” It is not preposterous to extrapolate these observations to the domain of thalassotherapy.

In fact, a thorough scientific project has been conducted by the department of hydrological and climatological medicine of the University of Nancy medical faculty and results made public in July 1999. The conclusions are positive and doubts as to the value of thermal cures seriously challenged. The congress (*assises*) of thermalism held in Toulouse, France in May 1999 (*Proceedings* were published in June 1999) confirms the medical dimension of thermalism. Positions held since 1996 seem thus appropriate. Furthermore, statistics show a substantial sustained drop in the frequency of hospital stays and the consumption of medicines.

The venture is foreseeably valid and sustainable. Boulangé in his 1989 and 1995 papers examining the scenario, both in the framework of the European unification and of the next century, for French thermalism and there is no reason for not extrapolating these views to thalassotherapy—sees a bright future (Boulangé et al. 1989).

With France and Germany still the leaders in centers, facilities and techniques, a noticeable increase in treatment centers has taken place in Tunisia and Morocco. One center in Austria, another in The Netherlands, import marine muds so it can claim thalassotherapy care. Marine muds of Israel are used in the Plombières (France) thermal center (Anonymous 1999a, b). A unique case is that of Ein Bokek, Israel on the Dead Sea. There is virtually no town but a cluster of hotels catering to patients seeking treatment mainly for (arthritic) psoriasis. The treatment commonly combines helio- and thalassotherapy; solar rays are filtered by an extra 300 m of atmosphere than elsewhere on earth and seawater here has a tenfold higher salt content. Results of treatments have been repeatedly commented in the *Journal of the American Academy of Dermatology* (1985) and the *International Journal of Dermatology* (1995, 1997).

Cross-References

- ▶ Beach Use and Behaviors
- ▶ Black and Caspian Seas, Coastal Ecology and Geomorphology
- ▶ Europe, Coastal Ecology
- ▶ Europe, Coastal Geomorphology
- ▶ Human Impact on Coasts
- ▶ Tourism and Coastal Development
- ▶ Tourism, Criteria for Coastal Sites
- ▶ Water Quality

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History of Coastal Geomorphology

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I am only too painfully aware how increasingly difficult it is to find time for a careful study of the work of our predecessors . . .

Geike (1897)

The coast has been of primary concern to man since he first set foot upon it and will undoubtedly continue to be so long into the future. It must have been "studied" by the earliest dwellers in their quest for food and also by later inhabitants for other reasons, such as a need of harbors or for defense. Men learned early about cliffs, rocky shores, and sandy beaches; they must also have discovered the operation and importance of tides, currents, and waves upon the shore. In light of man's long-continued interest in the coastal zone, it is surprising that its scientific study lagged so far behind that of many of the earth's other environments.

One explanation may well be that the shoreline, like so many of nature's other boundaries, failed to attract the attention of scientists. Scientifically, it was long a no man's land – neither "ocean nor land." Oceanographers were reluctant to tread where their ships would not go, and geologists were equally reluctant to tackle the sea near the shore.

Early Observations and Theories

Although the actual beginnings of coastal study are shrouded in the haze of history past, some of the speculations of the early Greeks and Romans have been preserved. They were aided by having a keen practical interest in things coastal; as early as the fourth century BC, they were well acquainted with the Mediterranean and Black Sea littorals. Their knowledge of coasts was advanced greatly by such men of action as Alexander the Great. The anonymous Greek *Periplus of the Erythraean Sea*, which described the coasts of the western Indian Ocean and even, albeit hazily, the coast east of India, was used by sailors during the days of Pliny. Although based on direct observation and unhindered by religious teachings, early theories about coasts were nonetheless tempered by the superstitions, legends, and myths that were in the heritage of all Mediterranean peoples at the time.

H. Jesse Walker: deceased.

The presence of marine fossils far from the sea led Aristotle (384–322 BC) and others to conclude that the sea had previously occupied higher levels. Strabo (54 BC–AD 25) even went so far as to write that it frequently changed levels rising at times, falling at others. The role of rivers in altering the landscape also attracted the attention of these natural philosophers, and, although they had several ideas about where and how river water originates, there seems to have been little doubt in their minds about what happens when it enters the sea. Aristotle, Strabo, and others recognized deltaic and alluvial plain deposits. They knew that river-carried sediment first shallows the sea, then converts it into marshland, and changes it eventually into dry, farmable land. Erosion as a coastal phenomenon was not ignored. Strabo, for example, noted that the ebb and flow of tides prevent deltas from advancing continuously outward into the sea.

The role of man as coastal agent was also studied. Strabo cited an example of the attempt to improve the Harbor of Ephesus in the second century BC:

The mouth of the harbor was made narrow by engineers, but they, along with the king who ordered it, were deceived. . . . He thought the entrance would be deep enough for merchant vessels. . . . But the result was the opposite, for the silt, thus hemmed in, made the whole of the harbor . . . more shallow. Before this time the ebb and flow of the tides would carry away the silt and draw it to the sea outside. (Russell 1967, p. 299)

Such was the state of knowledge and theory during those Greek and Roman times at about the beginning of the Christian era. Although these notions persisted in Europe only until the fall of the Roman Empire, they were nonetheless preserved and nurtured in the Arab world during much of the time leading up to the European Renaissance.

The Renaissance

The translation of Ptolemy's *Geography* in the early fifteenth century was a major step toward the great discoveries of the fifteenth through the eighteenth centuries. Much of the actual knowledge held by the Greeks, Romans, and Arabs about coasts came from navigators, and so it was to continue for many centuries. Many of the voyages during this period (such as those of the Cabots, Verrazano, and Gómez along the northeast coast of North America) were strictly coastal; there was no attempt at colonization or at exploration of the interior. Penrose (1955, p. 147) wrote: "A fair coastal survey was the sole result—nothing more. . . ."

Such surveys added rapidly to the body of knowledge that was beginning to accumulate and influenced those who would put their minds to coastal problems. Most of the material was pure description and just what influence it actually had on the development of coastal morphology as a science is uncertain. Nonetheless, it was available for such thinkers as Leonardo da

Vinci (1452–1519), who recognized terraces for what they are, and for Steno (1631–1686), who established a depositional sequence in explanation of the strata he observed.

Bernhard Varenius (1622–1650), because he successfully merged description with theory, has been credited with laying the foundations for geography as a science (Mather and Mason 1939). These foundations are recorded in his *Geographia Generalis*, a book that was used as a text in British universities for a century by such notables as Isaac Newton. On the topic of coastal morphology, Varenius wrote, "The ocean in some places forsakes the shores, so that it becomes dry land where it was formerly sea" (Mather and Mason 1939, p. 25). He reasoned that many factors may be involved in such a change, including those of erosion and deposition, ocean currents and tides, texture and structure, river flow, and wind. Process was very real in his science.

The Influence of Scripture

Nearly all "scientists" of the seventeenth, eighteenth, and much of the nineteenth centuries were influenced to greater or lesser degree by the Bible and the Church. In some cases, this acceptance was probably not unlike that of the Greek philosophers, influenced unknowingly as they were by mythology. The biblical quotations of greatest significance, as found in the King James' version of the Bible, are:

Let the waters under the heaven be gathered together unto one place, and let the dry land appear . . . (Genesis, 1: 9). Let the waters bring forth abundantly the moving creature . . . (Genesis, 1: 20). And . . . the flood was forty days upon the earth; . . . and the waters prevailed exceedingly upon the earth; and all the high hills, that were under the whole heaven, were covered. (Genesis 7: 17, 19)

Because, according to the Bible, the separation of land from sea occurred 2 days before animal life in the sea was created, the utilization of *fossils* in explaining the history of the earth, as the Greeks had done, was heresy. However, the Bible offers the universal flood of Noah's time as a possible out, one that was used in many explanations. A major difficulty was time. Because of the strict application of scripture, time was unavailable as a basis for applying observable processes to earth history. The major challenge was how to present the facts as offered by landscape observations without running counter to theological precepts.

Theories of Landscape Development and Coastal Morphology

Within such a Church-dominated intellectual environment, it is not surprising that a number of theories were proposed to explain the earth and its surface forms. During the eighteenth and nineteenth centuries, many were proposed, accepted,

modified, merged, and abandoned. Contemporary and subsequent history has labeled many of these theories by their predominant theme and some also by the name of their principal proponent. Examples include neptunism (Wernerism), plutonism, diluvialism, and fluvialism (Huttonism). These theories and their major advocates are treated at length in volume one of *The History of the Study of Landforms*, by Chorley et al. (1964). Basically, most of the earlier theories were catastrophic in approach, reflecting a high degree of biblical influence; later ones tended toward uniformitarianism, although there was much overlap. In any event, each theory in its own way has played a role in the development of coastal geomorphology.

The neptunists, of whom Werner (1749–1817) was the principal spokesman, invoked a primeval universal ocean in accounting for both stratigraphic sequence and landform types. The earth's rocks were formed, according to the neptunists, through the successive accumulation of chemical precipitates within the ocean. Although some surface forms were caused by submarine deposition and erosion during the presence of the universal ocean, most forms resulted from erosion during a rapid recession of ocean water. From this viewpoint, every landform originally was coastal – at least, in the sense that it was formed by turbulent oceanic currents and waves.

The diluvialists, as exemplified by Buckland (1784–1856), held a theory that was similar in some ways to that of the neptunists. The catastrophe they invoked, however, was Noah's flood. In the pure form of diluvialism, the flood was more a destructive than constructive agent. Again, it was rapidly receding water that carved the landscape. Some of the many types of otherwise unexplained phenomena that were accounted for by the flood were erratics, underfit streams, and terraces both river and marine.

Most of the catastrophists recognized that coastal and riverine erosion and deposition did occur but at such a slow pace that they had no place in earthly or "Godly" schemes. Nonetheless, even before and especially during the dominance of catastrophism, uniformitarian ideas surfaced and by the mid-nineteenth century were dominant.

The uniformitarians were united on the concept that currently operating processes were capable of creating present-day landscapes. They were not united, however, as to which processes were most important. In general, there were two camps: one, represented by Hutton (1727–1797) and Playfair (1748–1819), believed that the river was the most important agent in landscape development; the other, represented by Lyell (1797–1875) and Ramsay (1814–1891), advocated marine abrasion as dominant.

Form and Process: 17th–19th Centuries

Although it is basically correct to conclude that the dominance of the catastrophic schools during their heyday retarded

the study of geomorphology, the study of *coastal forms and processes* actually progressed to some extent. Because such forms and processes were considered insignificant from the standpoint of the overall scheme of landscape development, they could be looked at without fear of countering religious dogma if one did not try to conclude too much. Coastal cliff erosion and deposition are so conspicuous, especially in parts of the British Isles, that they did not escape the consideration of layman, geologist, and engineer alike. Hutton and Playfair, although mainly fluvialists, recognized coastal processes. Hutton, for example, wrote, "... we never see a storm upon the coast, but that we are informed of the hostile attack of the sea upon our country" (1788; quoted in Chorley et al. 1964, p. 39). Playfair, in the same vein, emphasized the obviousness of coastal erosion when he wrote, "If the coast is bold and rocky, it speaks a language easy to be interpreted." He also noted that once fragments of rock are detached they "become instruments of further destruction . . ." (1802; quoted in Chorley et al. 1964, p. 60). Thus, throughout the period of and subsequent to the Renaissance, statements appeared that indicated some relatively advanced thoughts about coasts, some of which did overstep the bounds of Church dogma.

For example, John Ray (1627–1705), a keen observer, went so far as to propose that the combination of subaerial erosion and coastal cliff retreat would eventually reduce all land to a level below the sea. Guettard (1715–1786), famous for his geological maps, believed the sea to be the major agent in land erosion and that cliff coasts were the remnants of former extensive hill systems. He observed that sediments brought to the sea by rivers mixed with material eroded from adjacent cliffs and submerged rocks. However, he tempered this view by noting that the action of the waves would have little effect beneath the surface of the sea. Lavoisier (1743–1794) adopted the Guettard idea that littoral beds are composed of materials from varied sources but went a step further and noted that the coarsest materials are highest on the shore and are followed downslope by coarse sands, fine sands, and clay. The width of each band, he maintained, varies with the steepness of the slope.

One of the most perceptive of the natural historians of this period was the little recognized John Walker (1731–1803). He was the first effective teacher of geology at the University of Edinburgh where he held the chair of natural history from 1779 to 1803. His students included the geologists Playfair, Hall, and Jameson. His lecture notes – not published until 1966 – contained many advanced notions that must have guided much of the thinking of the geologists of the early part of the nineteenth century. Some of Walker's notions, of relevance to coastal geomorphology, dealt with continental drift, "... why not America from Europe and Asia and indeed every one continent from another"; coral reefs, he was apparently the first geologist to actually describe the growth of coral reefs; and subsidence, he not only described the processes of alluviation but wrote that sediments "... are found in great quantity and to considerable depth, being the

sediment of rivers. . .” (Walker 1966, pp. 178, 183) anticipating R.J. Russell’s work of 150 years later. Unlike most geologists of the time, Walker believed that sea sand was formed by the weathering (his term) of rocks rather than by chemical precipitation from the sea (Walker 1966).

Marine Planation

With Ray’s seventeenth-century and Guettard’s eighteenth-century views of marine erosion, coupled with the fact that even the most dedicated of fluvialists placed great importance on marine processes, it is not surprising that some men in the nineteenth century considered the sea as being more powerful than the river. Lyell was one of them, although he had not always been so. During his career, he gave increasing importance to marine erosion, finally considering it to be the major modifier of the landscape. His ideas were modified by Ramsay and eventually evolved into the theory of marine planation. Ramsay’s concepts included two ideas: one, that the sea is capable of planing surfaces over which it moves, regardless of rock composition; the other, that unequal hardness in cliffs will result in differential erosion and the creation of an irregular coastline. He also maintained that marine planation accompanies shifts in sea level, using as evidence the presence of plains at different elevations above sea level. The escarpments between these levels he explained as old sea cliffs.

The importance of planation theory is emphasized by Chorley et al. (1964, p. 313): “Even when the idea of universal marine erosion had been discredited, the planation part of the theory lived on in Davis’s cycle of erosion and in the writings of mid-twentieth-century geomorphologists.”

As far as the coastline is concerned, Ramsay and others emphasized increasing irregularities because of differential erosion, a view that was not difficult to accept in the British Isles. However, Dana (1813–1895), a confirmed fluvialist, disagreed. He believed that “. . . waves tend rather to fill up the bays and remove by degradation the prominent capes, thus rendering the coast more even, and at the same time, accumulating beaches that protect it from wear” (1849; quoted in Chorley et al. 1964, p. 363). These conclusions were based on observations made during Dana’s 4-year voyage in the Pacific Ocean.

During this period, thought was being given to a number of agents previously little considered. Hutton, it is true, had recognized the importance of chemical action in soil formation, and others had discussed the transport of matter being carried to the sea in solution. Nonetheless, it was von Richthofen (1833–1905), a staunch follower of Ramsay, who applied such ideas to coasts. He wrote:

The weathering and loosening of rock by sea salts, carbonic acid, the formation of ozone, and the gripping of plants and animals-to which must be added the action of frost in higher latitudes-aids the mechanical action of the striking billows. (1882; quoted in Mather and Mason 1939, pp. 515–516)

Gilbert and Lake Bonneville

It is somewhat curious that the Western explorations in the United States, especially those during the last half of the nineteenth century, should be important from the standpoint of the history of the study of coastal morphology. This importance is even more surprising when one realizes that much of the Western field work is combined with the studies being made on the Ganges (Everest 1793–1874) and the Mississippi (Humphreys 1810–1883) rivers and in the heavy rainfall areas of the tropics in helping to reestablish the notion that the river is the dominant agent in geomorphology.

Especially significant are the coastal concepts presented by Gilbert (1843–1918) in *Lake Bonneville* (1890) and *The Topographic Features of Lake Shores* (1885). He treated a variety of shore-related topics, including beaches, cliffs, terraces, barriers, lagoons, waves, currents, undertow, backwash, and sorting, among many others. His deductions, based on intensive fieldwork, were lucidly presented. The main limitation to their usefulness was that, having been derived from work on lakes, they are not always applicable to oceanic situations. A major case in point is the concept of the bottomset, foreset, and topset bed composition of delta terraces, a concept that has little value when dealing with major oceanic deltas. Nonetheless, by describing coastal landforms in terms of physical processes, Gilbert set the style for present-day research in coastal geomorphology.

Gilbert, like most of the other geologists involved in the Western explorations, was not tradition bound and thus was able to distinguish between the relative importance of subaerial and marine processes, as he did in his Lake Bonneville research.

Possibly his most important contribution in the development of geomorphology is Gilbert’s concept of grade, a concept that he used in the development of his ideas on beach equilibrium. He wrote: “. . . in order that the local process be transportation only, and involve neither erosion nor deposition, a certain equilibrium must exist between the quantity of the shore drift on the one hand and the power of the waves and currents on the other” (1885, p. 101).

Gilbert utilized lake shorelines and lake deposits as indicators of past climates and tectonic history. For example, he was able to correlate lake terrace width with rock type in a lake’s discharge channel and the lack of horizontality and parallelism of shores with orogenic movement.

Davis, Gulliver, and Johnson

The role of William Morris Davis (1850–1934) in the development of geomorphology has been analyzed many times (most thoroughly by Chorley et al. 1973). Davis influenced in some way nearly every aspect of geomorphology, including coastal geomorphology. This influence was realized in

several closely linked forms, including his own research and publications on coastal topics, the wide adoption of his concepts of the cycle of erosion, and the work of his students (especially Gulliver and Johnson) who emphasized the study of coasts.

Much of Davis's research on coastal problems was related to reestablishing support for Darwin's subsidence theory of coral reef growth. By writing some three quarters of a century after Darwin, Davis was able to incorporate in his writings data about sea-level changes during the glacial period. He believed that the only way to properly understand coral reefs was through an examination of the "... physiographic features of the coasts, either insular or continental, that are bordered by fringing reefs or fronted by barrier reefs ..." (1928; quoted in Chorley et al. 1973, p. 592) and not from the reefs themselves.

Another example of Davis's contributions in coastal geomorphology is his *The Outline of Cape Cod* (1896). It illustrates an attempt at geomorphic reconstruction: "Let the activities of the sea be resolved into two components: one acting on and off shore, the other along shore; and let the effects of the first of these components be now examined alone ..." (Davis 1896, p. 700).

Davis incorporated the ideas of Gilbert's beach equilibrium within his cyclic concepts: "Here the sea is able to do more work than it has to do. Its action is like that of a young river. ... When a graded profile is attained, the adolescent stage of shore development is reached" (Davis 1896, p. 701). This paper also presents numerous diagrams, a Davisian hallmark, illustrating the development of graded profiles (both normal and longitudinal) and of bars and spits.

The cyclic concepts presented in the Cape Cod paper preceded by 3 years the publication of his most famous and influential paper, *The Geographical Cycle* (Davis 1899). The Cape Cod paper is only one example of the fact that the cyclic concept had been in Davis's mind for many years before the turn of the century. The influence of this concept in coastal geomorphology is further evidenced in the dissertation by Gulliver (1865–1919) that was entitled simply *Shoreline Topography* (1899) and was published in the same year as *The Geographical Cycle*.

Davis was continuously working with the cycle and considered that it is "... not arbitrary or rigid, but elastic and adaptable. ..." For example, "Like the processes of surface carving, the processes of shoreline development are subject to variation with climate, from the work of the ice foot in polar regions to the work of coral reefs and mangrove swamps in the torrid zone" (Davis 1905, p. 290).

The most influential of Davis's disciples was Douglas Johnson (1878–1944). Despite Gulliver's early coastal work under Davis's direction, it remained for Johnson to publish the first inclusive book dealing with coastal morphology, a book that Zenkovich (1967) considered to be the most

complete theoretical study of coasts available. Entitled *Shore Processes and Shoreline Development* (Johnson 1919), it was aimed at presenting an analysis of the forces operating along the shore together with a systematic discussion of the cycles of shoreline development. This book had a major influence on the study of coasts for at least 40 years, an influence that must be considered to have been detrimental in some regards. Johnson's emphasis on the importance of submergence and especially emergence in shoreline development, as well as the incorporation of these aspects of shore profile development in his classification scheme, delayed more meaningful approaches to coastal understanding for several decades. Nonetheless, much of his material is still useful and should be consulted by any serious student.

Coastal Geomorphology: World War I to World War II

Although the Davisian and Johnsonian evolutionary and qualitative approaches to geomorphology dominated coastal research from World War I to World War II, there were nonetheless a number of important developments that occurred during this period of time. One of the most significant of these developments was that coastal research became a respectable research endeavor after Johnson's book was published. Earlier most coastal research was of a practical nature—harbor construction, coastline defense, coastal mining, and the like. Few university scholars studied coasts for their own sake. Geologists, for example, often resorted to coasts but only because coastal cliffs provided them with good exposures of the rocks they were studying, not because of any interest in their existence as coastal forms.

A second and concurrent development was the rapid rate at which human utilization of the coast developed. The increase resulted directly from increases in coastal populations and indirectly from man's increased mobility, desire for coastal vacations, and use of coastal resources. Fortunately, this increased utilization of the coast was accompanied by an increase in its study, although not to the extent that might have been desirable.

The sponsorship by the United States National Research Council of separate studies on tides and sea-level changes in the 1920s, the creation of the Commission of Coastal Studies in the USSR and the Coastal Engineering Research Center in the United States, and the publication of such volumes as *Recent Marine Sediments* (Trask 1939) are examples of the organized endeavors that began between the two World Wars.

Despite a number of technical advances, there were surprisingly few actual substantive developments. In a very real sense, this period of time was transitional and set the stage for the rapid rise in coastal research that began after World War II. The way in which this transition proceeded might be illustrated by considering the way in which depositional

landforms came to be recognized as significant elements of the landscape. One of the pioneers of depositional geomorphology was R.J. Russell (1895–1971). Trained in the Davisian School with its emphasis on erosion, Russell had to develop a completely new perspective in order to understand the depositional landforms he found when he moved to Louisiana. He utilized the thousands of cores taken during the drilling for oil in the Mississippi River delta, sets of detailed topographic maps (the drafting of which was prompted by a severe flood in 1927), and aerial photography (Russell 1936). This research of Russell and his co-workers over a 20-year period led to the acceptance of the importance of three-dimensional studies in geomorphology, a clarification of the different types of subsidence, a reevaluation of Johnson's concepts of emergence and submergence and of Davis's concept of old age, and to new notions about sea-level fluctuations.

World War II and the Impetus It Gave Coastal Research

By the end of the 1930s, research on coastal problems, as well as that on other scientific topics, was beginning to recover from the difficult times that accompanied a worldwide depression when World War II erupted. Many scientific and engineering activities such as those being planned by the revitalizing Beach Erosion Board (BEB) of the United States had to be redirected toward war efforts. An even more extreme response to wartime demands was the dissolution of the Commission of Coastal Studies in the USSR mentioned above. Before its demise, however, the Commission sponsored a number of coastal engineers and geomorphologists (such as Vsesolod Zenkovich) many of whom became very productive after the war.

In the case of the BEB, wartime coastal research efforts were focused on gathering information on foreign beaches. Krumbein (1944) of the United States and Williams (1947) from England, along with their associates, concentrated on those coastline characteristics that could impact beach landings including beach slope, sediment texture, wave climate, and tidal parameters. During the War, the BEB produced more than 50 reports on beaches around the world including some in the little previously studied tropical zones.

More important from the standpoint of coastal research than the descriptive details of many of the world's beaches that came from these efforts were the war-generated developments that impacted virtually all of science. The developments included (1) the assemblage of scientists into research groups including those that became known as research and development laboratories, (2) many technological advances in equipment useful to coastal research, (3) specialized educational opportunities at universities, (4) the spread of both scientists and scientific research around the world, (5) the multiplication of regional, national, and

international organizations devoted to geomorphology including coastal morphology, (6) an increasing emphasis on scientific conferences, workshops, and symposia, and (7) an increase in opportunities to publish on basic, as well as applied, topics.

Coastal Research During the Third Quarter of the Twentieth Century: A Period of Expansion

One of the major advances that blended many of these developments was the creation, in 1952, of the Commission on Coastal Sedimentation within the International Geographical Union (IGU) – a commission whose name was changed a few years later to the Commission of Coastal Geomorphology (Schou 1964). The senior members of the Commission – A. Schou (Denmark), J.A. Steers (England), V. Zenkovich (USSR), R.J. Russell (USA), and A. Guilcher (France) – were among the most productive geomorphologists during and immediately following World War II. With the diverse topical and regional backgrounds they and numerous corresponding members represented, they recommended a very diverse set of topics that needed research (Schou 1964).

A couple of years earlier (1950), coastal engineers began holding biennial conferences from each of which stemmed proceedings volumes on coastal topics. By 1996, more than 2,300 papers had been produced, most of which have relevance to coastal geomorphology.

Whereas the 1950s showed much progress, some coastal geomorphologists still considered the subject immature. For example, Williams (1960) observed that:

... theories of coastal behaviour have been based on the most simple visual observations, and there is no doubt that some of these observations have been misleading ... [however, he added further that] today new techniques are being developed which should lead to a more certain knowledge of the subject. (p. xi)

Similarly, as late as 1968, Russell still considered "... the subsistence of coastal morphology as one in relative infancy ..." but, like Williams, one with great promise. During the 1950s and 1960s, a number of coastal scientists began to base their research on the concept of process and response. They in general followed the procedure advocated by Strahler (1952) for fluvial morphologists. Strahler stated that:

... dynamic-quantitative studies require, first, a thorough morphological analysis in order that the form elements of landscapes may be separated, quantitatively described, and compared from region to region. (p. 1118)

Such a procedure proved to be especially valuable for morphologic research in the coastal zone because of the great number of forms and processes present there and because of the complex nature of the interrelationships that exist between them. The process-response models that resulted from such studies proved of value in both the production of specific

types of coastal behavior and the provision of a clearer understanding of the integrated nature of the coastal system.

During the period of the 1950s and 1960s, such varied techniques as radioactive tracers, aqualung diving, satellite imagery, electron microscopy, and high-speed computers began being used to provide data and analyses about coastal forms and processes at scales both larger and smaller than had been possible during prewar years (Walker 1977).

Although the scientific study of coasts had traditionally been in the hands of Western Europeans and Americans, during the years following World War II, it truly became international. Evidence of this development is indicated by the frequency with which international symposia were being held, by the increasing numbers of research papers being produced in non-Western countries (Walker 1976), and by the increasing frequency of research along arctic, desert, and tropical coasts.

In the third quarter of the century, the number of publications in coastal geomorphology increased dramatically. Some of the books, like that of Douglas Johnson 40 years earlier, were broadly based. Included were Guilcher's *Morphologie Littorale et Sous-Marine* (1954), Zenkovich's *Processes of Coastal Development* (1967), Bird's *Coasts* (1969), King's *Beaches and Coasts* (1972), and Davies' *Geographical Variation in Coastal Development* (1973). Others such as Ippen's *Estuary and Coastline Hydrodynamics* (1966), Shepard and Wanless' *Our Changing Coastlines* (1971), and Komar's *Beach Processes and Sedimentation* (1976) deal with special aspects of coastal science.

Also, during this period of time, collections of papers began to become common. Many of them were special issues of standard periodicals, such as *Dynamics and Morphology of Sea Coasts* edited by Longinov (1969) as volume 48 of the *Transactions of the Institute of Oceanology*; Tedrow and Deelman's *Soil Science* (1975), which is devoted to soil formation in sediments under water; and Fairbridge's *Contributions to Coastal Geomorphology* (1975) and Kaiser's *Küstengeomorphologie* (1968), as special issues of *Zeitschrift für Geomorphologie*. Yet another category of volumes that resulted from symposia includes *Estuaries*, edited by Lauff (1967); *Waves on Beaches and Resulting Sediment Transport*, edited by Meyer (1972); *Coastal Geomorphology*, edited by Coates (1973); *Nearshore Sediment Dynamics and Sedimentation*, edited by Hails and Carr (1975); and *Research Techniques in Coastal Environments*, edited by Walker (1977). Still, a third type of compilation was developed toward the end of this period. Especially valuable from the standpoint of the development of coastal concepts, its volumes brought together the key papers representing the development of particular topics. A prime example is the *Benchmark Papers in Geology*, of which *Spits and Bars* (1972) and *Barrier Islands* (1973), both edited by Schwartz, and *Beach Processes and Coastal Hydrodynamics*, edited by Fisher and Dolan (1977), are especially appropriate.

Contemporaneous with the publication of volumes such as those listed above was the development of a number of concepts that modified the focus of many coastal researchers. In 1962, Per Bruun, in response to the emerging concern over sea-level rise, proposed what is now known as the *Bruun Rule* (Schwartz 1967). In essence, it states that a beach will maintain equilibrium through concurrent erosion and deposition as sea level rises. Criticized and modified subsequent to its proposal, the *Bruun Rule* nevertheless continues to be the focus of numerous studies on into the twenty-first century.

Although coastal sediment transport had been the subject of discussion for decades, the coastal circulation cell concept in relation to the compartmentalization of erosion, transport, and deposition of sediment along coasts was proposed in the 1960s. Bowen and Inman (1966) applied the concept in California, while Stapor (1971) used it in connection with a Florida study.

Probably the most innovative development to appear during the third quarter of the twentieth century was the scheme proposed by Inman and Nordstrom (1971) in which plate tectonics was used as the basis for classifying coasts. With this scheme, coastal morphologists were forced into giving "... more thought to long-term but continuing processes in their attempts at explaining present form and location" (Walker 1975, p. 4).

The Last Quarter of the Twentieth Century: A Period of Sophistication and Diversification

Whereas the third quarter of the twentieth century was one of rapid expansion in coastal research, the last quarter might be considered as one in which research became more sophisticated and diversified. Improved equipment, expanded multidisciplinary cooperation, modified methodologies, and enhanced funding contributed to a number of new research avenues in coastal geomorphology. Possibly one of the most important realizations, following from the cell concept, was the recognition of the interrelationships that exist between different parts of coastal systems.

Even though coastal research became more diversified during the 1980s and 1990s, the research dedicated to beach morphology and processes continued to receive the most attention. Much of this attention is devoted to small-scale hydrodynamics and sediment transport as it occurs in the surf/swash zone. During the 1980s, several large-scale field experiments were conducted on and in the surf zone. Included were the *Nearshore Sediment Transport Study* (Seymour 1987), the *Duck Experiments* (Mason et al. 1987), and the *Canadian Coastal Sediment Study* (Willis 1987), among others (Horn 1997). Not surprisingly, considering the complicated nature of such a dynamic zone, different views as to the nature of swash dynamics developed. One group, as

examined by Guza et al. (1984), holds that swash is dominated by low-frequency infragravity motions and can be attributed to standing long waves; whereas another group, represented by Hibberd and Peregrine (1979), holds that swash motion is mainly driven by incident waves that collapse at the shoreline and propagate up the beach face.

For the geomorphologist, the value of these hydrodynamic studies is in their help in predicting sediment transport and beach morphology. New instruments, such as the optical backscatter sensor (OBS) and the acoustic backscatter sensor (ABS), are enabling the measurement of suspended sediment concentrations in the surf zone and are being used in conjunction with acoustic Doppler velocimeters (Downing et al. 1981).

As noted above, much of the research along the shoreline has been concerned with the contact zone of the beach and surf. Nevertheless, increasing attention, especially during the past two decades, has been given to mudflats and coastal dunes and especially to their marsh/mudflat and beach/dune systems (Viles and Spencer 1995). Although mudflats, like the marshes and mangroves behind them, traditionally were considered wasteland and thus prime areas for reclamation, their importance ecologically, once recognized, has served to bring them into the research spotlight.

In China, where mudflats have been reclaimed for millennia, basic research into their characteristics was initiated primarily by the Coastal and Estuarine Research Laboratory (later one of China's key laboratories) in Shanghai under the leadership of J. Chen and the School of Geoscience, Nanjing University. Much of the earlier mudflat research was on mudflat distribution, but by the late 1980s, it included examination of how tidal currents influence grain size distribution (Wang 1989) and mudflat sediment exchange (Tang-Yinde 1989). The stability/erodibility of mudflats during tidal cycles and in response to mudflat biotic activity has also been the subject of recent research (Viles and Spencer 1995). During the 1990s, several investigators began dealing with the effects of sea-level rise on muddy coasts (El Ray et al. 1995; Han et al. 1995).

The seminal research on sand dunes by Bagnold (1941) served as the rationale for dune research for more than 30 years following World War II. It was not until the late 1970s that some researchers began to emphasize the distinctions between interior and coastal dunes. An increased appreciation for the facts that the presence and stability of dunes are important in coastal protection, that human activity can impact heavily on dunes, and that coastal dunes play an important role in the beach/dune system led to an increase in the quantity and variety of research on coastal dunes during the last quarter of the twentieth century.

Much of the new research on coastal dunes has been devoted to the determination of vertical velocity profiles across beaches and within dune fields and to the measurement of sediment transport by winds (Carter et al. 1990). Included

is the basic research by Hotta (1988) and by Hsu (1977) who examined the boundary layer conditions in both sand dunes and coastal ice ridges along an arctic coastline. Hydrodynamic research, which has become so important in many aspects of coastal research, has been applied to coastal dunes by Hesp (1988) and his colleagues. For example, Hesp used morphodynamics as a basis for classifying foredunes by linking vegetation and morphology to nearshore processes. Other factors that have received attention recently are the importance of beach wetness to sediment transport (Jackson and Nordstrom 1998) and large-scale budgeting in the beach/dune system (Illenberger and Rust 1988).

Dune research, like that on beaches, has become very international in scope. Some of the most important morphologic researches on coastal dunes has been conducted in Australia, South Africa, the Netherlands, Japan, the United Kingdom, and the United States.

Among the other coastal topics that have recently attracted increased attention are shore platforms, sea-level rise and coastal erosion, extraterrestrial coastal geomorphology, and humans as geomorphic agents.

The focus in shore platform research has been on morphology and processes (Stephenson 2000). For example, Trenhaile and Byrne (1986) concentrated on tides and sea-level change, Sunamura (1992) emphasized wave dynamics, and Viles and Naylor (2002) investigated the role of biogeomorphic processes on platform development.

Along with the proliferation of research on global warming and sea-level change since the 1970s has come a surge in the investigations devoted to the impact of sea-level change on the coastline. Because all types of coasts are affected when sea level changes, it is not surprising that research of such impacts has been done on beaches, reefs, estuaries, deltas, and even coastal cliffs (Viles 1989). Morphological and ecological changes in coastal marshes have been intensively examined by Reed (1995), mangrove response by Woodroffe (1995), foredune erosion by Carter and Stone (1989), and, as mentioned above, mudflats by Han et al. (1995). This kind of research is destined to intensify even more so long as sea level continues to rise as it has been doing the past few decades.

Extraterrestrial geomorphology is one of geomorphology's newest subdisciplines (Baker 1993). It, like most other themes in geomorphology, has a long history (see, e.g., Gilbert 1893). Not surprisingly, the field of extraterrestrial geomorphology also has a coastal component (Baker 2001). Parker et al. (1993), after examining high-resolution images of Mars, concluded that that planet not only had liquid water in the past but also possessed lakes and even oceans. Recent studies have identified relic lake sediments, deltaic deposits, and eroded "massifs" (resembling the wave-eroded headlands in Lake Bonneville) (Parker and Currey 2001). It is believed that other extraterrestrial bodies such as Venus and Titan, the

moon of Saturn, may also support such features and thus provide research opportunities for additional coastal geomorphologists well into the future.

The importance of humans as agents of geomorphologic change has long been recognized and documented (Thomas 1956; Turner 1990). The coastal zone, as one of the world's unique landscapes, has been impacted intensively along much of its extent. Indeed, along the coastline of some of the world's most densely populated coastal zones, the bulk of the shore is now artificial. This provides an opportunity and challenge to coastal geomorphologists because, as Viles (1990) noted:

Studies on coastal landforms, processes and change are becoming more and more relevant as human intervention (at scales ranging from sand mining on a single beach to global warming) increases. (p. 238)

The rapid increase in human intervention in the coastal zone is intimated by the fact that in the year 2000, the number of coastal dwellers was equal to the global population of 1950 (Haslett 2000). It is not surprising then that many coastal geomorphologists as well as coastal engineers are heavily involved with coastal research.

Much of this research has been focused on coastline protection (Walker 1988; Pilarczyk 1990), beach nourishment (Finkl and Walker 2002), beach stabilization (Silvester and Hsu 1991), beach management (Bird 1996), and wetland restoration (Mitch and Gosselink 2000). Sherman and Bauer (1993), referring to this trend, predict that:

... human altered coastal systems will be a major focus of research for coastal geomorphologists over the next twenty years ... [and, equally as relevant] that coastal scientists will have less choice and less input to what their subject of study will be ... (p. 240)

Early Twenty-First Century: The Age of Instrumentation, Climate Change, and Natural Disasters

Many of the late twentieth century's major fields of coastal research, such as sediment transport, beach and dune morphology, and rocky shelf development, have continued into the new century; while coastal research on the impacts of climate change and associated global warming has greatly expanded. The twenty-first century has seen a number of major coastal natural disasters that have been the focus of much research. Major seismic events have triggered tsunami killing hundreds of thousands of individuals and causing billions of dollars in infrastructure damage. Coastal scientists have relied heavily on ever-evolving technology to study these events and others. This section will focus on the dominant themes of coastal research in the early twenty-first century: technology, climate change, natural disasters, morphology and processes, and extraterrestrial geomorphology.

Advances in technology has infiltrated all areas of coastal research in the new century with new and more advanced technologies and associated data being utilized. High-resolution radar imagery, such as Synthetic Aperture Radar (SAR) and Light Detection and Ranging (LIDAR), with their ability to map features in 3D, has emerged as a key technology. SAR imagery has a spatial resolution of 1 m and can be acquired during both daytime and night and in all types of weather. SAR data are used to study breaking waves and nearshore currents (Hwang et al. 2013), storm surge flood inundation (Ramsey et al. 2011, 2013), dune fields (Ballarini et al. 2003), and beach profiles (Marghany et al. 2010).

LIDAR data typically has a spatial resolution of 1–3 m, with high-density LIDAR having submeter resolution, and vertical accuracies of 12–15 cm (Klemas 2011). LIDAR data are acquired using an airborne platform. Most LIDAR used for coastal research must be acquired during either morning or afternoon during periods of little to light wind. LIDAR has been used to study coastal inundation from sea-level rise, storm surge from tropical weather systems (Stockdon et al. 2009; Stoker et al. 2009; Gesch 2009), dune recovery after extreme storm events (Houser and Hamilton 2009), and dune and beach morphology (Lee et al. 2004).

Near-infrared (NIR) LIDAR uses wavelengths ranging from 1,000 to 15,000 nm for surface topography 3D imaging. It is used to examine temporal and volumetric changes in dunes (Woolard and Colby 2002), sea cliff retreat, and beach morphology (Young and Ashford 2006).

LIDAR is also used to provide high-resolution imagery of shallow nearshore bathymetry. Bathymetric LIDAR uses the green band (532 nm) of the electromagnetic spectrum to penetrate clear shallow water. NIR-LIDAR is being used to map shallow bathymetry (Allouis et al. 2010).

Dynamic coastal processes, such as sediment transport, wave dynamics, and beach processes, are studied using video imaging with high-speed cameras capable of recording several frames per second (Lee et al. 2004). The cameras usually record only visible light; thus, can only record in clear daylight conditions.

Unmanned aerial vehicles (UAVs), better known as drones, have become an important tool (Klemas 2015; Turner et al. 2016). Drones provide a low-altitude stable platform for aerial photographs, LIDAR, and video image acquisition. Drones have also been used to study dune and beach morphology and the beach-surf zone processes (Casella et al. 2014, 2016).

The early 2000s experienced an average increase in sea surface temperature of $\sim 0.2^{\circ}\text{C}$ (Huang et al. 2015); this coupled with the $\sim 0.4^{\circ}\text{C}$ increase in sea temperatures since the late 1800s has had a profound impact on the world's coastlines. In addition to rising sea levels, climate change is causing sea temperatures to increase. The dual threat of rising and warming seas has led to the inundation of low-lying

coastal areas, an increase in the inundation of storm waves and surges, and an increase in beach and cliff retreat and coastal landslides (Leatherman et al. 2000; Zhang et al. 2004). Researchers have rushed to study and predict these impacts and to understand altered coastal processes. This is especially true in the Arctic where rapidly thawing permafrost is accelerating coastal bank erosion.

Worldwide, the response of river deltas to sea-level rise (SLR) has continued to be important. The drowning of the Mississippi and Nile River deltas and efforts to offset the erosion/subsidence are ongoing. Researchers focus on sediment availability and wetland response to the rising seas.

Modeling the predicative effect of SLR on coastal systems has increased in the twenty-first century. Researchers have modeled the potential impacts of SLR on sandy coasts and their sediment transport (Davidson-Arnott 2005; Vitousek et al. 2017). Letterman et al. (2000) determined that shoreline retreat of sandy beaches is 150 times the rate of sea-level rise. Rocky cliff coasts, especially soft rock cliffs, will experience an increase rate of retreat due to intensified wave action and an increase in precipitation (Trenhaile 2014).

The early twenty-first century has witnessed significant coastal disasters, such as Hurricanes Katrina and Sandy in the United States and the Sumatran and Fukushima earthquakes and tsunamis. These events have exposed the vulnerability of the millions of people who live in the coastal zone. Catastrophic tropical systems have the ability to erode barrier islands and dune fields, destroy coastal wetlands, and flood inland areas.

In December of 2004, a magnitude 9.0 earthquake occurred just off the coast of the island of Sumatra, Indonesia triggering a tsunami that affected most of the coastal areas of the Indian Ocean. Another magnitude 9.0 earthquake occurred several years later, in March 2011, just off the coast of Fukushima, Japan. Similar to the Sumatran event, a large tsunami was generated which came ashore along Japan's heavily populated eastern coast. Both of these events had profound coastal impacts. Sandy beaches washed away, some coastal areas experienced uplift, new tidal inlets formed, and extensive inland deposition occurred. Initial research surveyed and documented the damage (Tanaka et al. 2012); this shifted to long-term studies of beach recovery and changing sediment patterns.

Dunes continue to be a focus of much research. Dune processes such as foredune evolution, modeling, and blow-outs dynamics are key areas of study (Hesp 2002). Modeling the movement of sediment transport and the interaction of sand and vegetation has emerged as an important area of research. Different types of models, including using fractals and chaos theory (Baas 2002), are being used to study this interaction.

A number of studies have been conducted on the response of dunes to extreme weather events, such as hurricanes

(Houser et al. 2008; van Rijn 2009). Dune erosion, sand transport, retreat, and/or recovery from storm surge continues to be important especially in this age of climate change.

Bulging coastal population is placing increasing stress on coastal dunes. Dunes are increasingly being destroyed and replaced by buildings and parking lots. Indirect damage to vegetation is occurring by visitors who trample the vegetation (Martinez et al. 2008).

Proving the existence of Martian landforms that are either relic oceans (Parker et al. 2010) or coastal-related features, such as submarine landslides (DeBlasio 2011) and shorelines (Ghatan and Zimbelman 2006), is an ever-growing subfield. Researchers are using high-resolution imagery to understand the paleoclimate of Mars based on their understanding of earth's coastal process. Rodriguez et al. (2016) identified paleo-shorelines and possible tsunami run-up lobes of boulders and backwash channels.

Organizations, Conferences, Proceedings, Journals, and Books

The fourth quarter of the twentieth century saw that the number of organizations, conferences, proceedings, journals, and books dealing with coastal geomorphology continues to increase. In addition, the coastal component of a number of geomorphology organizations also increased in prominence. For example, at the First International Conference on Geomorphology (1985), more than one fifth of the papers presented dealt with fluvial and coastal topics. Coastal geomorphologic papers have continued as an important ingredient of subsequent conferences. Recently established journals (i.e., those founded since 1975) that publish coastal papers of interest to geomorphologists include *Applied Ocean Research* (1979), *Coastal Society Bulletin* (1977), *Earth Surfaces Processes and Landforms* (1976), *Geomorphology* (1987), and the *Journal of Coastal Research* (1985). In addition, many long-established journals have recently published special numbers devoted to coastal topics. They include *The Geographical Review* with a number called *Coastal Geomorphology* (1988), *Catena* with a number on *Morphology and Sedimentation in Fluvial-Coastal Environment* (1997), *Marine Geology* with numbers on *Beach Ridges* (1995) and *Large Scale Coastal Behavior* (1995), and *Physical Geography* with a number on *Coastal Dunes* (1994).

One of the most extensive series of special numbers devoted to coastal topics is that published as Special Issues by The Coastal Education and Research Foundation (CERF). Since 1986, more than 30 such special issues have been published under such titles as *The Effects of Seawalls on the Beach* (1988), *Impacts of Hurricane Hugo: September 10–22, 1989* (1991), *Coastal Hazards* (1994), *Sediment*

Transport and Buoyancy in Estuaries (1997), and *Tidal Dynamics* (2001).

In addition to such special numbers are the numerous review articles that have recently appeared detailing the advances in coastal geomorphology. *Progress in Physical Geography*, for example, has had a number of status reports including *Coastal Landforms* by McCann (1982), *Coastal Depositional Landforms; a Morphodynamic Approach* by Wright and Thom (1977), *Coastal Geomorphology* by Viles (1990), *Sea-level Rise as a Global Geomorphic Issue* by Stoddart and Reed (1990), *Beach Research in the 1990s* by Horn (1997), and *Mid-Holocene Sea-Level Change and Coastal Evolution* by Long (2001).

Possibly the most revealing characteristic in the development of coastal research subsequent to World War II is illustrated by the articles appearing in what must now be considered the bellwether journal of coastal science, the *Journal of Coastal Research* (JCR). Since its foundation in 1985 as an “International Forum for the Littoral Sciences,” it has published some 1,400 articles in more than 18,000 pages. Although published in the United States, JCR reflects the international character of today’s coastal research in that more than half of the articles stem from countries other than the United States. Of the total, the United Kingdom is represented by more than 80 articles: New Zealand, Australia, and South Pacific Islands by 116, Canada by 57, Latin America by 73, Scandinavia and the Netherlands by 63, the Orient by 51, and Russia by 24 (Finkl 2002, personal communication).

Equally as revealing of the trend in research in coastal geomorphology is the specialized nature of the books that have appeared since 1975. Although prior to that date, as noted above, conference proceedings were often quite topical, books tended to remain inclusive. By 1980, however, coastal books began to reflect the maturation of coastal geomorphology in two significant ways. First, details about coastal environments had developed to a point where specialized volumes became justified, and second, theories of coastal science had been enhanced sufficiently to lead to new approaches in considering the field especially in those portions of coastal geomorphology relevant to human activities.

The first category of books is well exemplified by a series, edited by E.C.F. Bird for John Wiley & Sons, that includes *Coral Reef Geomorphology* (1988) by André Guilcher; *Coastal Dunes: Form and Process* (1991), edited by K. Nordstrom, N. Psuty, and B. Carter; and *Geomorphology of Rocky Coasts* (1992) by T. Sunamura. The second-type approach is represented by three books recently published by UK authors, namely, *Coastal Problems: Geomorphology, Ecology and Society at the Coast* by Viles and Spencer (1995), *Coastal Systems* by Haslett (2000), and *Coastal Defences* by French (2001).

The number of coastal-related journal articles has increased by approximately 36% since the end of the century.

The largest number appeared in the *Journal of Geomorphology*, *Journal of Coastal Research*, and *Marine Geology*. The majority of the papers have focused on climate change, modeling/remote sensing, coastal hazards, and coastal processes/evolution.

Since 2000, several journals have published special issues dedicated to coastal geomorphology. The Geographical Society of London published an entire volume dedicated to rock cliffs entitled *Rock Coast Geomorphology: A Global Synthesis* (Kennedy et al. 2014). The *Journal of Coastal Research* has had two special issues on the application of LIDAR in the coastal zone and one dedicated to coastal erosion.

While the actual number of coastal-related academic and textbooks has decreased since 2000, notable publications include *Coastal Geomorphology: An Introduction* (Bird 2008) and *Introduction to Coastal Processes and Geomorphology* (Massolink et al. 2014). Topical books have covered a range of coastal subjects ranging from dune restoration to coastal morphology modeling. Notable titles include *Restoration of Coastal Dunes* (Martinez et al. 2013); *Beaches and Dunes of Developed Coasts* (Nordstrom 2000); *Coastal and Marine Hazards, Risks, and Disasters* (Shroder et al. 2014); and *Tsunamis and Earthquakes in Coastal Environments: Significance and Restoration* (Santiago-Fandiño et al. 2016).

Conclusion

Although, as noted in the introduction of this History of Coastal Geomorphology, the coast was late in attracting the attention of scientists, it eventually became one of the major subdisciplines in the field of geomorphology. The rapidity with which it has developed during the past few decades might have pleased R. J. Russell who, even as late as 1968, considered the field in its infancy. Since World War II, a number of conditions evolved to place coastal geomorphology on a sound footing. Advancements in monitoring systems, development of new theories, increased funding opportunities, broadening the scales of investigation, inputs from other disciplines, and debates of controversial conclusions have all played a part in the maturation of coastal geomorphology.

As has been true throughout the history of science, each new discovery rewards investigators with intriguing questions. Coastal geomorphology is no exception. The variety of the earth’s coastal forms coupled with the dynamics of the forces operating on them will tax the expertise of coastal geomorphologists far into the future.

Cross-References

- [Beach Processes](#)
- [Changing Sea Levels](#)

- Coastal Changes, Rapid
- Coastal Climate
- Coastal Management Practices
- Geohydraulic Research Centers
- History, Coastal Protection
- Monitoring Coastal Geomorphology
- Physical Models
- Sea-Level and Climate Change
- Shoreline and Coastal Terrain Mapping

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History, Coastal Ecology

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Introduction

The history of the investigation of the way in which coastal habitats and species interact with each other and respond to human intervention is relatively recent. Studies of the distribution of individual species populations, pattern and process in species and groups of species, and the relationship with the physical environment, has mostly taken place during the last century. An ecosystem approach which embodies all of the above and lays stress on the complexity of the interactions, is the most recent manifestation of the developing science. Nowhere is this more obvious than in the study of the ecology of coastal habitats, species, and coastal systems.

Some of the earliest work involved quantitative descriptions and successional studies of vegetation. These were undertaken around the end of the nineteenth century, and included work on the Dutch Wadden Sea islands and on sand dunes in Germany. Studies in Denmark also took place on dunes and were the forerunners of the study of plant ecology (van der Maarel 1993).

This early work was carried out on habitats which were mostly, apparently free from human interference. Thus, coastal salt marshes and sand dunes with seemingly simple successional sequences were amongst the first to be studied in relation to the way they responded to natural forces. Understanding of the role of human activity was not considered to be a priority. This situation changed dramatically in last half of the twentieth century as human populations put increasing pressure on marine and coastal ecosystems throughout the world. This entry attempts to describe some

of the ecological principles which have been developed over the last 100 years or so, and discusses how they are applied to coastal systems today.

Early Work, Unravelling Complex Coastal Systems

Studies of the autoecology and ecophysiology (of plants) dominated early coastal ecological work. Salt marshes and sand dunes were obvious candidates for studying ecological change because of their seemingly regular patterns of succession and their dynamic nature. Some of the best known British ecologists of the first half of the twentieth century, for example, included detailed work on these coastal habitats (Tansley 1949; Salisbury 1952). The early descriptions were concerned both with the pattern of plant communities and also the process through which these patterns developed (e.g., Chapman 1976). Thus, coastal vegetation became recognized as a series of types progressing from early pioneer stages to more complex forms which could be related to the physical parameters affecting their development. The role of animals in this process was not considered and studies of animal populations continued along a largely separate path.

Plant Zonation and Succession

Zonations of seaweeds can be related to tidal influence and/or wave exposure, and on rocky shores these can be very

pronounced (Stephenson and Stephenson 1972). Similar patterns can be discerned for many coastal, terrestrial habitats, though the reasons for them may be less easy to interpret. Early studies looked at the role of individual plants in overcoming the rigors of what was considered to be a hostile environment, in an attempt to establish whether there were recognizable factors determining the sequence of development. *Suaeda fruticosa*, a plant of gravel shores, for example, was shown to be able to establish itself over time by slowing down the landward movement of the beach (Oliver and Salisbury 1913). On sandy shores pioneer plants and other obstacles do the same by arresting the movement of sand grains (Fig. 1). The growth habit of the plant itself, including its root system, soil water, and mineral nutrient relationships were the subject of classic studies on sand dunes (Salisbury 1952) and helped to define the adaptations of individual species in overcoming environmental perturbations. Similar studies were carried out by others on sand dunes and salt marshes (e.g., Chapman 1934).

The mechanism through which succession took place was perhaps most clearly elucidated for salt marshes. Primary colonizers such as *Salicornia europaea* *Suaeda maritima* (in north-west Europe) become established on accreting sedimentary tidal flats. The pioneer plants, tolerant of immersion in seawater, were shown to be dependant on periods in the early stages of plant establishment when they are free from tidal movement. As sediment height increases, the marsh is subject to progressively fewer tidal inundation, less sediment



History, Coastal Ecology, Fig. 1 Strandline and pioneer sand dunes, Magilligan Point, Northern Ireland



History, Coastal Ecology, Fig. 2 Parallel zones on a salt marsh. *Spartina* expansion following erosion, Severn Estuary, England/Wales

is deposited and a richer complement of plants and animals replaces the specialist salt-tolerant species (Ranwell 1972). Thus, from an ecological point of view salt marshes provided evidence of primary succession, with the development of a parallel spatial zonation (Fig. 2) which could be related to tidal inundation. The fact that this occurred apparently “largely without human interference” made them ideal for studying the processes associated with “natural” vegetation development.

By 1972 when Ranwell published his book on the ecology of salt marshes and sand dunes recognition of the importance of understanding the complex relationships, which determined the nature of salt marshes and sand dunes, was much more clearly understood (Ranwell 1972). These and other early studies showed that there are zonations attributable to environmental gradients in coastal vegetation. Anyone looking at the early stages in salt marsh or sand dune growth will need little convincing that this is so. However, unravelling the precise relationships was much more difficult than it at first appeared.

Salt marshes not only respond to tidal movement, but also to relative sea-level change. The fact that these operate over very different temporal scales is an important factor in understanding the mechanisms involved. The movement of estuary channels, which may cause erosion operates over yet a different timescale. The development of salt pans or the effects of rotting seaweed cause changes to the surface vegetation. Added to this

is the fact that succession may be cyclical as sediments are exposed when erosion takes place.

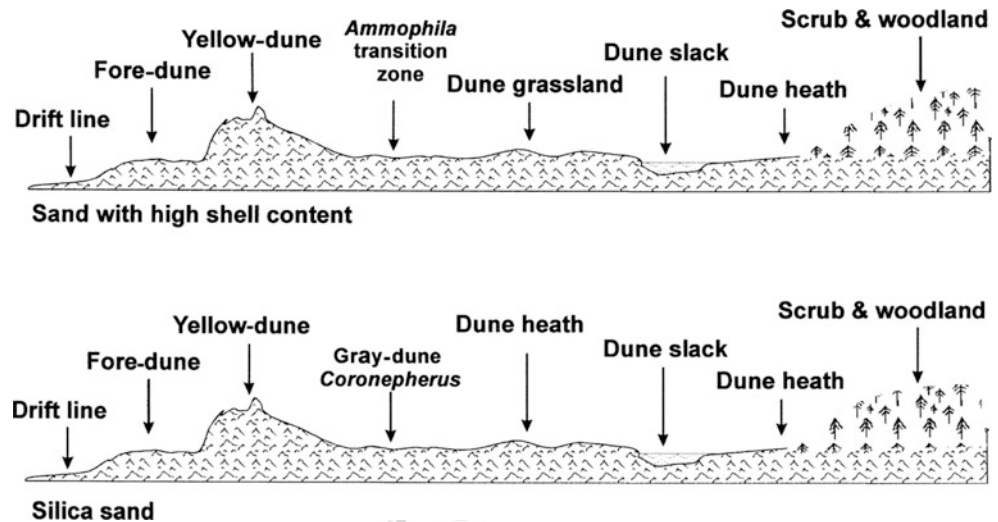
Sand dunes also show forms of successional development in the early stages of growth by the accretion of sand, aided by specialist plants. However, once the main body of the dune is formed other processes come into play and the change from mobile foredunes and yellow dunes to grassland, heath, scrub, and woodland is rarely a straight forward progression, though it may be depicted as such (Fig. 3). Blowouts occur with or without the intervention of man and can be the precursors of dune slacks. Similarly, the reprofiling of dune ridges under the influence of changing wind patterns bring an infinitely variable topography, the origins of which may be difficult or impossible to unravel.

Coastal Networks

As the mechanisms through which the habitats developed were unravelled, so the sheer complexity of the relationship with the “natural” environment became more apparent. Individual species relationships, to some extent, provided only a first level in understanding of the way in which coastal systems and their biological components operate. The physical conditions in which plants and animals exist provide powerful controls on populations, especially where there are major perturbations in these conditions. Superimposed on this are the various natural cycles involving predator–prey relationships and intra- and inter-specific competition. Thus, it is not surprising that as the study of ecology developed, these

History, Coastal Ecology,

Fig. 3 A sequence of sand dune succession on calcareous and acidic sand. Notice the dune building phase (high dune) followed by deflation and the development of more stable communities



complex relationships came under scrutiny. These interactions are the “stuff” of ecology and the studies have been many and various. Nowhere is this more apparent than in coastal wetlands.

Nutrient cycles and energy flows within wetland communities are key and often quoted examples. Attempts to describe the complex interactions between detritus, nutrient inputs, primary producers, detritus feeders, predatory fish, and birds are usually presented as “generalized” food webs. However, these pictures can only represent a very simplified view of the systems involved. A detrital food chain in a mangrove forest alone, for example, may include 11 different groups of animals with untold species involved.

and history of change on the coast was undertaken and described in two classic works (Steers 1946, 1972). These provided a much better understanding of the coastal landscape and its conservation needs and helped lay the foundation for much of the subsequent coastal policy. In America, comprehensive accounts of the development and the processes by which coastal systems came into being were led by geologists (e.g., Johnson 1919). More recent academic publications look at examples from around the world and provide more detail, not only of the way in which the coast responds to the many natural driving forces, but also the influence of human use in shaping its condition (Bird 1984; Carter 1988; Carter and Woodroffe 1994).

The Importance of Geomorphology

Engineers were perhaps amongst the first to appreciate the interaction and complexity of whole coastal ecosystems. Writings early last century described the way the coast responded to natural conditions such as sea-level rise, tides, storms, wave action, wind, and rain and how these affected the shoreline (Wheeler 1903; Carey and Oliver 1918). Later studies were concerned with coastal erosion and land reclamation (Du-Plat-Taylor 1931; Matthews 1934). These helped lead the way to unravelling the relationships between the physical and biological components of the coastal environment, including an appreciation of the dynamic nature of the habitats involved.

The description of coastal landforms and the factors which have helped to shape them come under the general heading of geomorphology. Some of the most important texts, which have helped secure a better understanding of the coast and its management needs come from these studies. In Great Britain, during the 1940s and 1950s, a survey of the geology

Human Occupation and Use

As has already been intimated above, from an ecological perspective, coastal habitats are often considered amongst the more natural ecosystems. This has led to the impression that the process of succession takes place in a sequence which is determined largely by natural forces. The classic studies of the salt marshes and sand dunes on the North Norfolk coast (Chapman 1938, 1941, 1959) or the Dovey Estuary in west Wales (Yapp et al. 1917) emphasized the natural status of the vegetation. However, this hides a long history of human use. On the sedimentary shorelines of the Wash, for example, artificial embankments to enable salt-making were present in some numbers in Roman times (Simmons 1980). Hay making, oyster cultivation, turf and reed cutting, and samphire gathering all take place or have taken place on upper marshes throughout Europe (Dijkema 1984). The deliberate planting of *Spartina anglica*, itself a hybrid fashioned from the interaction of a native and an introduced species in southern England (Marchant 1967), has been a major influence

on salt marshes throughout the world. Occupation of sand dunes probably dates back several thousand years as archeological studies have shown. Many sites have examples of middens with the remains of shell fish in them. An analysis of flint tools suggests that settlers of Torrs Warren, a sand system in south-west Scotland, may have appeared between 5500–7000 years ago (Coles 1964). It seems that cultivation has taken place since then and in 1572 three farm houses were recorded on the Warren. As long as 4500 years ago a small settlement existed on the shores of Skaill Bay in Orkney. The Neolithic people which inhabited the site, known today as “Skara Brae,” seem to have been agriculturists as well as hunters/gatherers (Ritchie and Ritchie 1978) living on the coast until their village was overwhelmed by a sand storm. Since Medieval times, dunes have been used extensively as rabbit warrens and are grazed by domestic stock to the present day. Research on the origins of the deltas of the Mediterranean show a link between population growth and decline, deforestation in the hinterland, and the growth and retreat of the deltas themselves.

This knowledge has altered our perception of coastal habitats. Far from being natural, many coastal systems are highly modified by human activity. Their ecology and conservation thus depends both on understanding the “natural” processes by which they develop and the impact of human action (Doody 2001).

Conclusion

Unravelling the ecological interactions of species and their environment has been the main thrust of ecological studies for most of the last century. It has been suggested that this has been largely undertaken without any direct investigation of the human factor and its impact on the individual elements in the coastal environment, whether they are concerned with vegetation succession or change in animal populations. As our understanding of the relationship between the physical and biological components of the coastal environment has grown, so has concern for the impact of human use both on the coast itself and its ability to sustain human uses. This has brought into sharp focus the need for a more integrated approach to the study of coastal ecosystems. At the same time the role of ecology has moved substantially from the study of systems “apparently” free from human interference, to one where socioeconomic forces are equally, if not more important, to understanding how coastal landscapes function.

Sustaining Coastal Habitats

The study of ecology has also formed the basis for the identification, protection, and management of areas of nature conservation importance. Indeed, it was the preeminent ecologists of the day who helped develop the present approach to

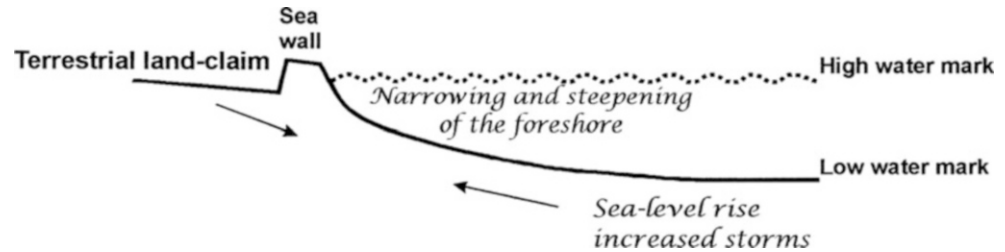
the protection and management of nature reserves. In the United Kingdom, for example, in 1942 the British Ecological Society established a Nature Reserves Committee under the Chairmanship of Sir Arthur Tansley (Sheail 1987). It was this committee which provided the foundation for the series of Nature Reserves and other protected areas that are so important to the conservation of the remaining areas of natural and seminatural habitats today.

In the early postwar period, the understanding of successional processes, the concept of climax vegetation, and the notion of zonation, helped to lead conservation managers toward a preoccupation with the protection of existing interests, identified when the sites were first assessed. To many the “naturalness” of the system was the prime reason for its conservation. However, as has been argued above, many so called “natural” systems are not natural at all. Most if not all, in temperate regions at least, have in some way been modified by human activity. The extensive unmodified lagoons and deltas with their apparently natural salt marshes and sand dunes in Albania and eastern Turkey, grow rapidly today because of deforestation in the hinterland. Even here drainage and land “improvement” for agriculture has destroyed large areas of transitional vegetation and pollution is a major consideration.

This understanding has important consequences both for the development of conservation policy and the ecological principles upon which it is based. Agenda 21, a program for action agreed at the United Nations, Earth Summit in Rio in 1992, aims to help achieve the twin goals of sustainable human development and the maintenance of biodiversity. The statements clearly point the way toward recognition of the interrelationship between the so called “natural environment” and human economic and cultural activity. If policies are to be developed which fulfill the aspirations both the politicians and the traditional conservationist, future ecological studies must include human use as a key component of the “natural” system.

The coastal squeeze. An important component of human use centers around the loss of coastal and tidal land. The process by which this affects the margin between the land and the sea can be described as “coastal squeeze.” Here human action pushes the limits of intensive use of the land (for agriculture, industry, and the like, or land claim of tidal areas), toward the sea. At the same time, where sea level is rising relative to the land, the upper limits of tidal influence are pushed landward (Fig. 4). Taken together, these effects cause a narrowing of the shore and the loss of coastal habitats. In its turn this can result in a reduction of the capacity of the shore to withstand, and recover from, episodic events including major storms. In low-lying coastal areas, for example, coastal properties are put at greater risk from flooding and/or erosion, as the protection afforded by a wide beach is reduced, as it becomes steeper and narrower.

History, Coastal Ecology,
Fig. 4 “Squeezing” the coast



The cumulative effects of habitat loss may also reduce the ability of the coast to recover from major environmental perturbations. This may include difficulties in replenishing living and nonliving resources. In their turn these result in an inability to sustain the economic fabric of some areas.

A question of sediments. Other factors, which have been increasingly recognized as being important to the sustainable use of the coast, include the nature and availability of sediments. In this context, there is also an increasing recognition that the reduction in availability of sediments, whether due to offshore exploitation, damming of rivers, or reducing the flow from longshore drift, has a predictable outcome. Many deltas have grown through the transport and deposition of sediments eroded from the hinterland following deforestation. Studies suggest that this situation has been reversed as the damming of rivers has reduced the available sediment supply. As a result, today the outer margins of many deltas are eroding as they become wave-dominated rather than ones where freshwater river flows exert the major force. This pushes the margins of the delta landward and with it increases saline water intrusion into the underground aquifer. The implications for the continued economic use of these areas for agriculture (e.g., rice cultivation), ground water abstraction, or problems associated with erosion are largely ignored by the political and economic forces, which continue to dictate land use policy and coastal management.

Ecological Change as a Healing Force, a New Paradigm?

Change is an important part of the development of coastal systems. This was most dramatically revealed by major events, such as the landslide between Axmouth and Lyme Regis (Devon, Dorset, southern England) on Christmas Day 1839, which created a chasm 1 km long and up to 122 m wide, taking with it a small village. The steady erosion of the cliffs at Dunwich (Suffolk, England) has now thrown all seven of the Medieval churches of this once-important port into the sea. At the same, time saltmarsh erosion and accretion are relatively natural phenomena, taking place in response to changes, for example, in the location of the tidal channels or, over a longer time period, in sea level. Recognizing these changes as factors to work with rather than against, may provide a more sustainable approach to coastal development.

In this context, the alliance between the ecologist, geomorphologist, and coastal engineer in re-creating new coastal habitats may be a first step toward accepting change as a means of securing more sustainable living on the coast. For example, the re-creation of salt marshes, or other tidal wetlands from land given back to the sea may not only secure new nature conservation opportunities, but also provide a more sustainable sea defense. Perhaps, ultimately we will emulate our ancestors, and in some areas initiate major change and then step back and let nature take its course. Under these circumstances the role of the ecologist may be paramount in predicting the outcome of a particular management option. This will require a marriage between the traditional approach which sees the ecological process as natural, and a more pragmatic one where human factors play a significant role. This will not reduce the importance of the “natural” succession, but set it in a wider human context. In the same way that our ancestors helped fashion the present coast and its biological and non-biological resources by accident, we can do it by design!

Cross-References

- [Deltas](#)
- [Dune Ridges](#)
- [Estuaries](#)
- [Europe, Coastal Ecology](#)
- [Monitoring Coastal Ecology](#)
- [Rock Coast Processes](#)
- [Salt Marsh](#)
- [Vegetated Coasts](#)

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History, Coastal Protection

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Though the claim of the Frisians to have been the first to devise an embryonic system of coastal defense is probably

appropriate for the northern European area, others may equally assert their right to first place; the history of coastal engineering can be traced back in China to the East Han dynasty era. Indeed large coastal defense projects were initiated between about 25 and 220 before our era (Xu 1993). In the Mediterranean, likewise, coastal engineering had an “early start” particularly among Greeks, Etruscans, and Romans, and also Carthaginians, Minoans, Phoenicians, Sumerians, and Egyptians.

Classical Times

Dikes were built under the reign of Apollonios Ptolemaeus II Philadelphus, Egyptian ruler, during 259–258 bc. He donated to his employee Stothoetis, a large section of land in Ghoram, Fayoum (Philadelphia); where the owner decided to exploit the embankments by establishing a network of canals and dikes, well before “westerners” drained land in more northern areas. A cursory examination of a picture of the Alexandria Lighthouse, one of the seven wonders of the world of the classical times that crumbled to the bottom of the sea due to a fourteenth-century earth tremor, shows rather sophisticated dikes around Pharos Island. A French expedition funded by the Electricité de France brought back to the surface, in 1998, the colossal statue of Ptolemaeus II that stood in front of the lighthouse.

Coastal defense history is, naturally, closely linked to harbor creation and development. Primitive breakwaters were put in place occasionally with ramps allowing the top of the waves to pass over them. The Phoenicians had devised wave catchers by excavating holes and establishing trenches in the rocks lining the shores (Raban 1988). Carved breakwaters were cut out of bedrock: a suitable wave-absorber profile was created which had a gentle grooved slope at the waterline. The classical example dating from the second century bpe is still visible at Ventotene (Italy), it had an overspill. One may see in it an early version of today’s Fontvieille breakwater (Principality of Monaco).

Greek and Etruscan breakwaters and seawalls consisted of rubble mounds topped by cut rocks; though no mortar was used, neighboring blocks were sometimes held together by clamps and joints made out of metal. With the discovery of pozzolanic ash hydraulic-cement, solid breakwaters could be built underwater, and several vertical composite concrete walls have been preserved from the second century to the fifth century before our era. There are illustrations of a vertical breakwater made in situ within a wooden frame and tie-rods. Toe protection against scouring was provided occasionally by a bronze slab (Oleson 1988).

Not only had these engineers mastered the art of erecting cofferdams for construction “in the dry,” but they also thought of caissons, forerunners of contemporary building methods.

Watertight wooden cellular caissons were used to cast large concrete breakwaters, for example, at Caesarea. “Permeable” breakwaters and “arched moles” were installed in various sites (Franco 1996a). Apparently well before the Dutch dyke-builders thought of sinking old ships, the Romans sank old hulls, filled them with concrete and had a breakwater placed in no time, under Claudius’ reign, Caligula’s “monster” ship was sunk at Ostia (50) to provide a breakwater (Testaguzza 1970). Remnants are still visible at 4 km from the Fiumicino airport.

Trajan built a rubble mound breakwater on an “island,” reshaped by nature and workers leading to a subsequent mild slope profile (Franco 1996b).

Medieval Times and Renaissance

Grillo (1989) reports that the earliest written document dealing with shore protection dates back to 537 when fagines, wicker faggots, supplemented by timber piles and stones, held up earthen dikes adding their protection to that of the dunes.

“Timber and rock revetments and groynes have been used (in the Venice area) until 1700 to halt beach erosion and silting” notwithstanding a lengthy transport for rocks and short life span of wood (Franco 1996b). Strict environmental regulation governing shore protection can be traced back to legal documents of 1282 and 1339: prohibition to cut or burn trees from coastal forests, to pick mussels from rock revetments, to let cattle walk the dikes, to remove sand, and vegetation, from beach or dune, and to export materials used in coastal defense (Grillo 1989). The *Magistrato alle acque*, created in 1501, invited suggestions to reduce the high cost of the coast defense, so, in the eighteenth century, there appeared rip-rap revetments, gabions, staircase-placed limestone blocks, and the use of mortar and steel links and flexible steel strips became common.

Whereas beaches protect the littoral, they had to be maintained, and artificial nourishment with offshore dredged sand was initiated as early as the seventeenth or eighteenth century, hence long before California used the method (1919).

Massive *murazzi*, devised by Zentini and his team were constructed from 1741 on; with an average width of 12 m (39 ft), their crest peaked at 5 m (16 ft) above mean sea level. It took 40 years to construct a protective seawall 20 km (12.4 miles) long (Charlier and De Meyer 1998). *Murazzi* have withstood the test of time, though storms took their toll and toe protection was eventually provided by a rubble mound structure and, more recently, jet-grouting diaphragms. Commenting on the Genoa breakwater, an actual fortification with a superstructure, Franco (1996b) underscores its importance as in 1245 it was proclaimed a “pious work” thereby compelling every citizen of the Republic of Genoa to provide in his will for the breakwater’s maintenance.

Of course no Renaissance technology review can pass over Leonardo da Vinci whose talents included hydraulics and is the father of a proposed triangular-shaped island breakwater. He as well championed the credo of “working with Nature,” rather than against it: *ne coneris contra ictum fluctus: fluctus obsequio blondiuntur* (Nature should not be faced bluntly and challenged, but wisely circumvented).

Franco (1996b) has virtually provided a catalog of Italian designed breakwaters: use of irregular blocks with pozzolanic concrete crown and large rock armor porosity, (Crescentio in 1607), a monolithic superstructure over a leveled rubble mound foundation; (De Mari in 1638), armored with precast blocks (San Vincenzo mole at Naples, 1850), vertical composite structures (1896), and caisson construction (1915, 1931, 1936, 1938, 1995).

Contemporary Approaches

If new approaches to coastal defenses were slow, and little new technology was introduced during the centuries-though improvements and refinements were often made-the pace of change accelerated considerably in the twentieth century, even more so during the last decades. The large beach nourishment achievement in Belgium (1980s) has been surpassed, profile nourishment and berm feeding have been implemented-aimed simultaneously at restoration and protection-not less than 40 alternative methods have been proposed and tried out. But the problem has not been solved; for instance, all 30 of the US coastal states suffer from erosion and some see Hawaii’s tourist industry in jeopardy.

At the turn of the twentieth century, the response to coastal erosion remained the construction of hard defense structures: groins, jetties, breakwaters, revetments, gabions, placement of tetrapods, and the like. However, the groins placed at Miami Beach and Long Island N.Y. did not stall retreat of the shore. The same situation prevailed in Europe where beaches were shrinking in Denmark, Germany, The Netherlands, Belgium, and France. The Mediterranean beaches were not spared either. The problem is worldwide. Variants were tried out, sometimes meeting limited success, that is, floating, permeable, offshore breakwaters. If hard structures do protect or extend beaches on one face, the downdrift side is starved.

Beach planners, engineers and geologists proposed to artificially re-nourish beaches and the approach was tried on northern California beaches in the early 1920s. Since then the “re-charging” of beaches has been carried out around the globe, wherever there were sufficient funds to undertake an operation that is not inexpensive. Major schemes were undertaken in the United States, Belgium, and France. The beaches require regular additions of sand and in some cases a major storm may carry back to sea a major part of the artificial deposit.

Artificial beach nourishment entails many operations among which selection of the materials and of the source spot, method of material transportation, study of the waves and weather climate, of the physical and biological impacts. Improvements on the simple method of direct material dumping have been sought and thus appeared profile feeding, establishment of a feeder berm (e.g., De Haan, Belgium), and combinations of hard and soft defenses. Beach dewatering has been presented as a new alternative (e.g., Carolinas in the United States) but in fact it is more a complementary than an alternative method. At any rate it helps a beach retain the nourished material for a longer span of time. Sand back-passing and by-passing have proven valuable approaches (Hillsboro Inlet, Florida; Durban, Republic of South Africa); sand is transferred from one side of a structure or formation to the other which is starved. In compensation dredging, a somewhat similar method, the material is dredged and carried to another site that needs to be nourished.

There are over 40 methods that have been proposed and/or patented to halt coastal erosion. Though most have merits, most also have objectional side effects or hard-to-accept environmental impacts. Two of them, Berosin[®] and Beachbuilder[®], apparently are free of them but are, at least temporarily, unaesthetic. The latter of the two attempts to promote beach accretion by using the power of the erosive waves to build up the beach.

Still other approaches have been tried: artificial reefs, artificial or restored dunes, fields of algae or synthetic fronds, creation of inlets for seawater. However, one should not lose sight of the fact that shore landward migration is inexorable, that people can only slow down the inland progress of the sea, and that only nature can definitely reverse the trend.

Conclusion

In the 19th and 20th centuries, coastal protection against an advancing sea has been centered on a variety of hard structures (groins, breakwaters, seawall, tetrapods, etc.) and artificial nourishment. Environmental concerns have steadily played a more important role. Some 40 types of alternative schemes have been proposed over the last decades, with a large number of them faulted for negative environmental impacts. As the economic consequences of the landward migration of the shore are often disastrous, the search of solutions remains.

Cross-References

- [Beach Drainage](#)
- [Beach Nourishment](#)
- [Bioengineered Shore Protection](#)

- [Coastal Management Practices](#)
- [Dredging of Coastal Environments](#)
- [Engineering Applications of Coastal Geomorphology](#)
- [Human Impact on Coasts](#)
- [Navigation Structures](#)
- [Shore Protection Structures](#)
- [Shoreline Response to Littoral Drift Barriers](#)

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Holocene Coastal Geomorphology

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Definition

Holocene coastal geomorphology is a geoscience discipline dealing with materials (rock, sediments), forms, and forming processes (anorganic and organic) in the overlapping fringe between marine and terrestrial environments (called “littoral”) during the Holocene period, in particular the time frame of a postglacial high sea level over the last about 7000 years to modern times. It is a wide field of science, with certainly tens of thousands of related publications. A text on this subject therefore has to be restricted to some general points.

Normally forming processes on Earth need time measured in geological scales (i.e., many thousands to millions of years) to show adequate results with mature forms. In contrast to this general rule, all coastal forms are not older than 6000–8000 years, when oceans again reached modern levels

from a low-stand near -120 m during the last cold phase with extended glaciers (last glacial maximum (LGM); Nicholls and Cazenave 2010; Scheffers et al. 2012; Shennan et al. 2012). The Holocene as a geological epoch, however, started 11,700 years ago with the shifting to modern climate and temperatures, but it lasted several thousand years to melt the extended ice sheets and fill the oceans again. Therefore, coastal forms we see today belong only to the last phase of the Holocene. There might be some inherited features incorporated in them, formed during former sea-level high stands in warmer periods of the ice ages, the last one dated to about 125,000 years ago.

The world's coastlines are the most extended geomorphological features on Earth, measuring several 100,000 km on small-scale maps, but more than one million km in nature. They are present in all geotectonic and geodynamic situations, along all petrographic units of hard rocks and loose sediments and in all climatic and biogeographic provinces of the world. Therefore, their geomorphological inventory and the amount of forming processes are numerous.

Beside the tectonic situation (stable, subsiding, rising) and the type of rock (more or less resistant, massive or stratified, limestone or silicate, hard rock or loose sediments), the kind of wave impact along a certain coastline is the most important factor. It depends on facts like nearshore water depth, strength of wind, tidal range, and more (Dillenburg and Hesp 2008; Lim et al. 2011).

These elements all have a wide range, in particular the tides, from micro-tidal (less than 1 m) to macro-tidal (with spring tide ranges exceeding 10 m like in southern England and along the French coast, or even more than 15 m in Fundy Bay of Eastern Canada). These tides are distributing wave energy and currents on a wide horizontal area, resulting in tidal flats exposed during low water, marshes from high water and storm deposition, or mangrove fringes in warmer latitudes. Surf beat with effects on coastal destruction (abrasion) depends – beside water depth – on strong winds from one direction over a long time and on a wide ocean area (called fetch) and from the availability of particles as abrasive agents (Erdmann et al. 2015; Goto et al. 2015; May et al. 2013; Nel et al. 2014; Paris et al. 2011; Scheffers et al. 2012; Short 2006; Trenhaile 2008). Zero energy coasts in calm climates differ from those with gale forces and surf waves more than 10 m high around the southern parts of South America, South Africa, Tasmania/New Zealand, or Western Europe. Even in warmer latitudes, hurricanes may occur with similar geomorphological effects. They usually come with a storm surge, i.e., rising water driven by strong wind against the coastline, and additionally rise during low-pressure situations. This has the same effect of flooding on coastal lowlands as winter storms in higher latitudes. In direction of the highest latitudes, sea-ice cover may suppress wave action during many months of the year (Lantuit et al. 2013). This ice, however, has

geomorphological effects moving along a shoreline during break up in springtime.

Beside the influences of the lithosphere, atmosphere, and hydrosphere, the biosphere at the coastlines of the world will promote forming as well. There are mangrove coasts, coral reef coasts, or those with dominant kelp, seagrass, or driftwood (Hopley 2011; Scheffers et al. 2012).

All in all the world's coastlines show a latitudinal zonality from sea-ice belts to coral reefs, depending on climate and water temperature in the first order and on aridity or humidity, in the second order.

Even small secular sea-level variations (by neotectonic or isostatic movement of the land, or eustatic changes of the water body itself) have a strong influence on coastal forms: relative rising sea levels lead to more abrasion or drowning and falling sea levels to emergence and accumulation. Congruent to the ongoing sea-level rise during the last one to two centuries (also caused by anthropogenic greenhouse effect with global warming), a significant erosion of beaches and coastal dunes as well as wetlands occurs (Martinez et al. 2008; Nel et al. 2014; Provoost et al. 2011; Shennan et al. 2012; Short 2006). Holocene sea-level variations, however, are still under debate. It seems that there are large provinces with different sea-level stories depending on geotectonic factors as well as eustatic ones, sediment load, compaction, and more.

Similar in all histories of modern coastal forms is a rising sea level with more than 1 cm/year on average in early Holocene periods (i.e., about ten times faster than today's rise), but during oscillations, many cm/year also occurred from the last glacial maximum to the present warm phase. This transgression on formerly terrestrial environments did barely leave marked forms but reworked the weathered mantle on the drowned landscapes finding abundant coastal sediments. The transformation of the landscape by waves, however, was minimal, and the type of "primary coasts" with partly submerged terrestrial forms dominated nearly everywhere. After reaching the modern level about 7000 years ago, the balance between marine processes on one side and terrestrial ones on the other side was concentrated along a fixed level, leading to a significant transformation and sculpturing the many types of "secondary coasts," dominated by littoral processes.

The most striking evidences for abrasion and coastal destruction are cliffs, stacks, sea arches, rock platforms, and sea caves. Their amount of retreat depends on surf energy and rock resistance and can reach several 100 m during Holocene times, leaving slightly inclining rock platforms in front of the cliffs. There is not only surf that destroys the coasts, but salt weathering (Trenhaile 2008) or bio-erosion may be important in certain environments as well.

Advancing coasts show beaches with accompanying coastal dunes (regionally cemented into beachrock or

eolianite in more arid environments), sometimes formed into spits or barriers (or chains of barrier islands in regions with higher tidal range), or marshlands. Deltas develop even along submerging coasts, if sediment discharge is strong enough. Beside the destructive and constructive coastal forms, the category of ingressive features is very variable, representing all relief forms of the Earth in a partly drowned status (fjords, rias, tropical karst, etc.; see Scheffers et al. 2012).

A fully genetic classification of the world's coastlines is presented in Table 1: All belong either to the class of advanced or retreated coasts. Advancing may be caused by emergence or construction/accumulation, retreating by submergence or destruction. Next categories point to organic or inorganic processes or the type of partly drowned terrestrial relief's, tidal range, etc. (Table 1).

The coasts – in particular the accumulative ones – are important archives for even small environmental changes, because their ecosystems are very sensitive. They more and more fill the gap between terrestrial proxies (from inland ice, peat, lakes) and oceanic ones (from deep-sea sediments).

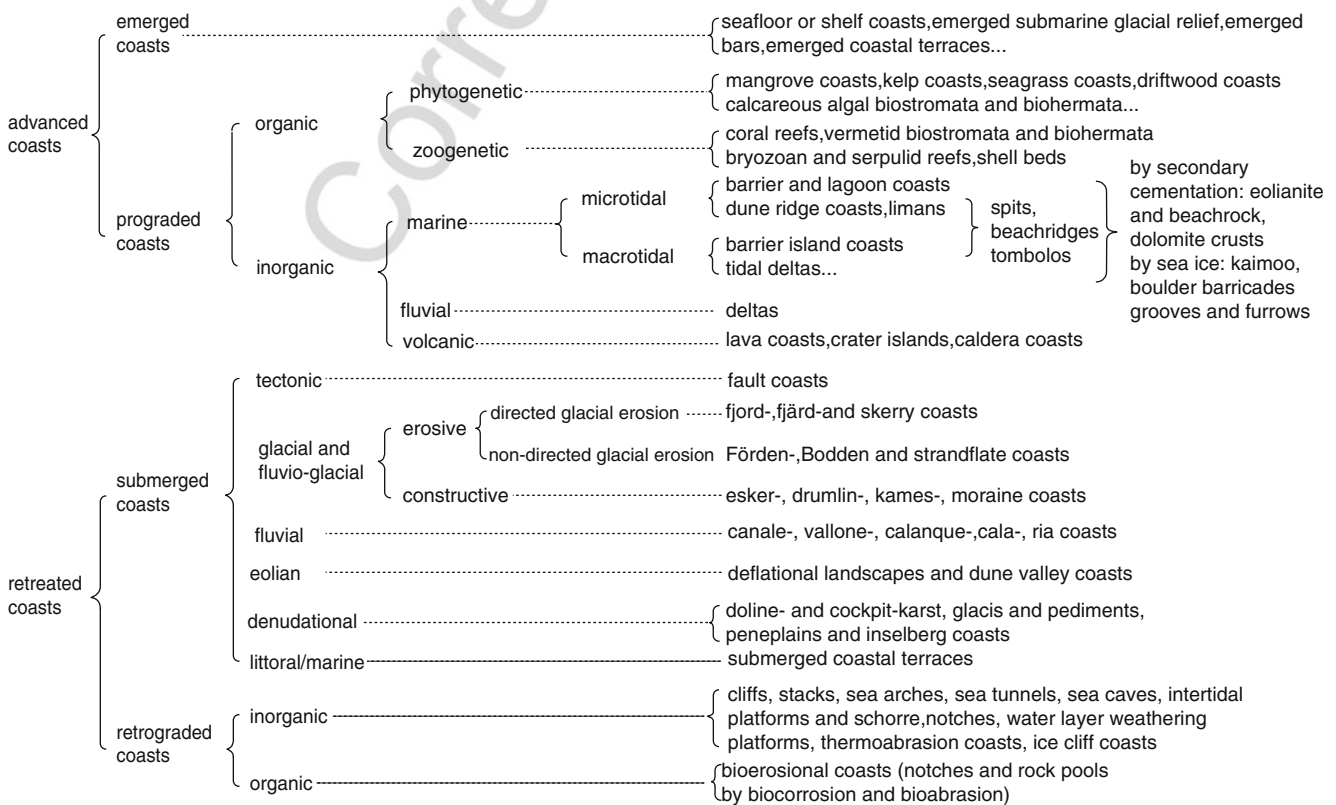
In a worldwide scale, the coastlines are more and more transformed by man, either directly (land reclamation, coastal protection with sea walls and dikes, mining of beaches, aquaculture in mangrove areas, and others) or indirectly (e.g., drilling oil or water from the ground), with the consequence

of subsidence (Mississippi delta, Venice). Catching river sediments in reservoirs far from the coast has the consequence of starving beaches, industry and agriculture may pollute near-shore waters by chemicals and pesticides, and sediment suspension from deforestation and agriculture may contribute to the destruction of coral reefs. A general problem is global warming by emission of greenhouse gases, in particular by permafrost decay forming “thermo-karst” features (Lantuit et al. 2013). Studies on geo-archaeology contribute to man-made influence on coastal variations over a long time back (Jöns and Harff 2014; Marriner and Morhange 2007).

Conclusion

Textbooks on coastal geomorphology mostly give the impression of a consecutive, slow forming in coastal environments. New research brings into discussion whether sudden events of extreme energy (like tsunamis) maybe as important or more important for the development of some coastal forms and deposits (Bryant 2008; Erdmann et al. 2015; Goto et al. 2015; Kelletat et al. 2007; Scheffers 2006, 2008; Scheffers et al. 2012). All in all many questions in coastal geomorphology remain to be solved, as the coastlines of the World are not adequately investigated so far.

Holocene Coastal Geomorphology, Table 1 Genetic classification of the world's coastlines (Kelletat 2013)



Cross-References

- [Beach Processes](#)
- [Bioconstruction](#)
- [Bioerosion](#)
- [Changing Sea Levels](#)
- [Cliffs, Lithology Versus Erosion Rates](#)
- [Coral Reefs](#)
- [Deltas](#)
- [Eustasy](#)
- [Isostasy](#)
- [Rock Coast Processes](#)
- [Surf Zone Processes](#)
- [Tectonics and Neotectonics](#)
- [Tides](#)
- [Tsunami Deposits](#)
- [Waves](#)

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Holocene Epoch

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The Holocene, or “wholly recent,” Epoch is the youngest phase of earth history. It began when the last glaciation ended, and for this reason is sometimes also known as the post-glacial period. In reality, however, the Holocene is one of many interglacials which have punctuated the late Cainozoic Ice Age. The term was introduced by Gervais in 1869 and was accepted as part of valid geological nomenclature by the International Geological Congress in 1885. The International Union for Quaternary Research (INQUA) has a commission devoted to the study of the Holocene, and several International Geological Correlation Programme (IGCP) projects have been based around environmental changes during the

Holocene. A technical guide produced by IGCP Subproject 158B (“Palaeohydrological Changes in the Temperate Zone”) represents a comprehensive account of Holocene research methods (Berglund 1986). Since 1991, there has also existed a journal dedicated exclusively to Holocene research (*The Holocene* published by Arnold).

During the Holocene, the earth’s climates and environments took on their modern, natural form. Change was especially rapid during the first few millennia, with forests returning from their glacial refugia, the remaining ice sheets over Scandinavia and Canada melting away, and sea levels rising to within a few meters of their modern elevations in most parts of the world. By contrast, during the second half of the Holocene, human impact has become an increasingly important agency in the modification of natural environments. A critical point in this endeavor was when *Homo sapiens* began the domestication of plants and animals, a process which began in regions like the Near East and Mesoamerica very early in the Holocene, and which then spread progressively to almost all areas of the globe. For short histories of the Holocene, see Roberts (1998) and Bell and Walker (1992).

Although there are different schools of thought about how the Holocene should be formally defined (see Watson and Wright 1980), the most common view, and one which is supported by INQUA, is that the Holocene began 10,000 radio-carbon (^{14}C) years ago. But ^{14}C chronologies count AD 1950 as being the “present day” and also underestimate true, or calendar, ages by several centuries for most of the Holocene. Nonetheless, there is evidence of a global climatic shift remarkably close to 10,000 ^{14}C year BP (years before present), often involving a sharp rise in temperature (see Atkinson et al. 1987).

Various attempts have been made to subdivide the Holocene, usually on the basis of inferred climatic changes. Blytt and Semander, for instance, proposed a scheme of alternating cool–wet and warm–dry phases based on shifts in peat stratigraphy in northern Europe. Some researchers believe there is evidence of a “thermal optimum” during the early-to-mid part of the Holocene. During the 1980s, the Cooperative Holocene Mapping Project (COHMAP) members established a comprehensive paleoclimatic database for the Holocene (Wright et al. 1993), and showed that variations in the earth’s orbit were the principal cause of differences in climate between the early Holocene and the present day. For this reason, the early Holocene is unlikely to provide a good direct analog for a future climate subject to greenhouse-gas warming (Street-Perrott and Roberts 1993).

Cross-References

- [Geochronology](#)
- [History of Coastal Geomorphology](#)

- [Holocene Coastal Geomorphology](#)
- [Sea-Level Fluctuations over the Last Millennium](#)

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Honeycomb Weathering

George Mustoe

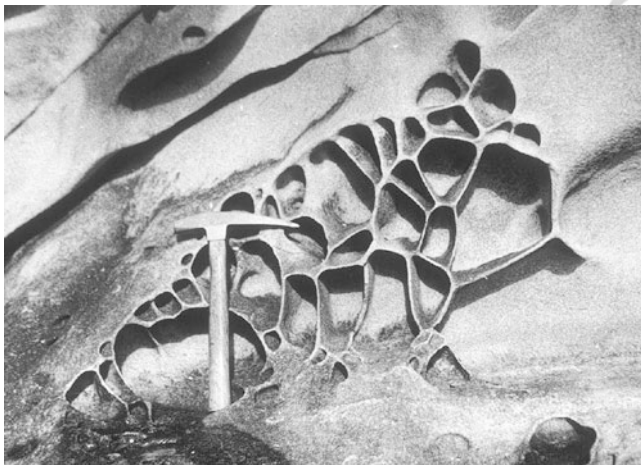
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Honeycomb weathering produces extensive networks of small cavities that form on rock surfaces. These patterns initially develop as many shallow depressions, but continued development produces deep chambers that are separated by thin septa of unweathered rock (Fig. 1). Individual cavities are typically several centimeters in width and depth, the shape often being controlled by bedding planes, foliation, or other structural features of the rock in which they occur. In many localities the holes occur in association with a hardened surface layer formed when dissolution of ferruginous minerals has been followed by precipitation of ferric hydroxides near the outcrop surface. The thickness of this hardened layer may range from a few millimeters to several centimeters.

Honeycomb weathering has worldwide distribution, typically found in coastal outcrops and inland deserts. The most-studied occurrences are those of Australia, the western United States (Fig. 2), and South Victoria Land, Antarctica. This type of weathering is most commonly observed in sandstone, but it also occurs in granite, gneiss, schist, gabbro, and limestone. The mode of formation of these cavities is not well established, and several hypotheses have been advanced. Possibly more than one type of mechanism may be involved.

Honeycomb Weathering,

Fig. 1 Honeycomb weathering in sandstone near Baku, Azerbaijan on the west coast of the Caspian Sea. (Photo, M.L. Schwartz)



Honeycomb Weathering, Fig. 2 Honeycomb weathering in sandstone at Larrabee State Park, near Bellingham, Washington, USA. (Photo George Mustoe)

In formulating any explanation it is necessary to account for the extremely selective nature of the erosion. One possibility is that the rock possesses internal variations in hardness, composition, or porosity. This conflicts with the fact that honeycomb weathering commonly occurs in rocks that have a high degree of physical and chemical homogeneity. Thus, it is more likely that attack by the weathering agents occurs in a differential fashion. Early investigators invoked a diverse variety of geomorphic processes to explain honeycomb weathering, but these cavities are now generally accepted to be caused by salt weathering, where evaporation of wave splash or saline pore water produces salt crystals that wedge apart mineral grains (Evans 1970). Chemical dissolution of silicate minerals may also play an important role (Young

1987). Just how these process works to produce a delicate honeycomb pattern remains an enigma. Mustoe (1981, 1982) believed that coatings of lichens and algae on the rock surface control the pattern of cavity development. Laboratory experiments by Rodriguez-Navarro et al. (1999) suggest that erosion may be related to variations in wind currents flowing over the rough outcrop surface.

Cross-References

- ▶ [Bioerosion](#)
- ▶ [Cliffs, Lithology versus Erosion Rates](#)
- ▶ [Coastal Climate](#)
- ▶ [Coastal Hoodoos](#)
- ▶ [Coastal Wind Effects](#)
- ▶ [Desert Coasts](#)
- ▶ [Notches](#)
- ▶ [Shore Platforms](#)
- ▶ [Weathering in the Coastal Zone](#)

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Human Impact on Coasts

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Definition

Human activity along the coast has both positive and negative impacts on the natural environments. As people migrate to the coast for residences, business, and recreation, they interact with the natural coastal environments. For the most part because the coast is a very dynamic system and changes can occur rapidly, people make the effort to control the nature and magnitude of these changes. The most important of these changes occur on the beach and typically are related to erosion of the beach. Beach erosion is widespread and causes significant problems for both the use of and protection by beaches. Various structures such as seawalls, groins, and breakwaters have been constructed to help protect beaches from erosion. Many of such structures cause more problems than they solve. Such attempts to control the nature of changes on the beach are among the more negative impacts from humans. During the past few decades, beach nourishment has been used to help manage this environment and with general success. Anything that humans do to change or impede the natural processes along the coast tends to have a negative impact on the coast.

Introduction

Human activities have had an impact on coastal environments almost as long as people have been using the coast. It was not long before attempts to control erosion resulted in various types of structures such as jetties, groins, and seawalls. Access to the beaches of barrier islands resulted in fill-type causeways. Development of barrier islands accessed by these causeways resulted in various types of construction that have negatively impacted on the coastal zone. Harbors and the navigational channels leading to them have resulted in some problems for estuaries. Overall, there are many ways whereby human activities have caused problems for a broad spectrum of coastal environments. Some of these are a result of direct acts of development, and others are indirectly the result of these activities. The following discussion will consider some of the more obvious and problematic human impacts. Many of the examples will come from Florida where such development-related activities have been underway for several decades and where we can learn from our experiences.

Direct Impact

Hard Protection

Among the first, the most widespread, and the most problematic human impacts on the coast is the erection of hard structures: those that are immobile and that are not modified by coastal processes. This category includes seawalls, groins, breakwaters, and jetties. Their primary purpose is to protect the coast from erosion and to stabilize tidal inlets. The variety of such structures is tremendous (CERC 1984), and a comprehensive discussion of them would fill a book. The basic characteristics of each general type and some examples will be provided to acquaint the reader with such structures.

Seawalls of various types have been around for more than a century. The basic situation is protection of the open coast landward of a beach from wave attack. Such structures may be poured concrete, metal sheet piling, pressure-treated wood, or various types of riprap: well-placed boulders. Sometimes longard tubes or other sand-filled plastic or cloth tubes are used as temporary protection. Such an approach is viewed as a semihard type of protection. There are problems with these types of structures. Regardless of their type, they tend to be temporary. Eventually, time or severe storms will cause them to fail or become useless for various reasons. Some of the problems associated with seawalls are the scouring that occurs at the base. In many cases, these walls are vertical which takes the full impact of the waves. Riprap placed in front of the seawall dissipates wave energy, but continued wave attack will eventually result in failure (Fig. 1).

Groins are placed perpendicular to the shore in an attempt to keep sediment from being carried away by longshore currents. Most of these are overdesigned and act like dams along the beach causing downdrift erosion (Fig. 2). The best performing groins, also among the largest, are along the North Sea coast of The Netherlands and Germany (Fig. 3).

Jetties and stabilized inlets are also a problem for beach erosion. They act as large dams, prohibiting longshore transport from moving across the inlet. This is, of course, their purpose because they are constructed in order to stabilize the inlet and maintain its navigability. The downdrift erosion produced by such jetties is chronic (Fig. 4) and can only be avoided by some type of sediment or bypass system. Another problem with jetties is that many leak sediments are short enough to allow sediment to pass around the end so that regular dredging becomes a necessary procedure for maintenance of inlet navigation. Attached breakwaters also have similar problems.

Detached breakwaters (Fig. 5) are probably the most benign of the hard open coast structures. These are typically parallel to the shore and are designed to provide some combination of protection from shore erosion and safe mooring for small boats. The negative effects are that they commonly cause a salient to form in the lee of the structure that in many

Human Impact on Coasts,

Fig. 1 Riprap in front of concrete seawall. This is one of riprap's most common types of erosion protection on the open coast environment



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Fig. 2 Groin showing a significant amount accumulation on the updrift side



cases forms a tombolo. This connection between the structure and the shore is an unwanted occurrence, which can sometimes be eliminated by lowering the breakwater to permit some wave action and, therefore, longshore currents to keep sediment from accumulating. In some cases, the shore erodes even in the presence of these structures leaving them in place but without function. Removal is generally very expensive, commonly more than their emplacement.

Soft Protection

More recently there has been a dramatic and nearly total shift to soft protection for erosion control along the open coast. This has come in two basic approaches: (1) beach

nourishment and (2) vegetation and protection of dunes. These approaches have received the endorsement of the engineering and the environmental community alike. The result is much more esthetic shore protection; however, these techniques are not permanent solutions to coastal erosion. The only permanent solution is abandonment of the barrier islands and open coast shores.

Beach nourishment. Although not new as a method of erosion control, beach nourishment became a standard beginning in the early 1980s (NRC 1995). The basic scheme is to artificially rebuild the beach that was removed by erosion and to do it as close to the original, natural beach as possible. The most important factor for such a construction project is the

Human Impact on Coasts,

Fig. 3 Large groin on the North Sea coast where there is little difference in accumulation of beach sand on either side of the structure

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Fig. 4 Jetty at a tidal inlet showing a significant amount of downdrift erosion as the result of the inability of sediment to cross the structure and the inlet



location of a sediment source that is similar to that on the natural beach and that is of sufficient volume to do the job. Most of these projects require an average of about one million m^3 of beach quality sand being placed on the eroding shore area. Some are smaller and a few have been much larger.

For a variety of reasons, the borrow site for such beach sand is typically seaward of the beach rather than landward, although a few upland sources have been used. Most of the offshore sources have been the ebb-tidal deltas associated with tidal inlets and large sand bars or old beach deposits from the present shoreface. Some projects have been nourished using beach quality material dredged from inlets,

but these are typically at the smaller end of the spectrum. In addition to the limitations posed by the availability of beach quality sand, the distance of transport is a major consideration because this is what comprises most of the cost of the project.

The design of the nourishment project is undertaken by coastal engineers and requires considerable input of coastal processes, storm surge levels at the site in question, desired protection, and historical characteristics of beaches at the site. Commonly, the elevation of the construction berm is associated with the surge level of a 10–20-year storm. In many parts of the Gulf Coast of the United States, that would be at about 1.5–2 m above mean sea level. Some extreme elevations along the Atlantic Coast have been constructed at 3–3.5 m

above mean sea level. Width is also variable but is almost always at least 15–20 m and sometimes will be up to 50 m.

Once the borrow material has been located, all of the permits are obtained, contracts let, and pre-nourishment surveys completed, and then dredging and construction begins. The typical approach is to use a large suction dredge to remove the sand, pump it through pipes or onto barges, and eventually be pumped onto the subject shore. Barges are used for transport when the borrow site is more than a couple of kilometers from the construction site. Large tractors are then used to spread the new beach into the specifications in the construction plans (Figs. 6 and 7).

Such nourishment projects typically cost from about \$3/m³ up to \$15/m³ meaning that many of these projects are over US\$10 million. Overall, this has been a very successful approach. Some areas such as in New Jersey and parts of North Carolina have not been uniformly successful, but others such as the Florida Gulf Coast have been successful beyond predictions (Trembanis and Pilkey 1998). We are still learning about the best way to conduct such protection activities, but the cost/benefit ratio on most of them has been high on the benefit side. Probably, the most successful of these projects, and the largest, is the one at Miami Beach (Fig. 8). It was completed in 1980 at a cost in excess of US\$60 million but is

Human Impact on Coasts,

Fig. 5 Photograph of a detached breakwater designed to protect the shore landward of it. (Courtesy of D. FitzGerald)



Human Impact on Coasts,

Fig. 6 Construction of a new beach using a conveyor belt to offload sand from a large barge



Human Impact on Coasts,
Fig. 7 Pumping and spreading
 nourishment sand in the con-
 struction of a new beach



Human Impact on Coasts,
Fig. 8 Photograph of Miami
 Beach before and after
 construction of the huge beach
 nourishment project. (Photo
 courtesy of US Army, Corps of
 Engineers)



performing very well and is protecting billions of dollars of upland properties.

Dune protection and stabilization. The other environment that is critical to protection of the upland environment and properties is the dunes that are immediately landward of the beach. Destruction of dunes for development purposes has been eliminated along most coasts, and preservation of these dunes is a high priority. In many places dunes have been rebuilt, vegetation has been established, and other measures have been put in place to stimulate dune growth and to preserve those that are there.

Actual construction of dunes is not a widespread practice although small dunes are commonly built to initiate dune development. This approach is most effectively used

after a severe storm has caused removal of all or portions of foredunes or if there have been no dunes along a backbeach area. Most commonly, the approach for stimulating dune growth is through planting of appropriate vegetation. This vegetation provides a very efficient mechanism for trapping windblown sand from the beach, and dunes form very quickly (Fig. 9). These plantings are even irrigated in some areas. Other efforts to enhance dune growth are various types of fencing that will trap sand. Originally, the same type was used that has been used in northern climates for trapping snow along the highways but that has now given way to various types of biodegradable material that will deteriorate when buried for some time. In some countries such as the



Human Impact on Coasts, Fig. 9 Artificially planted vegetation that is designed to trap sand and promote growth of dunes

Netherlands, rows of twigs and shrubbery are planted to trap the windblown sand.

Once established, there are methods of preserving the dunes. The most common and most effective is construction of walkovers to prevent foot erosion from people. This is widespread along coasts where there is considerable development and traffic to the beach. In more remote areas, the paths to the beach are simply developed so as to be in opposition to the prevailing and predominant wind directions. The paths also may have multiple changes in direction to prevent wind from blowing along them and eroding the sand.

Soft shore protection is now the standard due to its compatibility with the natural coastal environment and its esthetically pleasing appearance. Although costs are high and maintenance is mandatory, this approach has prevailed for about two decades and will likely do so in the future.

Indirect Impact

There are numerous ways in by which human activities along the coast can indirectly impact the behavior of coastal environments. These range from the obvious situations where

jetties at an inlet may impact beach erosion kilometers down the beach to more subtle situations where activities in an estuary can influence open coast morphodynamics. The discussion here will focus on the activities that impact on tidal inlet stability. All of these activities are involved, one way or another, with coastal development.

Before proceeding farther in the discussion, it is important to briefly consider the important factors in inlet morphodynamics. First, the volume of water that passes through an inlet during a given tidal cycle is the tidal prism, a water budget. The prism is the product of the area of the back-barrier area served by the inlet and the tidal range. It is the tidal prism that determines the size and stability of the inlet. Large prisms tend to maintain stability in the inlet position and cross-sectional area, whereas small tidal prisms lead to instable inlets that migrate and often close. It is the latter condition that is a main reason for the construction of jetties as considered in an earlier part of the discussion.

There are various types of human activities in the back-barrier/estuarine environments that can impact on the tidal prism and therefore on tidal inlets. In nearly every case, the result of these activities is a decrease in tidal prism and a lack of stability in the tidal inlet(s) involved.

Construction of Fill-Type Causeways

One of the first major activities of barrier island development is the construction of roadways to access these recreational and resort areas. This was typically accomplished by dredge and fill construction of causeways (Fig. 10), many of which were built in the United States in the early 1920s. This was the time that many private vehicles were on the road and people wanted to “drive to the beach.”

Such causeways acted like dams in the area between the mainland and the barrier islands. They became efficient tidal divides because, except for navigational passages, there was essentially no tidal flux across these ribbons of fill. As a consequence, there was significant change in the tidal prism of the tidal inlet(s) that served this back-barrier area. The result was change in the inlet stability, typically a reduction in the size and commonly the tendency for migration. As an example, Dunedin Pass in the Clearwater area of the Florida

Gulf Coast (Lynch-Blosse and Davis 1977) showed a major reduction in its cross-sectional area shortly after construction of the Clearwater Causeway in 1922, and then after construction of the Dunedin Causeway in 1964, it was reduced substantially again (Fig. 11). This same phenomenon has also taken place on the coasts of Texas, New Jersey, and North Carolina.

Another type of human activity that has had a major influence on inlet stability is the dredge and fill activity that has taken place on the landward side of the barrier islands. The typical situation on this side of a barrier island is domination by some type of wetland environment: either salt marsh or, in the low latitudes, mangrove communities. As pressures of development increased and space on barriers became limited, these wetlands were included. Such lands brought only a small price because they were deemed worthless. The potential for development was, however, very good.

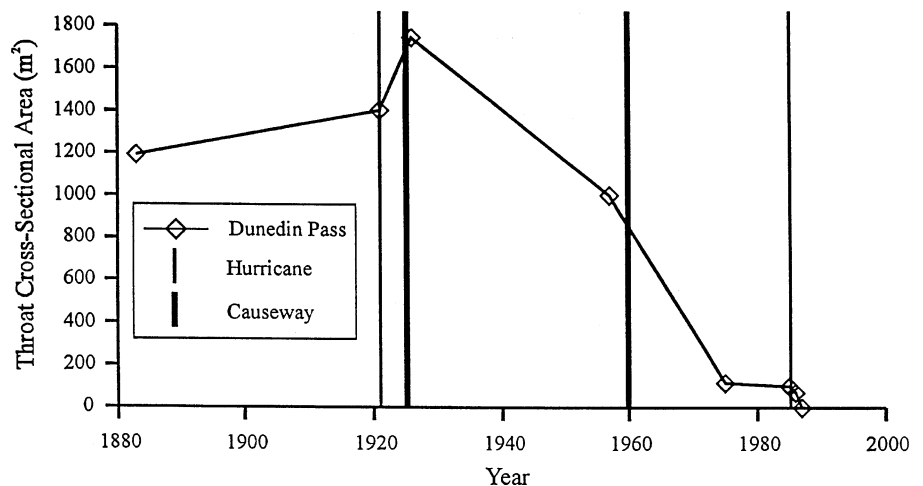
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Fig. 10 A fill-type causeway constructed between the mainland and a barrier island



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Fig. 11 The effects on Dunedin Pass of the construction of two causeways along the central Gulf Coast of the Florida peninsula. (Modified from Lynch-Blosse and Davis 1977)



These wetlands were converted to buildable uplands by dredge and fill techniques. Finger canals were dredged through the wetland, and the spoil was cast to the side, thus producing a supratidal finger of land which, when stabilized by seawalls, was suitable for building houses. These finger canals and their associated small peninsulas of land are widespread along many back barriers but are most common in Florida (Fig. 12).

Obviously such development practices have terrible negative impacts on the wetland environments, essentially destroying all of them. Fortunately, they were stopped in the 1960s. The impact of such activities on the tidal prism is also significant. These practices of dredge and fill construction reduce the area of the back barrier which in turn reduces the tidal prism (Davis and Barnard 2000). A comparison of two examples from the west-central coast of Florida shows the contrast. Caladesi Island is a virtually pristine barrier island that is about the same size and shape as Long Key, a fully developed barrier. The outlines of these barriers display how the area of the back barrier has been reduced (Fig. 13). An

extreme example of this has occurred in Boca Ciega Bay in the St. Petersburg area of Florida (Davis 1989). This large back-barrier estuary has had its surface area reduced by more than 25% since the 1920s.

Construction of the Intracoastal Waterway

The Intracoastal Waterway (ICW) is a dredged and maintained navigational channel that extends from Brownsville, TX, across the Gulf Coast and then continues along the east coast of the United States up to New England. It was originally constructed for commercial traffic to protect vessels, especially barges, from severe weather and energetic wave conditions. Through most of its extent, the ICW has a design width of 50 ft (~15 m) and a depth of 8 ft (~2.5 m). This channel cuts through some land areas but mostly follows along the open water, back-barrier areas. Along most of its extent, the spoil from dredging was cast alongside in piles forming small islands. Aerial photos along many stretches show dozens of these islands that have become valued for fishing and as bird rookeries. The negative side of this dredge

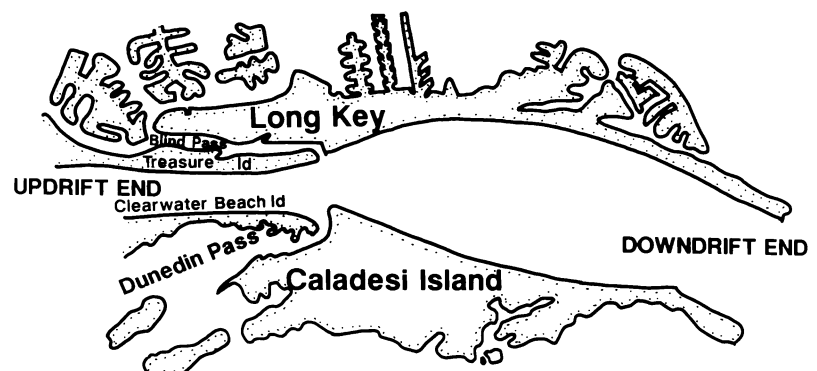
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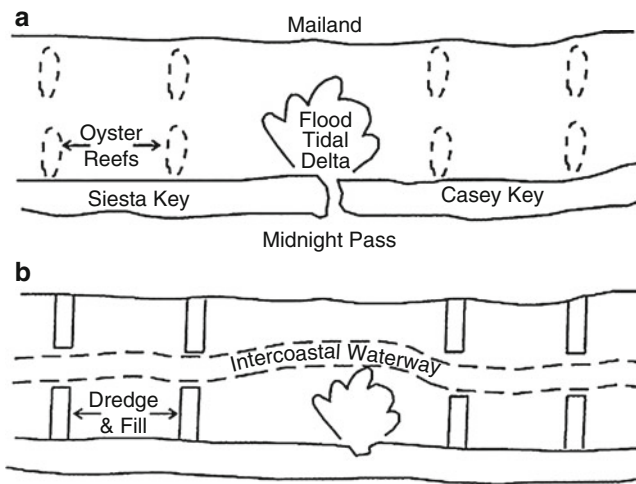
Fig. 12 Finger canal development on the back-barrier area as the back result of dredge and fill construction



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Fig. 13 Comparison of the natural shore of Caladesi Island and the drastically modified shore of Long Key, two originally similar barrier islands on the Gulf Coast of Florida. (From Davis 1989. Reproduced by permission of the ASCE)





Human Impact on Coasts, Fig. 14 Schematic maps of Little Sarasota Bay showing (a) the natural conditions and (b) the results of dredging of the Intracoastal Waterway and dredge and fill of oyster reef areas

spoil is that in many places it was dumped on wetlands and destroyed the area covered. Now a few decades old, the ICW is in need of major maintenance dredging in many locations. Whether or not this happens depends on plans to dispose of the dredge spoil. It is unlikely that disposal of this spoil will be permitted along the channel for environmental reasons. This spoil is not suitable for beach nourishment because of its high mud content and its toxic content at some locations. The cost of disposing it in deepwater offshore is probably too high to be a viable solution. A solution is still not forthcoming.

Another important impact of ICW dredging is its impact on the tidal prism. In narrow back-barrier areas where the prism is comparatively small, such a channel can divert the tidal prism from a natural inlet, along its path up and down the channel. Such a circumstance took place in Little Sarasota Bay, FL in conjunction with the ICW dredging and some other construction practices. The result was closure of a natural tidal inlet that had been open throughout historical time (Davis et al. 1987).

Coincidentally with the dredging of the ICW was the dredge and fill construction over several elongate oyster bars that were covered with mangroves. Dredge material was placed on these features, and sheet piling was used to contain it. These areas now have several houses on each one. The result is that all circulation within Little Sarasota Bay is now channeled through the ICW and the natural inlet that served this bay, Midnight Pass, has been closed due to its greatly reduced tidal prism (Fig. 14).

Summary

Human impacts along the coastal zone are numerous, widespread in kind, and typically detrimental to the environments

where they take place. This discussion has focused on the open coast impacts emphasizing the beach and inlet environments. Other types exist, especially in various coastal bay environments.

In general, the impacts on these environments are the result of development pressures for more space to be occupied by residential or commercial properties. Because the land along the coast is so expensive, the pressures are great, both economical and political. The consequences have been disastrous in nature. Fortunately, most governmental jurisdictions have taken action to prevent such activities in the future, at least in the developed countries. Many of the developing countries are still way behind in their planning and management of the coast. Unfortunately, there are indications that too many of them are not learning from the many mistakes that have already taken place and that are well-documented. The other unfortunate circumstance is that even though we have rules and ordinances in place, there seem to be too many exceptions that are granted to violate these regulations.

Cross-References

- [Beach Nourishment](#)
- [Bioengineered Shore Protection](#)
- [Dredging of Coastal Environments](#)
- [Environmental Quality](#)
- [Estuaries: Anthropogenic Impacts](#)
- [Navigation Structures](#)
- [Shore Protection Structures](#)
- [Tidal Inlets](#)
- [Tidal Prism](#)

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Hydrology of the Coastal Zone

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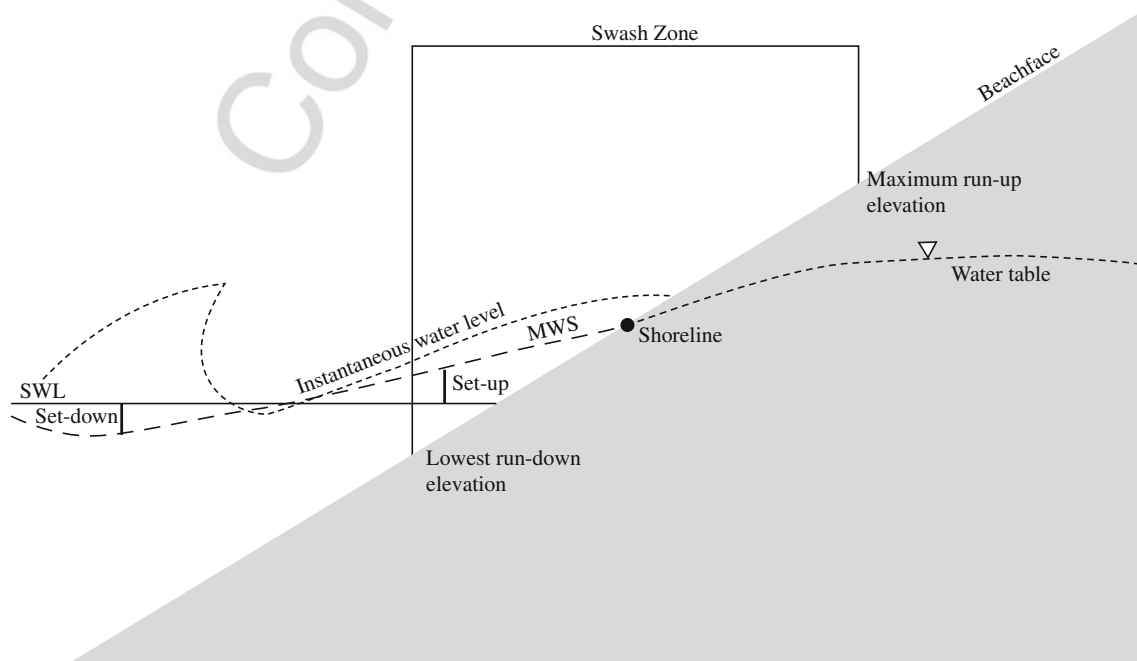
Concepts and Definitions

In the broadest sense, the hydrology of the coastal zone could include the distribution and movement of any water in the coastal zone; however, in practice, this term generally refers to groundwater. A few studies have looked at groundwater movement in coastal barriers, water table fluctuations in estuarine environments and gravel beaches, or sand moisture content effects on aeolian sediment transport. However, most studies of coastal groundwater dynamics have concentrated on groundwater in beaches and, in particular, in the swash zone of sandy beaches. The *swash zone* is the section of the foreshore where final wave energy dissipation occurs, which is alternately covered and exposed by water motions.

The complex interaction of surface and subsurface water in the swash zone is defined below and illustrated in Fig. 1. The *still water level* (SWL) is the water surface in the hypothetical situation of no waves. When the local water-surface elevation is averaged over a time span much longer than incident and infragravity periods but shorter than the tidal period, the result

is the local mean water level, which traces the *mean water surface* (MWS). The MWS in the surf and swash zones generally has a gradient which balances the change in the *radiation stress*, defined as the excess flow of momentum due to the presence of waves. Changes in radiation stress are balanced by changes in hydrostatic pressure, in other words, by changes in water level. This difference is known as *setup* or *setdown*. *Setup* is a wave-induced increase in the MWS, whereas *setdown* is a wave-induced decrease in the MWS. Setdown occurs seaward of the breakers, where radiation stress is at its maximum. The positive gradient due to radiation stress is balanced by a negative water-surface gradient, resulting in a lowering of the MWS to below SWL. Setup occurs inside the surf zone, where the decrease in radiation stress due to energy dissipation is balanced by the raising of the MWS to above SWL. As long as energy dissipation continues, setup continues to increase in the onshore direction and is greatest at the shoreline where the wave setup is of the order 40% of the offshore wave height (Hanslow and Nielsen 1993). The concept of setup is discussed in more detail in the entry on *surf zone processes*.

The *shoreline* is the position where the MWS (including the setup) intersects the beach face, in other words, the line of zero water depth. The shoreline represents the land-water boundary, which moves across the intertidal beach at a range of frequencies from incident waves (3–15 s) to tides (daily or twice daily for high-tide-low-tide cycles and approximately 2 weeks for spring-neap cycles). The shoreline also experiences episodic landward excursions due to the passing of extreme weather events which induce *storm surges* due to

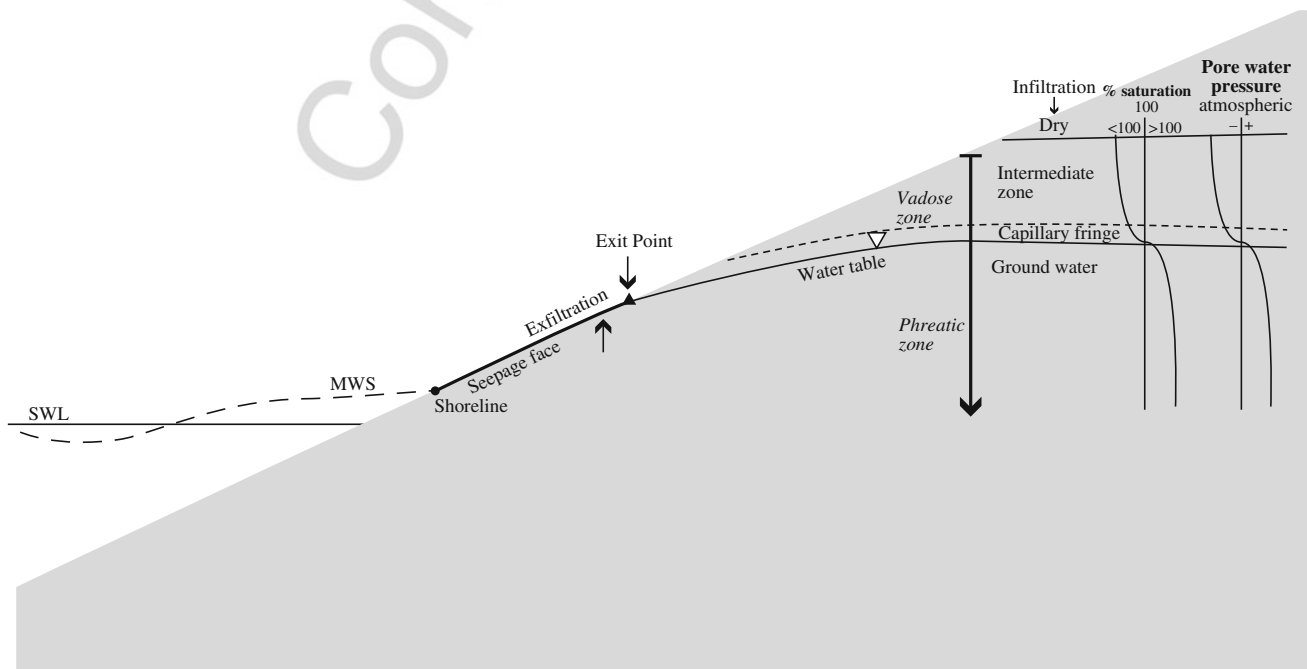


Hydrology of the Coastal Zone, Fig. 1 Definition sketch of surface and subsurface water levels in the swash zone

large storm waves, strong onshore winds, and low atmospheric pressure (see the entry on *Storm Surges* for further details). The limits of shoreline excursion define the boundaries of the swash zone, which migrates up and down the foreshore of the beach over a tidal cycle. The seaward and landward limits of the swash zone are, respectively, the point of collapse of the wave or bore and the landward limit of wave action. There are two components to the water motions in the swash zone. The first is *swash* (also referred to as *uprush*), which is a landward-directed flow characterized by the upslope transport of water. The second component of the cycle is the *backwash*, which is the downslope movement of the water which follows maximum run-up. The uprush-backwash cycle is essentially an oscillation superimposed on the maximum MWS (including setup) inside the surf zone. Total wave run-up represents the combined effect of setup and swash at incident and infragravity frequencies. The *maximum swash height*, or *maximum run-up elevation*, is the maximum vertical height above SWL reached by the uprush. Wave *run-down elevation* is the lowest vertical height reached by the backwash of a wave before the uprush of the next wave begins to run up the beach face. The run-down elevation may be below SWL.

The terminology used by groundwater hydrologists to describe subsurface water may not be familiar to coastal researchers; therefore, these terms are defined in some detail here and are illustrated in Fig. 2. The best general reference on groundwater hydrology is Freeze and Cherry (1979), from which many of the definitions in this entry are taken.

An *aquifer* is a saturated geologic unit that can transmit significant quantities of water under ordinary hydraulic gradients (Freeze and Cherry 1979). An *unconfined aquifer*, or *water table aquifer*, is one that is freely connected with the atmosphere. The *water table* is an equilibrium surface at which pore water pressure is equal to atmospheric pressure (i.e., a gauge pressure of zero). The water table may also be referred to as the *phreatic surface*. *Pore water pressure* is the fluid pressure in the pores of a porous medium relative to atmospheric pressure. Below the water table, pore water pressure is greater than atmospheric pressure; above the water table, pore water pressure is less than atmospheric pressure due to capillary suction. Hydrologists generally use the term *groundwater* to refer to water below the water table, where pore water pressures are positive, and use *soil water* to describe water above the water table where pore water pressures are negative (subatmospheric). However, to equate beach sediment with a soil would be misleading, so in beach hydrology, the term *groundwater* is commonly used to mean any water held in the sand below the beach surface. The *phreatic zone* is the permanently saturated zone beneath the water table. The *vadose zone*, which is sometimes called the *zone of aeration* or the *unsaturated zone*, is the unsaturated zone of moisture that exists above the water table due to capillary suction which may extend to the sand surface near the shoreline. In the saturated (phreatic) zone, pore spaces are filled with water and pore water pressures are equal to or greater than atmospheric pressure. In the zone of aeration, the pores are filled with both water and air and pore water



Hydrology of the Coastal Zone, Fig. 2 Definition sketch of beach groundwater zones when the water table is decoupled from the tide

pressures are less than atmospheric. For this reason, beach groundwater zones are better defined by pore water pressure distribution than by saturation levels. A *capillary fringe* develops immediately above the water table as a result of the force of mutual attraction between water molecules and the molecular attraction between water and the surrounding sand matrix. The capillary fringe may also be referred to as the *tension-saturated zone* (groundwater hydrologists often use the terms *tension* or *suction* – which can be used interchangeably – to describe a pressure which is negative relative to atmospheric pressure). In the capillary fringe, pore spaces are fully saturated, but the capillary fringe is distinguished from the water table by the fact that pore water pressures are negative. The thickness of the capillary fringe in sand beaches is inversely related to the sediment size and may vary between a few millimeters for coarse sediments to over a meter for finer sediments. Some workers also refer to an intermediate zone which may occur above the capillary zone where the degree of saturation may vary but remains less than 100%.

One of the most influential parameters in controlling beach groundwater behavior is the *hydraulic conductivity*, K , which may be defined as the specific discharge per unit hydraulic gradient. The hydraulic conductivity reflects the ease with which a liquid flows and the ease with which a porous medium permits the liquid to pass through it and relates the mean discharge flowing through a porous substance per unit cross section to the total gravitational and potential force. Hydraulic conductivity has units of velocity, usually ms^{-1} in the case of beach groundwater systems. Hydraulic conductivity should be distinguished from *permeability* (also referred to as *intrinsic* or *specific permeability*), denoted by k , which is the measure of the ability of a rock, soil, or porous substance to transmit fluids and refers only to the characteristics of the porous medium and not to the fluid which passes through it. Permeability has dimensions of L^2 .

Porosity quantifies the storage capacity of an aquifer and is defined as the volume of the voids in a sediment or rock divided by the total volume of the sediment or rock. Porosity is denoted by n and is usually reported as a decimal fraction or percent. The *volumetric water content*, θ , is defined as the volume of water in a sediment or rock sample divided by the total volume of the sediment or rock. In saturated conditions, where the pores are filled with water, the volumetric water content, θ , is equal to the porosity, n . In unsaturated conditions, where the pores are only partially filled with water, the volumetric water content, θ , is less than the porosity, n .

Importance of Beach Ground Water and Swash Zone Processes

Swash zone and beach groundwater processes are of interest to geomorphologists who wish to determine beach erosion or

deposition, or aeolian sediment transport, to marine biologists who are interested in intertidal fauna, and to engineers who require data on run-up, particularly on coastal structures such as breakwaters. Over the past few decades, data on the position of the shoreline, which is directly dependent on swash-zone processes, have emerged as one of the principal sources of information for monitoring coastal change. In some cases, the shoreline position (identified as the maximum extent of run-up) is used to establish legal boundaries, setback lines, or flood hazard zones.

Marine biologists have an interest in the swash zone, as the distribution and type of macrofauna inhabiting the intertidal zone of sandy beaches appear to be related to the swash climate. Both the interstitial fauna and the macrofauna of sandy beaches are directly affected by swash and groundwater processes: the former by infiltration, which is responsible for flushing oxygen and organic materials into the sand, and the latter by swash dynamics and the position of the seepage face, which influence tidal migrations and burrowing (McArdle and McLachlan 1991). Differences in the spatial distribution and abundance of beach fauna have been explained in relation to sediment size and beach slope, but have not yet been related successfully to swash dynamics.

Coastal engineers have long recognized the need for a better understanding of swash-zone processes, largely concentrating on the measurement and modeling of run-up on structures such as breakwaters. Such studies are needed to establish design criteria, particularly the elevation of the structure required to prevent overtopping by the run-up of extreme waves. Recently, the commercial possibility of modifying beach water table elevation to control beach erosion has been recognized, and several studies have investigated the use of beach dewatering as an alternative to hard engineering practices. Beach dewatering works by lowering the water table artificially through a system of buried drains and pumps (see the entry on *Beach Drain* for further details). Other engineering applications where knowledge of beach groundwater dynamics is important include water quality management in closed coastal lakes and lagoons and the operation of water supply and sewage waste disposal facilities in coastal dunes, contaminant cycling in estuaries and coastal water resource management issues such as saltwater intrusion into coastal aquifers, wastewater disposal from coastal developments, and pollution control.

An understanding of swash and beach groundwater dynamics is also important in the modeling of beach profile evolution. At present, most beach profile and shoreline change models either do not include, or vastly simplify, sediment transport processes in the swash zone. Cross-shore sediment transport models have demonstrated considerable success in predicting eroding beach profiles on relatively fine sand beaches. Predictions of accretionary events and the behavior of coarser sediment beaches are generally less

good, particularly in the inner surf zone and swash zone (Schoones and Theron 1995). Since an accretionary event is defined by the deposition of sediment above mean sea level, the lack of detailed knowledge of swash and beach groundwater dynamics is probably an important factor in the inability of beach profile models to simulate accretionary events accurately.

Erosion and accretion of the beach profile, and the resulting movement of the position of the shoreline, are a direct result of sediment transport processes occurring in the swash zone and inner surf zone. Beach groundwater-swash dynamics provide an important control on swash-zone sediment transport, which affects the morphology of the intertidal beach by controlling the potential for offshore transport or onshore sediment transport and deposition above the SWL. Cyclic erosion and accretion of the beach face as a result of relative elevations of the beach water table and swash have been substantiated by researchers for many years. Most of these studies suggest that beaches with a low water table tend to accrete and beaches with a high water table tend to erode. Recent observations indicate that flows in the swash zone can also affect the beach profile seaward of the intertidal profile, influencing sediment transport in the bar region.

Swash and beach groundwater interaction may play a particularly important role in profile evolution and sedimentation patterns on macrotidal beaches. Many macrotidal beaches have two, and sometimes three, distinct beach zones: a flat, dissipative low-tide beach and a steeper, more reflective high-tide beach. There is generally an abrupt decrease in beach slope on macrotidal beaches where the water table intersects the beach face, which may be also marked by a change in sediment size between coarse and fine material. The interaction between beach groundwater and swash flows may provide a mechanism for the shore-normal sorting of coarse and fine material that is often observed on macrotidal beaches, with dissipative sand low-tide terraces at the base of steep reflective high-tide gravel ridges. For example, Turner (1993a) surveyed 15 macrotidal beaches on the Queensland coast in Australia and found that a decrease in sediment size was strongly correlated to an increase in the relative extent of the lower gradient (saturated) lower region of the intertidal profile. Turner (1995) developed a simple numerical model that incorporated the interaction of the tide and the beach water table outcrop. This model predicted the development of a break in slope resulting from landward sediment transport and berm development across the alternately saturated and unsaturated upper beach, while the profile lowered and widened across the saturated lower beach. Hughes and Turner (1999) gave different empirical equations for equilibrium slope on unsaturated and saturated beach faces.

Common to all these studies is the observation that when the water table outcrops above the tide, two zones are

distinguished: a lower saturated zone that promotes downslope (offshore) sediment transport and an upper region that alternates between saturated and unsaturated conditions, with upslope (onshore) sediment transport potentially enhanced by infiltration. However, the relative importance of infiltration is not yet known and will be discussed further below in the section on *Mechanisms of Surface-Subsurface Flow Interaction*.

Behavior of Beach Ground Water

Turner and Nielsen (1997) identified a number of mechanisms which have been associated with observed beach water table oscillations: seasonal variations, barometric pressure changes associated with the passage of weather systems and storm events, propagation of shelf waves, and infragravity and incident waves. However, the majority of research has concentrated on tide-induced fluctuations of the beach water table.

A number of studies since the 1940s have described the shape and elevation of the beach water table as a function of beach morphology and tidal state. The majority of these studies have been limited to measurements of water table elevations across the beach profile, although limited observations of both longshore and cross-shore groundwater variations have been reported. The elevation of the beach water table depends on prevailing hydrodynamic conditions such as tidal elevation, wave run-up and rainfall, and characteristics of the beach sediment that determine hydraulic conductivity, such as sediment size, sediment shape, sediment size sorting, and porosity. Observations of beach water table behavior show that the water table surface is generally not flat. Several authors have showed that the slope of the water table changes with the tide, sloping seaward on a falling tide and landward on a rising tide. The slope of the water surface has been found to be steeper on a rising tide than on a falling tide. Other researchers have measured water table elevations with a humped shape, with the hump near the run-up limit. Water table oscillations have been shown to lag behind tidal oscillations. Observed water table elevations are asymmetrical, as the water table rises abruptly and drops off slowly compared to the near-sinusoidal tide which drives it. For a given geometry, the lag in water table response is due mainly to the hydraulic conductivity of the beach sediment. With increasing distance landward, the lag between the water table and the tide increases linearly and the amplitude of the water table oscillations decreases exponentially in accordance with the following equation:

$$\eta(x,t) = Ae^{-k_r x} \cos(\omega t - k_i x) \quad (1)$$

where η is the water table elevation (units of length L), A is the forcing amplitude (L), ω is the angular frequency (T^{-1}), x is

the landward distance, and t is time. $k = k_r + ik_i$ is known as the water table wave number which quantifies how quickly the amplitude decays (k_r) and the time lag increases (k_i). The wave number is a function of the nondimensional aquifer depth $n\omega d/K$ where n is the porosity, d is the aquifer depth, and K is the hydraulic conductivity. For example, as the hydraulic conductivity increases, the nondimensional aquifer depth and hence the wave number decrease resulting in less decay and a faster speed of propagation (i.e., shorter time lag). In addition, as the wave period increases, the angular frequency decreases leading to smaller wave numbers and less decay and time lag. So, groundwater waves in coarse sandy beaches forced by the tide will penetrate further than those in fine sandy beaches forced by wind waves. The presence of a capillary fringe also facilitates greater penetration of high-frequency waves into beaches compared to those with little or no capillary fringe. Shoushtari et al. (2016) provide an evaluation of the ability and limitations of current groundwater flow models to predict the water table wave number.

Wave run-up, tidal variation, and rainfall may produce a *superelevation*, or *overheight*, of the beach water table by raising the elevation of the beach water table above the elevation of the tide.

Many researchers have observed that the beach acts as a filter that only allows the larger or longer period swashes to pass. The frequency of the groundwater spectrum decreases in the landward direction. The further landward the given groundwater spectrum, the narrower its band and the more it is shifted toward lower frequencies as higher frequencies are filtered out by the sand matrix (Cartwright et al. 2006). Infiltration into the beach also acts to reduce the amplitude and frequency of the input swash energy. Based on observations such as these, the beach has often been described as a low-pass filter (meaning that only lower frequency oscillations are transmitted through the beach matrix). High-frequency, small waves are damped, and their effect is limited to the immediate vicinity of the intertidal beach face slope, whereas low-frequency waves can propagate inland. Comparison of run-up and groundwater spectra shows a considerable reduction in dominant energy and also a shift in dominant energy toward lower frequencies.

Decoupling between the tide and the beach water table occurs when the groundwater exit point becomes separated from the shoreline (shown in Fig. 2). This occurs because the rate at which the beach drains is less than the rate at which the tide falls, so the tidal elevation generally drops more rapidly than the water table elevation and decoupling occurs, with the water table elevation higher than the tidal elevation. The *exit point* is the position on the beach profile where the decoupled water table intersects the beach face. After decoupling occurs, the position of the exit point is independent of the MWS until it is overtopped by the rising tide. Below the exit point, a *seepage face* develops where the water table coincides with

the beach face. The seepage face is distinguished by a glassy surface. The seepage face is different from the water table in that its shape is determined by beach topography. However, water on the seepage face is at atmospheric pressure, as is water at the water table. The extent of the seepage face depends on the tidal regime, the hydraulic properties of the beach sediment, and the geometry of the beach face; thus, the degree of asymmetry in water table response will vary between beaches. The exit point is generally assumed to mark the boundary between a lower section of the beach which is saturated and an upper section which is unsaturated; however, this assumption is probably an oversimplification.

Modeling Beach Ground Water Dynamics

Beach groundwater systems are generally treated as unconfined aquifers because commonly the upper boundary to groundwater flow is defined by the water table itself rather than by some surface layer of impermeable material (Masselink and Turner 1999). The beach groundwater system is underlain by an impermeable boundary at a depth which is often unknown. The rate of flow (or *specific discharge*) of water through unconfined aquifers, q , is given by Darcy's law:

$$q = -K\nabla h = -K\left(\frac{\partial h}{\partial x} + \frac{\partial h}{\partial y} + \frac{\partial h}{\partial z}\right) \quad (2)$$

where h is the hydraulic head (units of length, L), K is the hydraulic conductivity (LT^{-1}), ∇ is the gradient operator, and x , y , and z are the Cartesian coordinates.

Darcy's law is valid in both the saturated and unsaturated zones as long as flow is laminar, which is a reasonable assumption for sand beaches. Darcy's law has also been successfully applied to simulate the flow in gravel beaches (e.g., Guo et al. 2010). See the entry on *Gravel Beaches* for further details. Darcy's law shows that the rate of groundwater flow is proportional to the hydraulic gradient or slope of the water table. The *hydraulic gradient* (∇h) is the change in hydraulic head (h) over distance. Water flows down the hydraulic gradient in the direction of decreasing head. The *hydraulic head* (h) is the sum of the elevation head (z) and the pressure head (ψ) and is measured in length units above a datum. There is no standard datum that is used in beach hydrology, but many researchers use the elevation of an impermeable layer beneath the beach sediment, so that the vertical coordinate z is measured from the impermeable base. Some workers have considered the hydraulic head in a beach groundwater system to be the elevation of the free water surface or water table elevation. However, this is only true when there is no vertical component to the flow, in other words, when Dupuit-Forchheimer conditions apply (see below).

Groundwater hydrologists generally model water flow using Darcy's law in combination with an equation of continuity that describes the conservation of fluid mass during flow through a porous medium. Historically, modeling beach groundwater flow in response to tidal forcing in sandy beaches used the one-dimensional (1-D) form of the Boussinesq equation:

$$\frac{\partial h}{\partial t} = \frac{K}{s} \frac{\partial}{\partial x} \left(h \frac{\partial h}{\partial x} \right) \quad (3)$$

where h is the elevation of the water table (L), t is time (T), K is hydraulic conductivity (LT^{-1}), s is the specific yield (dimensionless), and x is horizontal distance (L). The *specific yield* is defined as the volume of water that an unconfined aquifer releases from storage per unit surface area of aquifer per unit decline in water table (Freeze and Cherry 1979). While this equation performs reasonably well at modeling tide-induced groundwater waves, it ignores the unsaturated zone above the water table and the effects of vertical flows (non-hydrostatic pressure) which become important in non-shallow aquifers and in the case of higher-frequency waves.

Workers have investigated effects of the unsaturated zone on beach groundwater behavior by means of either correction terms added to the Boussinesq equation based on the Green-Ampt capillary fringe parameterization (e.g., Li et al. 1997b) or solving *Richards'* equation which derived using Darcy's law and the conservation of mass principle across both the saturated and unsaturated zone:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left(K(\psi) \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial z} \left(K(\psi) \frac{\partial h}{\partial z} \right) \quad (4)$$

where θ is the volumetric soil water content (unitless), h is the hydraulic head (L), ψ is the pressure head (L), z is the elevation head (L), and $K(\psi)$ is the hydraulic conductivity (LT^{-1}). Despite these advancements, none of the above models are capable of predicting observed decay and time lags of high-frequency waves (Shoushtari et al. 2016), and as such this remains an area of active research.

Models of beach water table fluctuations that incorporate wave effects have been developed only relatively recently. Nielsen et al. (1988) proposed the use of a linearized version of the Boussinesq equation (Eq. 2) with an additional term to account for the additional, time-averaged mass added to the system due to infiltration of wave run-up:

$$\frac{\partial h}{\partial t} = \frac{Kd_a}{s} \frac{\partial^2 h}{\partial x^2} + U_1(x, t) \quad (5)$$

where d is the aquifer depth and $U_1(x, t)$ is the time-averaged infiltration/exfiltration velocity per unit area, h is the elevation of the water table (L), t is time (T), K is the hydraulic

conductivity (LT^{-1}), s is the specific yield or drainable porosity (dimensionless), and x is horizontal distance (L). The 1-D nature of this particular model has found it to be inadequate in modeling infiltration due to wave run-up, particularly near the run-up limit (Cartwright et al. 2002). Li et al. (1997b) and Li and Barry (2000) have developed more complicated models to predict wave-induced water table fluctuations; however, none of the models which include wave effects have yet been tested against field or laboratory data.

More recently, more concerted efforts have been made to link swash hydrodynamic and sediment transport models. More information can be found in the next section and also in the reviews of Elfrink and Baldock (2002) and Bakhtyar et al. (2009).

Mechanisms of Surface-Subsurface Flow Interaction and Implications for Sediment Transport

Several mechanisms have been suggested to explain why beaches with a low water table tend to accrete and beaches with a high water table tend to erode. The mechanisms which are proposed most frequently are infiltration and exfiltration. The terminology used to discuss these mechanisms requires some clarification, as different terms may be used by hydrologists, engineers, and other coastal scientists. The physical process of interest is that of vertical flow within a porous bed and/or through a permeable boundary. Vertical flow exerts a force within the bed called *seepage force*, which is defined as a force acting on an individual grain in a porous medium under flow, which is due to the difference in hydraulic head between the front and back faces of the grain (Freeze and Cherry 1979). The seepage force, F , is exerted in the direction of flow and is directly proportional to the hydraulic gradient and is given by

$$F = \rho g \frac{\partial h}{\partial z} \quad (6)$$

where ρ is the density of the fluid (ML^{-3}), g is acceleration due to gravity (LT^{-2}), and $\partial h/\partial z$ is the hydraulic gradient (dimensionless). In the convention used here, a positive hydraulic gradient represents a downward-acting seepage force and a negative hydraulic gradient represents an upward-acting seepage force.

The vertical flows which produce this seepage force have been referred to in a number of different ways in the literature: bed ventilation, suction and blowing, piping, seepage erosion, groundwater sapping, etc. In the case of beach hydrology, however, the terms infiltration and exfiltration are most commonly used. *Infiltration* is the process by which water enters into the surface horizon of a soil or porous medium, such as

beach sediment, in a downward direction from the surface by means of pores or small openings. Infiltration is often used interchangeably with *percolation*, which more correctly refers to the flow of water through a soil or porous medium below the surface. Recently, the term *exfiltration* has been used to describe outflow from the bed. Infiltration/exfiltration velocity may also be referred to as seepage velocity.

Grant (1946, 1948) was among the first to suggest a link between beach groundwater behavior and swash-zone sediment transport, proposing a simple conceptual model which has been highly influential in beach hydrology research. Grant defined a dry foreshore as one with a low water table and an extensive infiltration zone. On a dry foreshore, most of the water infiltrates rapidly into the sand above the water table, which reduces the flow depth of the swash and thus the velocity, allowing sediment deposition. Grant's conceptual model also described conditions on a wet foreshore, one whose water table is high and contiguous with the surface of most of the foreshore. He reasoned that when the beach is in a saturated condition throughout all of the foreshore, the backwash, instead of being reduced by infiltration, retains its depth and is augmented by the addition of water rising to the surface of what he called the effluent zone (the seepage face). Grant also noted that groundwater outcropping at the beach surface can cause dilation or fluidization of the sand grains, allowing them to be entrained more easily by backwash flows.

Although most researchers have concentrated on infiltration/exfiltration and possible effects on swash/backwash asymmetry, Nielsen (1992) suggested that vertical flow within the beach alters the sediment transport characteristics due to a modification of the effective weight of the sediment, which will act to stabilize the bed under infiltration or destabilize under exfiltration. Turner and Nielsen (1997) identified several mechanisms by which vertical flow through a porous bed could affect swash-zone sediment transport, including an alteration in the effective weight of the surface sediment due to vertical fluid drag and modified shear stresses exerted on the bed due to boundary layer thinning due to infiltration or thickening due to exfiltration. Experimental work on the influence of seepage flows within sediment beds provides conflicting results concerning the effect on bed stability likely due to the fact that the effects of seepage force and boundary layer thinning tend to oppose each other. While infiltration results in a stabilizing seepage force due to an increase in the effective weight of the sediment, simultaneous boundary layer thinning has the opposing effect of enhancing sediment mobility due to an increased shear stress and vice versa for exfiltration (e.g., Hughes and Turner 1999). The relative importance of these opposing effects depends on the density of the sediment and the permeability of the bed (Nielsen et al. 2001). Nielsen (1997) proposed the following revised Shields parameter (which quantifies the extent of sediment mobility) that includes the effects of infiltration/exfiltration:

$$\theta_m = \frac{u_{*0}^2(1 - \alpha(w/u_{*0}^2))}{gd_{50}(s - 1 - \beta(w/K))} \quad (7)$$

where w is the seepage velocity (LT^{-1} , with infiltration negative), u_{*0}^2 is the shear velocity without seepage (LT^{-1}), s is relative density (dimensionless: ρ_s/ρ , where ρ_s is the density of the sediment and ρ is the density of the fluid), K is hydraulic conductivity (LT^{-1}), g is acceleration due to gravity (LT^{-2}), d_{50} is median grain diameter, and α and β are constants, defined by Nielsen et al. (2001) as 16 and 0.4, respectively. The factor β is intended to quantify the increase of the particle's weight due to the vertical seepage velocity and is 1 for particles in the bed but considerably smaller for particles on the surface (Nielsen et al. 2001).

The extra term in the numerator represents the increase in shear stress due to the thinning of the boundary layer, and the extra term in the denominator represents the effect of the downward seepage drag on the effective weight of the grains (Nielsen et al. 2001). Equation 7 suggests that for a fixed sediment density, as grain size (and therefore, hydraulic conductivity) decreases, the stabilizing effect will increase. Therefore, finer quartz sands ($d_{50} < 0.58$ mm) are likely to be stabilized by infiltration, whereas the net effect of infiltration on beaches of coarser sediment may be destabilizing (Nielsen 1997). Nielsen et al. (2001) extended this analysis to show that infiltration is likely to enhance sediment mobility for dense, coarse sediment where $\alpha(s-1) > \beta(u_{*0}^2/K)$ and impede sediment motion for light, fine K sediment where $\alpha(s-1) < \beta(u_{*0}^2/K)$.

Turner and Masselink (1998) also followed this approach but included the effects of the seepage flow on the bed shear stress. They used their modified Shields parameter, which incorporated an additional through-bed term, to calculate the swash-zone transport rate in the presence of infiltration/exfiltration relative to the case of no vertical flow through the bed. Their modeling showed that altered bed stresses dominated during uprush, indicating enhanced sediment mobility relative to the case of an impermeable bed. They found that altered bed stress effects were also dominant during backwash; however, the net effect of combined seepage force and altered bed stress was less pronounced during backwash than during uprush. Turner and Masselink (1998) concluded that the effects of combined seepage force and altered bed stress enhanced net onshore sediment transport on a saturated beach face. The field experiments of Butt et al. (2001) indicate that the influence of in-exfiltration on the swash boundary layer can be neglected for fine and medium sands with a diameter of $d < 580$ μ m. This is consistent with the laboratory experiments of Nielsen et al. (2001) which showed that infiltration through a horizontal sand bed under regular nonbreaking waves under conditions of steady downward seepage had the effect of reducing the mobility of

0.2 mm sand, and they suggested that infiltration effects on sediment mobility in the swash zone would be minor if infiltration rates are in the range reported by other researchers, where $w < 0.15 K$.

The potential of beach groundwater fluctuations to cause bed failure due to instantaneous fluidization has also been considered. *Fluidization* of sediment occurs when the upward-acting seepage force exceeds the downward-acting immersed particle weight (the stabilizing force). In particular, it has been suggested by a number of workers that tidally induced groundwater outflow from a beach during the ebb tide may enhance the potential for fluidization of sand and thus the ease with which sand can be transported by swash flows. This has been suggested as the mechanism by which there is a break in the beach slope (flatter closer to the low-tide shore line than the upper beach) at many macrotidal beaches. However, tidally induced groundwater outflow alone is unlikely to be sufficient to induce fluidization, because hydraulic gradients under the sand surface will tend to be relatively small, generally of the order of the beach slope (1: 100-1: 10). In addition, Turner and Nielsen (1997) found that, rather than fast water table rise in the swash being the cause of upward flow (and hence potential fluidization), rapid water table rise within the swash zone resulted from a small amount of infiltration of the swash lens since the tension-saturated capillary fringe extends from the water table to the sand surface for much of the beach profile. However, upward-acting swash-induced hydraulic gradients which are capable of fluidizing the bed have been measured within the top few centimeters of the beach. Horn et al. (1998) presented field measurements of large upward-acting hydraulic gradients which considerably exceeded the fluidization criterion, which occurs when the upward-acting hydraulic gradient is greater than (i.e., more negative than) about -0.6 to -0.7 (in the convention used here). The mechanism responsible for these upward-acting hydraulic gradients is not clear but may be related to the time-averaged wave-induced groundwater circulation pattern first reported by Longuet-Higgins (1983).

Baird et al. (1996) argued that fluidization is only generally possible in the presence of swash on a seepage face. As a swash flow advances over the saturated beach surface, there will be a rapid increase in pore water pressures below the beach surface. When under swash flow, the beach sediment behaves like a confined aquifer. The sediment is saturated, and movement of water into the beach is extremely limited since changes in porosity due to expansion and contraction of the mineral "skeleton" will be minimal. However, water pressures will propagate rapidly through the sediment. As the swash retreats, there will be a release of pressure on the beach face, potentially giving large hydraulic gradients acting vertically upward immediately below the surface. The resultant seepage force associated with these upward-acting

hydraulic gradients could be sufficient to induce fluidization of the sand grains at the surface. They showed theoretically how hydraulic gradients in the saturated sediment beneath swash can exceed, or at least come close to, the threshold for fluidization.

Baldock et al. (2001) compared field measurements of swash-induced hydraulic gradients in the surface layers of a sand beach to the predictions of a simple (1-D) diffusion model based on Darcy's law and the continuity equation. The model allows for dynamic storage (within the sediment-fluid matrix) due to loading/unloading on the upper sediment boundary. The model predicted minimal hydraulic gradients for a rigid, near fully saturated sediment which were in accordance with measurements close to the seaward limit of the swash zone. The model also provided a good description of the measured hydraulic gradients, both very close to the surface and deeper in the bed, for the region of the beach where the beach surface is frequently exposed between swash events. These model-data comparisons suggest that the surface layers of a sand beach store and release water under the action of swash, leading to the generation of relatively large hydraulic gradients as suggested by Baird et al. (1996). However, the model was not able to predict the very large near-surface negative hydraulic gradients observed by Horn et al. (1998), although, for the same swash events, the agreement was good deeper in the bed. Baldock et al. (2001) concluded that the very large upward-acting hydraulic gradients observed in the upper part of the bed were not simply due to pressure propagation during swash loading/unloading or swash-generated 2-D subsurface flow cells. Instead, they suggested that these very large negative hydraulic gradients are probably generated by alternative mechanisms, possibly due to non-hydrostatic pressures developing within the sheet flow layer that occurs during backwash.

The implications of these hydraulic gradients for sediment transport are not clear. Vertical seepage forces are not themselves capable of transporting sediment; however, this process may act to provide readily entrainable material which is then available for transport, onshore under uprush or offshore under backwash. Nielsen et al. (2001) noted that their experiments indicated only the effect of infiltration/exfiltration on sediment mobility and did not necessarily suggest anything about the direction of net sediment transport. This is likely to be affected by other factors such as the phase relationship between infiltration- and exfiltration-induced effects on sediment transport and swash flows. For example, Blewett et al. (1999) measured events where large upward-acting hydraulic gradients occurred when the head of water at the surface, and therefore the uprush or backwash flow, was zero. Under these conditions, even if the sediment were to be fluidized, it would not be transported. However, in other data sets, Blewett et al. (1999)

reported measurements with upward-acting hydraulic gradients of -1.7 , which were more than sufficient to fluidize the bed. These hydraulic gradients lasted for approximately 4 s in waves with a period of 6.3 s under a falling head of water, initially as deep as 40 mm, and under offshore directed flows of $0.7\text{--}1.4\text{ ms}^{-1}$. This suggests a possible erosional mechanism under backwash. Clearly, the phasing between these potentially destabilizing hydraulic gradients and swash flows is critical to the potential for sediment transport.

Nielsen et al. (2001) argued that if the beach face tends to be fluidized during backwash as suggested by Horn et al. (1998), a mechanism must exist to enhance sediment transport during the uprush in order to balance this effect – otherwise the beach would rapidly disappear. They suggested that this balancing effect might be delivered by fluidization due to strong horizontal pressure gradients near bore fronts. However, Hughes et al. (1997) suggested an alternative mechanism, arguing that onshore transport in the uprush is likely to be significantly influenced by turbulence and sediment advection from bores arriving at the beach face. The lack of a clear mechanism for onshore transport highlights the complexity of sediment transport processes in the swash zone, as the exact nature of the relationship between swash flows, beach groundwater, and cross-shore sediment transport is not yet known.

Some of the more recent research has focused on the variability in the swash-aquifer coupling as a function of position on the beach profile. In large-scale laboratory experiments, Tuner et al. (2016) found that the dominant influence on the groundwater level, flow paths, and fluxes within the beach face region of the sand barrier was wave action with the barrier-scale hydraulic gradients having relatively little influence. Sous et al. (2013) showed a strong dependence of the relative weight modification in the Shields parameter to position on the beach face with positive increases up to 4% on the beach face and up to 8% at the berm. Near the toe of the beach however the probability density function tended to be centered on 0%. With increased computational power, beach face sediment transport and morphological change have been examined using coupled process-based numerical models. For example, Bakhtyar et al. (2011) coupled a swash-zone model based on the Navier-Stokes equations to a groundwater flow model to obtain infiltration/exfiltration rates to feed into a modified Shields parameter similar to Eq. 7. The model results revealed that the maximum infiltration rate occurred in the upper beach resulting in the generation of a berm above the SWL and erosion below the SWL. The model also confirmed Grant's conceptual model with the largest berm forming when the water table was low. The model results also demonstrated that the swash-groundwater coupling and resultant berm formation were greater for coarse sediments than fine.

Saltwater Intrusion and the Geochemistry of Beach Groundwater Systems: The Subterranean Estuary Concept

Beach aquifers are at the interface between the ocean and terrestrial groundwater systems and consequently play a critical role in the mixing of the two water masses. In this context, the beach groundwater system is commonly referred to as being a *subterranean estuary*. First proposed by Moore (1999), a subterranean estuary is a reactive mixing zone where salty, oxygen rich ocean water mixes with fresh, often oxygen depleted, terrestrial groundwater. For example, chemical reactions between the water and sediment particles will modify the chemical composition of the water and the mixing of salt and fresh water is important for carbonate diagenesis. The mixing of dissolved oxygen between the two water bodies also plays a role in the discharge of nutrients to the coastal ocean. Further information can be found under *Submarine Groundwater Discharge*.

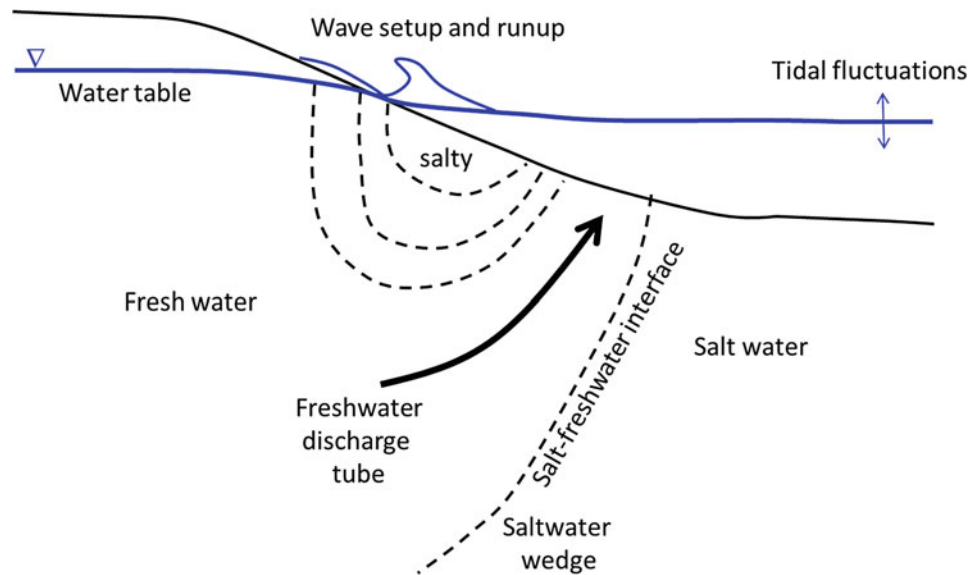
Saltwater intrusion into coastal groundwater systems is an important issue worldwide as it can impact on the availability of potable groundwater resources. As shown in Fig. 3, salty ocean water (higher density) creates a saltwater wedge in underneath the fresh terrestrial groundwater. This is then further complicated by oceanic tides which cause the shoreline to move laterally across the beach and, on the rising tide, add salt water on top of the fresh water which discharges to the ocean via a freshwater discharge tube (Robinson et al., 2006). Infiltration of salty ocean water due to the action of wave run-up adds even more salt to the system (Xin et al. 2010).

Santos et al. (2012) provide a review of the driving forces behind the transport of solutes in permeable coastal sediments and identified 12 such drivers, namely:

1. Terrestrial hydraulic gradients
2. Seasonal changes in the aquifer level on land moving the location of the subterranean estuary
3. Wave setup and tidal pumping
4. Water level differences across permeable barriers
5. Flow- and topography-induced pressure gradients
6. Wave pumping,
7. Ripple and other bedform migrations
8. Fluid shear
9. Density-driven convection
10. Bioirrigation and bioturbation
11. Gas bubble upwelling
12. Sediment compaction

They note that these drivers are often a source of new or recycled nutrients to the coastal ocean and also transform organic carbon into inorganic carbon. They estimated that, collectively, these drivers act to filter a volume equivalent to that of the entire ocean over a time scale of about 3000 years.

Hydrology of the Coastal Zone, Fig.3 Illustration of a subterranean estuary and the influence of oceanic forcing on the salinity distribution in beaches



Summary

The hydrological processes occurring in the coastal zone are the result of a highly dynamic and complex interaction between the coastal ocean and terrestrial water bodies. Forcing of the system occurs over a wide range of frequencies and magnitudes including seasonal to inter-decadal variations in rainfall and evaporation, high-intensity storm surge events over the time scale of hours to days, regular forcing from astronomical tides ranging from 12.25 hourly semidiurnal tides to the 14-day neap to spring cycle, and forcing due to wind waves running up a beach with a period of the order 10 s. This complex ocean-aquifer interaction has a range of implications for both the mobility of sediments on beaches (and hence coastal erosion and recovery) and also for the fate of contaminants and salty ocean water intruding into fresh terrestrial water. Many challenges still exist to better our understanding of this system so that we can better manage it in a long-term, sustainable way.

Cross-References

- Beach Drainage
- Coastal Wells
- Cross-Shore Sediment Transport
- Depth of Disturbance
- Storm Surge
- Submarine Groundwater Discharge
- Surf Zone Processes
- Tides
- Waves

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Ice-Bordered Coasts

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Introduction

Since the publication of *The Encyclopedia of Beaches and Coastal Environments* (Schwartz 1982), the term “cold coasts” has come into common use even serving as a chapter title in the book *Coastal Problems*. The authors, Viles and Spencer (1995), use the 1961 definition by R.L. Nichols that cold coasts “those where there is or has been abundant sea ice, lake ice, water-terminating glaciers or deeply frozen ground” (p. 254). The advantage of such a definition is that it avoids the latitudinal restriction placed by such locational designators as Arctic and Antarctic and thus can accommodate lower latitudinal examples including the tidewater glaciers of Chile and southern Alaska and the presence of sea ice along the Labrador and Hokkaido coasts or even the coast of Spain during the Pleistocene.

This entry treats two of these types of cold coasts: namely, water-terminating glaciers and sea ice. Glacial ice is land-derived and tends to be perennial; sea ice, on the other hand, develops seaward of the coastline and is usually seasonal. They both can serve as erosional, transportational, and depositional agents along coastlines, although the rates and intensities of their action vary greatly between the two and with time and location.

Glacial Ice and the Coast

One of the conspicuous features of the landscape at present and during much of the past 3 million years is glacial ice.

Although today it is dominant (as a coastal feature) only in Antarctica, Greenland, and a few smaller islands in high latitudes, it is still sufficiently abundant to affect more than 35,000 km of the world’s coastline. Glacial ice impacts the coastline in several ways including burying and depressing the coast as it moves across it into the sea, flowing in pre-existing valleys into the sea as tidewater glaciers, and replacing the traditional shore with an ice margin that then interfaces directly with the sea.

Ice sheets, such as those in Antarctica and Greenland, serve as the source of ice that moves under its own weight and in response to gravity from inland to sea. Glacial ice is so extensive in Antarctica that only about 5% of the coastline is ice-free (Fig. 1). Greenland contrasts with most of Antarctica in that glacial ice does not dominate the coastal zone but flows through relatively few high passes in the coastal mountains (Fig. 2). A major exception is where inland ice reaches Melville Bay along a 460 km front.

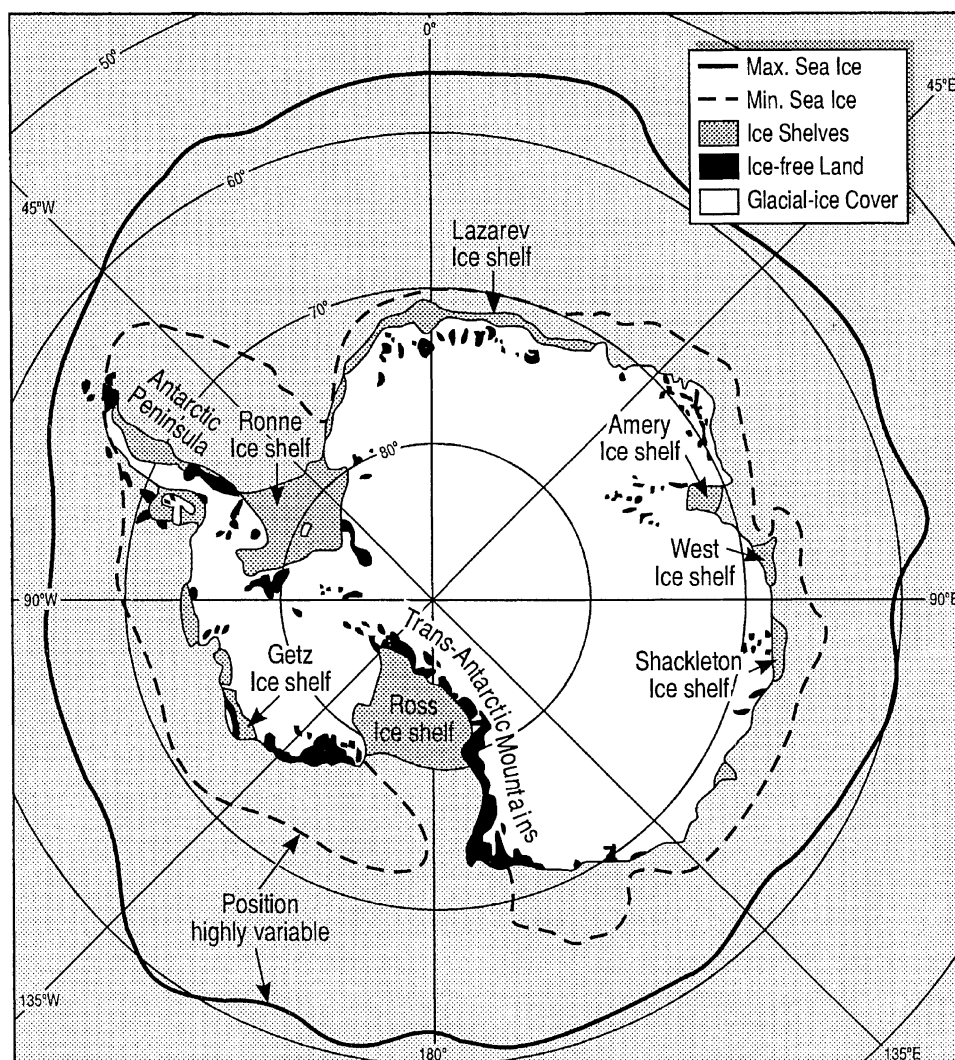
A further characteristic of ice sheets is that their great weight depresses the landmasses upon which they form. Near the coastline the depressed continental shelf slopes downward inland thereby reversing the general slope of the coastal zone. Large parts of Antarctica and Greenland have been depressed to such extents that their present elevation is below sea level. Although, with rebound, much of this depressed land would be above sea level, large areas beneath the West Antarctic Ice Sheet (which averages 440 m below sea level) would still be a series of islands with lengthy coastlines even after the rebound that would accompany deglaciation (Paterson 1994).

Ice shelves, which are large ice masses floating on the sea, range in thickness up to several hundred meters. They are especially common in Antarctica (Fig. 1), but also occur as small shelves in Greenland, Ellesmere Island, and Franz Josef Land. The ice shelves of Antarctica fill many of the continent’s embayments giving the continent a nearly circular form excepting the northward trending Antarctic Peninsula.

Thus, the 11,000 km (more than one-third of the total) of Antarctica’s coastline occupied by ice shelves, possesses

H. Jesse Walker: deceased.

Ice-Bordered Coasts, Fig. 1 Ice and the Antarctic coastline. (Modified from Mellor 1964; Paterson 1994; Crossley 1995)



ice cliffs (Fig. 3) at the ocean interface. These ice cliffs are impressive coastal features as illustrated by the Ross Ice Shelf (Fig. 1) where the front edge is more than 200 m thick with a 20–30 m cliff above water. The ice landward of the Ross ice cliff thickens to 700 m at land's edge (Robe 1980). The exposed ends of these floating glaciers are impacted by the same agents and processes as the coast proper including waves, tides, currents, and sea ice.

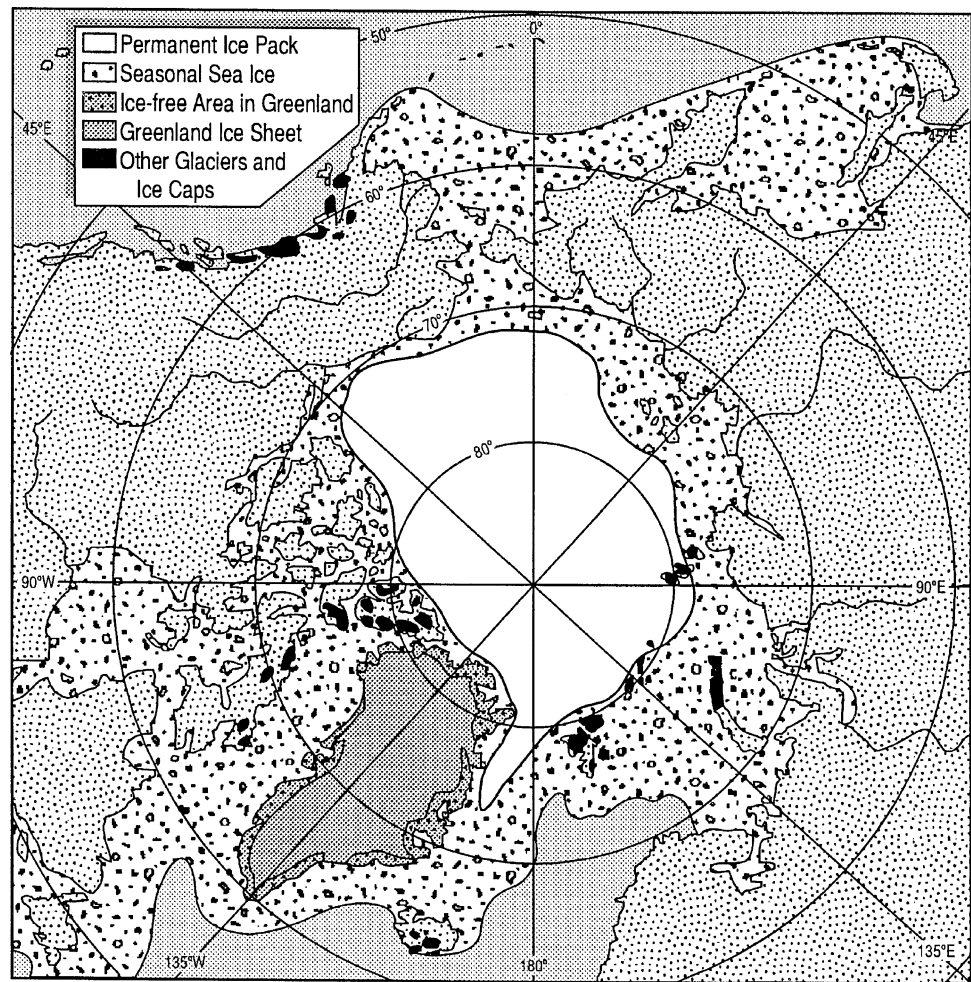
Many of the coastal glaciers in Antarctica (~38%) and most in the Northern Hemisphere do not have floating shelves or tongues but are grounded at their termini (Powell and Domack 1995). For example, part of the west coast of the Antarctic Peninsula has glaciers that have formed ice walls at the shore because of undercutting and calving at the ice front where they override a gravelly beach (Robin 1979).

Tidewater glaciers are found in nonpolar areas such as southern Alaska and Chile as well as at higher latitudes. In the Arctic, tidewater glaciers are present in Jan Mayen, Svalbard, Novaya Zemlya, Severnaya Zemlya, Ellesmere,

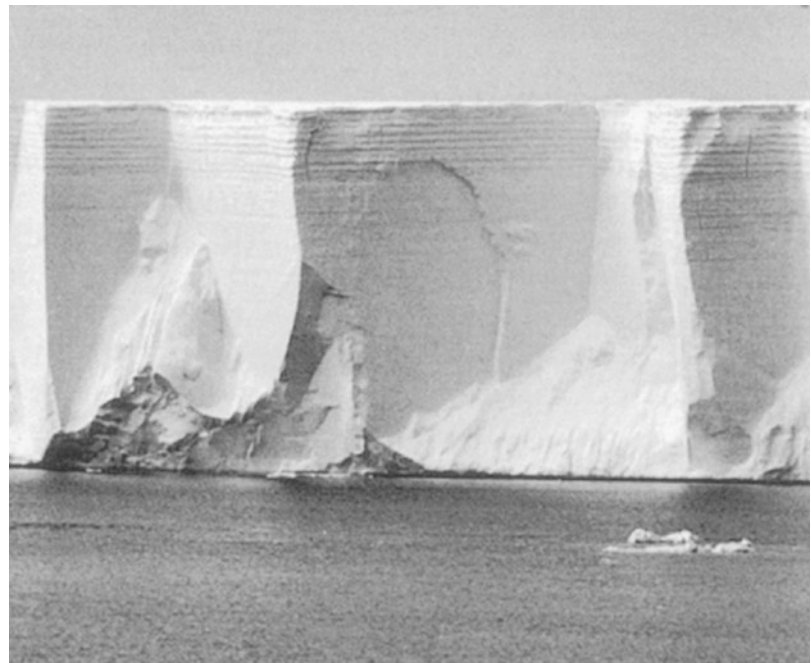
Baffin Island, Bylot Island, and Devon Island as well as Greenland and Alaska (Fig. 2). The termini of temperate tidewater glaciers are grounded as are many in higher latitudes (Hambrey and Alean 1992). Most tidewater glaciers are confined in fjords of variable length and terminate at some distance inland from their mouth. Many are very long, such as the combined 350 km long Nordvestfjord and Scoresby Sund in east Greenland. Such fjords not only have glaciers at their inner ends but also are icebound with icebergs and sea ice. Because of the nature of the coastal mountain rim around the Greenland Ice Sheet the coastal impact is highly varied at the front of outlet glaciers. Although many flow into the sea, others terminate on land (Fig. 4) and in glacially created lakes (Warren 1991).

A major characteristic of glaciers is their changeable rate of advance and retreat. Thus, their position and therefore their impact on the coastlines they border is continuously varying. Jakobshavn Isbrae (Fig. 4) in west Greenland, for example, retreated 26 km between 1851 and 1953 (Williams

Ice-Bordered Coasts, Fig. 2 Ice and the Northern Hemisphere Coastline. Note that within the ice-free coastal area of Greenland there are numerous small ice caps. (Modified from Mellor 1964; Field 1975; Williams and Ferrigno 1995)

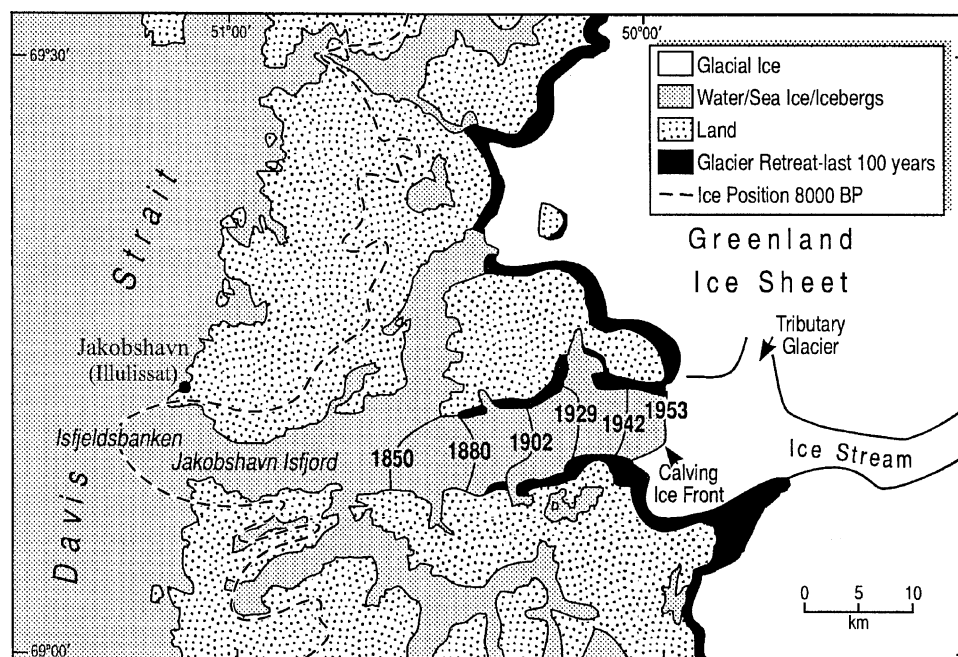


Ice-Bordered Coasts, Fig. 3 Ice cliff, Antarctica. (Williams and Ferrigno 1995. Photograph by Charles Swithinbank, courtesy of Richard S. Williams, United States Geological Survey)



Ice-Bordered Coasts,

Fig. 4 Tidewater glacial retreat (1850–1953) and ice-margin retreat in Jakobshavn Isfjord. (Modified from Williams and Ferrigno 1995)



and Ferrigno 1995) exposing sizable sections of the fjord's shore that it formally bordered.

The most important mechanism for the loss of glacial ice, whether it be the large ice shelves of Antarctica or the smaller tidewater glaciers of Alaska, is calving at their termini. Calving produces icebergs that range in size from very small, as those produced in constricted fjords, to very large such as those tabular icebergs that calve from ice shelves. On March 17, 2000, iceberg B-15, with a length of 295 km and a width of 37 km calved from the Ross Ice Shelf. It broke into two parts a few months later (Lazzara et al. 1999).

Icebergs, once formed, essentially become floating islands. The act of calving increases the number of sides in contact with the sea, hastening their disintegration. Many of them float for years as exemplified by those ice islands that have been used as research bases in the Arctic Ocean. Others become trapped, even if temporarily, in sea ice and some become grounded in shallow water. They can also be erosional agents, often in association with the floating pack ice that forms around them, creating deep gouges in the nearshore bottom.

Glaciers, whether they terminate onshore, at the shore, along fjords or at some distance offshore, are major morphological agents. If terminating inland, meltwater drainage carries the sediment formed by the glacial scour that accompanies the advancing ice to the sea creating depositional facies along the shore. Those glaciers that terminate at the shore and those that have overridden the shore leave behind ice-scoured surfaces and depositional forms including a variety of morainal types. Such coasts, once released from their overburden of ice, rebound as is happening in northern Canada,

Scandinavia, and many parts of Antarctica especially on the Antarctic Peninsula. Some of the raised beaches on the Peninsula are as much as 60 m above present sea level (Kirk 1985).

Sea Ice and the Coast

Sea ice, one of the most variable elements in the oceanic system, varies seasonally in areal coverage by nearly 500% in the Southern Hemisphere but by less than 200% in the Northern Hemisphere (Figs. 1 and 2). These percentages show that sea ice is more strongly seasonal in the Antarctic than the Arctic. It is again a reflection of the nature of the two ocean areas involved. Because the Arctic Ocean is surrounded by land and the Antarctic Ocean surrounds a continent, ice formation and movement and therefore, impact on the coastlines is quite different. Most of the Arctic Ocean sea ice is of the multi-year type whereas most (~85%) of that of the Antarctic is first-year ice.

During the Arctic winter, sea ice not only affects the coasts surrounded by the Arctic Ocean but also extends south to locations such as Hokkaido, Japan in the North Pacific Ocean and New Foundland in the North Atlantic (Fig. 2). In contrast, during summer virtually all Arctic coasts are free of sea ice for varying lengths of time. Exceptions include northern Greenland, Ellesmere, and part of the Canadian Arctic Archipelago where sea ice may last throughout the year. Of major interest and concern is the great variability in sea ice cover and thickness. An analysis of satellite passive microwave observations shows those areas exhibiting negative trends in

the sea ice season are larger than those exhibiting positive trends (Parkinson 2000). If such negative trends continue, ice-free periods along Arctic coasts will continue to lengthen.

During the austral winter, sea ice extends north several hundred kilometers from Antarctica even reaching 55°S in the Indian Ocean (Fig. 1). During that period of the year, sea ice is present along the entire coastline although nearshore in many locations are polynyas – some maintained by katabatic winds. As summer approaches, sea ice drifts out and away from most of the coast except for a few locations where it remains attached to the shore throughout the summer (Wadhams 2000).

Sea Ice and its Impact on Coasts during Summer

For a few months during summer most of the coasts in both the Antarctic and Arctic are ice-free. However, in the Arctic Ocean especially, the permanent (although highly mobile) pack is often close enough to shore to dampen waves and thus reduce their impact on the coast. During those periods of time when the pack retreats from the coast thereby increasing fetch over coastal waters, storms can cause severe erosion as happened at Barrow, Alaska in October 1986 (Walker 1991). The shorter the ice-free period and the narrower the shore lead, the more limited the wave action alongshore.

Under certain wind and current conditions pack ice can move onto shore even at the height of the summer season causing some ice scour and sediment transport. This situation is especially true along the Beaufort Sea coast.

Freeze-Up and the Ice Foot

The factors affecting the timing of freeze-up include temperature, wind, snow, waves, tides, and the nature of the shoreline. When the temperature is lowered to the value at which seawater freezes, ice forms on the foreshore and within the interstices of shore sediments. Ice buildup occurs through the addition of the spray, swash, slush ice, and ice floes brought

by waves and tides plus the addition of snow. The accumulation becomes the ice foot, a major characteristic of ice-bordered shore (Fig. 5). The form, structure, and extent of the ice foot varies with shore gradient as well as tidal range and wave conditions during formation (Taylor and McCann 1983). A gently sloping shore face and a high tidal range favor the development of a wide ice foot whereas variable wave conditions often produce complex structures (Owens 1982).

The ice foot, composed of a mixture of sediments, snow and ice, rests on a beach surface that is also frozen. As waves approach the ice foot, they continue to deliver sediment from offshore either as loose particles or as material already incorporated into the pancake and brash ices that are added to the ice foot mix (Evenson and Cohn 1979).

The ice foot is bottomfast and immobile. At its seaward edge it abuts floating ice that moves vertically with tidal and wave action. They are separated by tidal cracks along a line which has been referred to as a “hinge zone” (Forbes and Taylor 1994). The degree of roughness of the sea ice at the hinge zone tends to increase with increasing tidal range.

Once freeze-up is complete, wave action on the coast ceases and any sediment transport by longshore currents is confined to locations seaward of the bottomfast ice zone. However, exceptions do occur. In the case of the Beaufort Sea, for example, severe winter storms can produce override with ice being forced over the ice foot high up on the shore. These features are known among the Inuit as “ivu.”

Although bottomfast ice usually extends out to water depths of two or more meters, shorefast ice extends out over deeper water to distances of as much as 20–30 km (Taylor and McCann 1983) where it merges with drifting pack ice. Shorefast ice is relatively immobile especially when present between islands as in the Canadian Arctic Archipelago. However, along open coasts, as those facing the Beaufort and Chukchi Seas, shorefast ice may be subject to occasional drift. The area between the outer limit of shorefast ice and

Ice-Bordered Coasts,

Fig. 5 Stranded ice-foot at Barrow, AK in June



the drifting pack, known as the “stamukhi” zone, is characterized by large pressure ridges some large enough so that they last through the summer.

Ice Melt and Breakup

Although shore leads in the sea ice may open early during the breakup period, the ice foot is not directly affected. Its ice melts in place. The rate and timing depend mainly on temperature conditions and depending on the amount of ice and snow present, may last into summer (Fig. 5). During winter, snow accumulates on the irregular surface and may be quite thick especially if there is a cliff behind the beach. As sea ice begins to move, especially with offshore winds, it breaks apart and floats the outer edges of the bottomfast ice. In the process, much of the sediment that has been incorporated into the sea ice is transported off and alongshore. The last ice to be removed from the coast are the large ice masses that become stranded and often buried onshore. Their melt rate is affected by the sediment that may cover them (Fig. 6).

In contrast to the shore ice melt and breakup that occurs along the exposed shore is that occurring out from river

mouths. Off river mouths the gradients of subaqueous deltas are usually more gentle than those along other coastlines so that the bottomfast sea ice zone is wider. During river breakup, floodwaters progress out over the sea ice and in the larger rivers beneath it in the subice distributary channels that do not freeze to the bottom. The water flowing over the sea ice deposits much of its sediment on top of the ice before it reaches pressure-ridge cracks or potholes out from the bottomfast/shorefast ice boundary where it drains to continue flowing seaward (Walker 1974). The drainage through these holes (Fig. 7) create in the bottom what Reimnitz has labeled strudel-scour holes (Reimnitz and Bruder 1972). Most of the sediment deposited on top of the sea ice is later deposited in the delta as the ice melts. However, some of it is transported seaward and alongshore as the offshore ice begins to drift.

Sea Ice: Its Role in Erosion, Transportation, and Deposition

In addition to the transport of sediment and the development of strudel mentioned above, sea ice is involved in other geologic processes along coastlines. The movement of sea

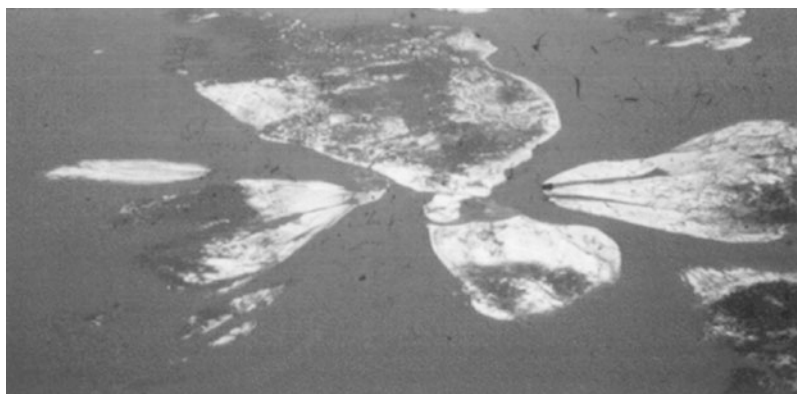
Ice-Bordered Coasts, Fig. 6

Five-meter-high sea ice pileup with accumulated snow onshore during melt season. Note the gradual settling of incorporated beach materials as ablation occurs



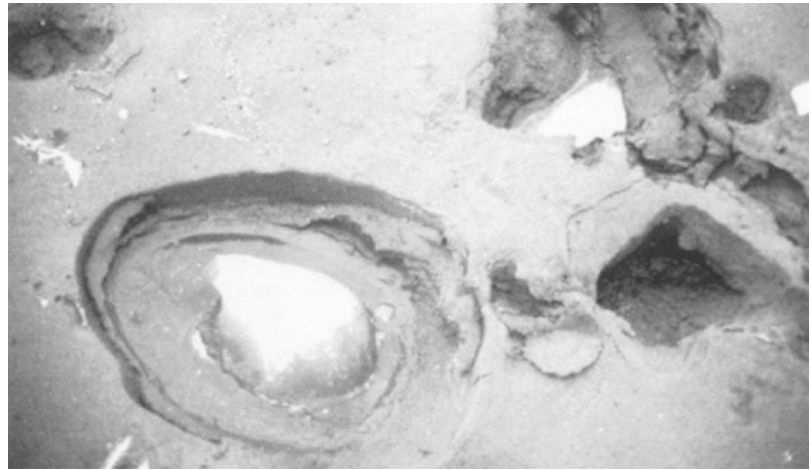
Ice-Bordered Coasts, Fig. 7

Drainage hole in sea ice at the Colville delta front. The floodwaters from the river drain from the ice creating strudel scour in deltaic sediments



Ice-Bordered Coasts, Fig. 8

Small kettles forming in the beach at Wainwright, AK as stranded ice blocks melt



ice on the beach and near the shore can produce scour marks and ridges. The uneven bottom of sea ice and especially when present as pressure-ridge keels gouge bottom materials as the ice drifts along the shore or up to the beach. As sea ice rides up onto the shore it both erodes and transports beach materials some distance above the high-tide shoreline. If the moving ice is impeded, ice pileups develop. They form not only at the high-tide shoreline but also at the hinge-line that joins bottom-fast and shorefast ice and around grounded floes. Such ice pileups impact heavily on both shore and nearshore sediments. Pileup heights of 20 m are common; some grow to double that (Forbes and Taylor 1994).

The beach which can have a very irregular surface at the end of the melt season because of the presence of ice-push ridges, kettles left by the melting of stranded ice blocks (Fig. 8), and a variety of scour forms is generally reworked by waves during the ice-free season. Only those forms that are especially large or that have been produced at the back of the beach may last through the summer.

Sea ice like glacial ice serves as both an agent of erosion and of protection. In the case of sea ice, the period of time it protects the coast from erosion varies from hours or days to year round in some very sheltered locations, whereas in the case of glacial protection the time period may be reckoned in millennia. With the rapid changes taking place in the icescapes of the world (e.g., the 15% reduction in sea-ice cover in the Arctic Ocean in the past 20 years (Krajick 2000)), the impact on the coastlines of ice-bordered coasts is continually in flux.

Cross-References

- [Antarctica, Coastal Ecology and Geomorphology](#)
- [Arctic, Coastal Ecology](#)
- [Arctic, Coastal Geomorphology](#)
- [Glaciated Coasts](#)
- [Paraglacial Coasts](#)

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Indian Ocean Coasts, Coastal Ecology

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Introduction

The coastal zone is the area covered by coastal waters and the adjacent shore lands, strongly influenced by each other. Coastal lands are some of the most productive and invaluable habitats of the biosphere, including estuaries, lagoons, and coastal wetlands. They are a place of high-priority interest to people, commerce, military, and to a variety of industries. Because it contains a dense population, the coast undergoes environmental modification and deterioration through reclamation, dredging, pollution, industry, and anthropogenic activities.

India has a vast coastline of approximately 7,000 km along the Arabian Sea in the west and the Bay of Bengal in the east. The western coastal plains lie between the Western Ghats and the Arabian Sea, further split into the Northern Konkan Coast and the Malabar Coast. The eastern coastal plains on the other hand, lie between the Eastern Ghats and the Bay of Bengal. Indian coasts have a large variety of sensitive ecosystems—sand dunes, coral reefs, sea grass beds, wetlands, mudflats, and rocky and sandy shores.

As shown in Fig. 1, there are number of backwaters, estuaries and coastal lagoons that support the rich and diverse flora and fauna (Fig. 2). These coastal habitats are considered to be highly productive areas in terms of biological productivity and one of the *hotspots* of marine biodiversity. Over 11,000 faunal (10,400 invertebrates and 625 vertebrates) and over 800 floral (624 algae, 50 mangrove, 32 angiosperms, 71 fungi, 14 lichens, 12 sea grass) species have been identified from Indian coastal areas (Untawale et al. 2000; Anon 2002). The majority of the shallow coastal and backwater

areas form an important spawning and nursery ground for commercially important fishes, molluscs, crustaceans, and various other species that constitute the coastal fishery of India.

Type of Marine Habitats

Mangroves

Mangroves are woody plants that grow at the interface between land and sea in tropical and subtropical latitudes where they exist in conditions of high salinity, extreme tides, strong winds, high temperatures, and muddy, anaerobic soils (Kathiresan and Bingham 2001). Mangroves are usually present in estuarine and muddy shores (Fig. 3), but can also be found on sand peat.

They are a complex and highly productive ecosystem that forms the interface between land and sea. The mangroves are widespread along the east and west coast of India. A detail of areas covered by mangrove habitat is given in Table 1.

On the Indian subcontinent, the mangrove ecosystems are distributed within the intertidal or tidal, supra-tidal or subaerial deltaic zones of both the east coast, facing the Bay of Bengal, and the west coast, facing the Arabian Sea. Mangrove flora of India comprises 50 exclusive species belonging to 20 genera, and 37 mangrove-associated floral species (Jagtap et al. 2002; Upadhyay et al. 2002). The maximum species diversity occurs in the Mahanadi delta along the Orissa coast, with 36 mangrove species present. Mangroves in India are estimated to cover about 4,871 km² (Upadhyay et al. 2002). The mangrove ecosystem on the Indian subcontinent is of three types.

1. *Deltaic mangroves*: These are found along the mouth of different major estuaries on the east coast, and two gulfs (Gulf of Kachchh and Gulf of Khambhat) on the west coast. However, deltaic mangroves cover up to 53% of the total Indian mangals, which are estimated to be about 2,560 km², out of which the Gangetic delta popularly, known as *Sunderbans*, alone covers about 78%. About 48 species of mangroves have been recorded from the east coast (Upadhyay et al. 2002). Mangrove distribution is scattered along the west coast with stunt growth and less species diversity.
2. *Coastal mangroves*: These are found along the intertidal coastlines, minor river mouths, sheltered bays, and backwater areas of the west coast. They extend from Gujarat to Kerala, and constitute to 12% of the mangals area of India. Due to less freshwater supply the mangals in the west coast are less, sparse, and show stunted growth. About 41 species have been reported from the west coast (Jagtap et al. 2002).
3. *Island mangroves*: They are found along the shallow but protected intertidal zones of bay islands, Lakshadweep



Indian Ocean Coasts, Coastal Ecology, Fig. 1 Some of the important coastal lagoons and marine biosphere reserves of India

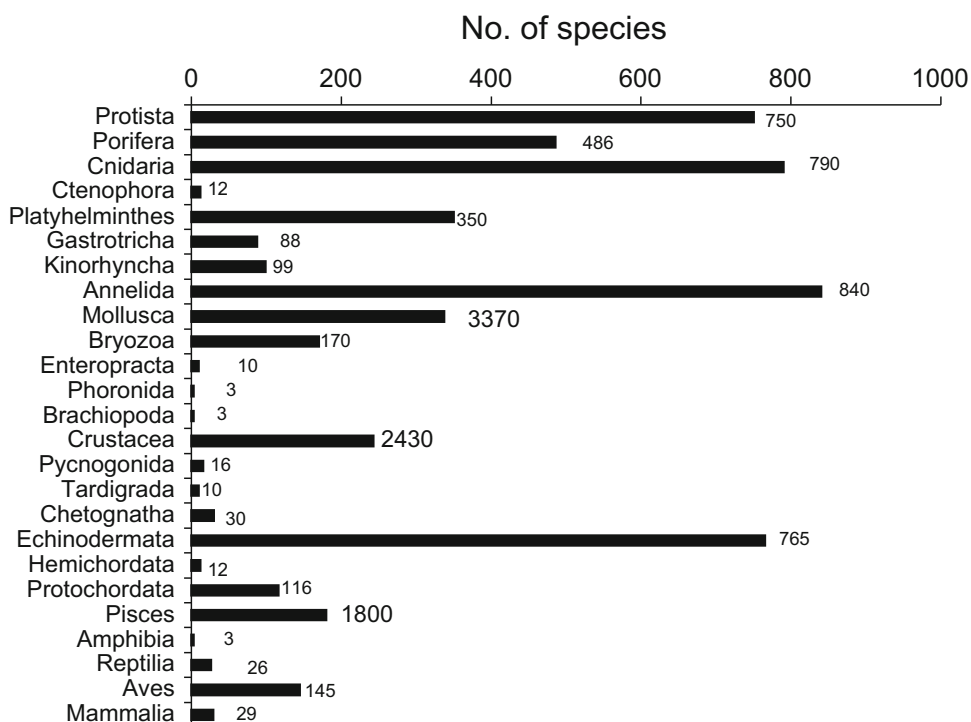
and Andamans. These are estimated to be about 16% (800 km²) of the total mangrove area. About 30 true species of mangroves have been recorded from the island areas.

Species of *Avicennia* and *Aegicera* are dominant vegetation in the Godavari–Krishna and Cauvery deltaic system while *Ceriops decandra*, *Sonneratia apetala* are dominant on the Mahanadi delta. About 33 species of mangroves have been reported from the Gangetic Sunderbans, with species such as *Heritiera fomes*, *C. decandra*, *Xylocarpus* spp., *Lumnitzera* sp., *Sonneratia alba*, *Kandelia candel*, *Nypa*

fruticans, and *Phoenix paludosa*. The mangroves of the West Bengal are dominated by *Excoecaria agallocha*, *C. decandra*, *S. alba*, *Avicennia* spp., *Bruguiera gymnorhiza*, *Xylocarpus granatum*, *Xylocarpus moluccensis*, *Aegiceras corniculatum*, and *Rhizophora mucronata*.

Species such as *R. mucronata*, *Rhizophora apiculata*, *Avicennia officinalis*, *Avicennia marina*, *Ceriops tagal*, *E. agallocha*, and *Acrostidium aureum* are most dominant along the west coast. Mangrove habitats harbor a variety of flora and fauna species (Figs. 4 and 5). Until recently, mangroves were treated as unwanted plants and were used largely as a source of timber and charcoal. Therefore, mangrove

Indian Ocean Coasts, Coastal Ecology, Fig. 2 Marine faunal diversity of India. (After Anon 1997, 2002; Untawale et al. 2000)



Indian Ocean Coasts, Coastal Ecology, Fig. 3 Estuarine mangrove habitat along the Indian coast



ecosystems have been severely depleted during the last two decades. According to recent surveys, deforestation has destroyed about 44% and 26% of mangroves along the west and east coast, respectively (Upadhyay et al. 2002). It is only in recent years that they have been recognized as ecologically vital. Mangroves play a very important role in protecting the shore from major erosion. The ecosystem forms an ideal nursery for the juvenile forms of many economically

important fish and prawn species. A large percentage of the detrital food, which supports a variety of young fish and shrimps, is generated from mangroves.

Management

Traditionally, mangroves have been utilized for their wood, mainly for construction, fuel, and stakes. The bark of *Rhizophora* and *Bruguiera* spp., are used for tannin

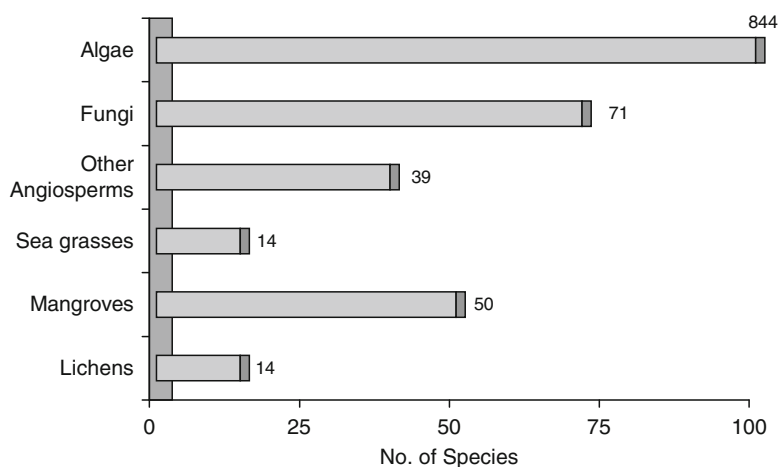
extraction. The leaves and fruits of *Avicennia* are used as fodder. *Avicennia* forests in the Gulf of Kachchh are constantly grazed by cattle. Mature fruits of *Sonneratia* as well as young fruits of *C. tagal* are consumed as vegetables in the human diet. Similarly, the young shoots of *A. aureum* and *Salicornia brachiata*, associated fern and an obligate halophyte, respectively, are also used as vegetables (Jagtap et al. 2002). Mud from the mangrove regions is used as manure for paddy and coconut fields. The roots and leaves of *Derris heterophylla* are used for narcotizing and stupefying fish. Extracts of *Acanthus* leaves are used for rheumatic disorders, while that of *Bruguiera* species for high blood pressure and *Rhizophora* extracts as a cure for jaundice. The bark and the leaves of *E. agallocha*, though poisonous, are used to cure rheumatism.

Indian Ocean Coasts, Coastal Ecology, Table 1 State-wise mangrove forest cover in India (in 1999)

	Area (km ²)
States along the east coast	
Tamil Nadu	21
Andhra Pradesh	397
Orissa	215
West Bengal	2125
Andaman & Nicobar Islands	966
Sub-total	3724
States along the west coast	
Karnataka	3
Goa	5
Maharashtra	108
Gujarat	1031
Sub-total	1147
Grand Total	4871

Source: Forest Survey of India (1998)

Indian Ocean Coasts, Coastal Ecology, Fig. 4 Floral diversity in mangrove habitat. (Compiled from Anon 2002)



Mangrove Fisheries

Mangroves have a rich and diverse fish assemblage and the habitat is commercially exploited for capture as well as captive fisheries. The capture fisheries mainly consist of various species of bivalve and gastropod, crabs, prawns, and fishes from the proper mangrove regions and estuarine waterways. The captive fishery includes fish and prawn farming in the mangrove regions as well as mussels and oyster culture in the estuarine region. The salt-affected, water-logged, tidal regions in the vicinity of mangrove environments are commonly used for paddy-cum-prawn farming and salt production.

Lagoons

Coastal lagoons are shallow water bodies lying parallel to the coastline and separated from the open sea by a narrow strip of land or salt bank (Fig. 6). They are a very rich and fragile natural ecosystem. As shown in Fig. 1, lagoons are distributed all along the Indian coast.

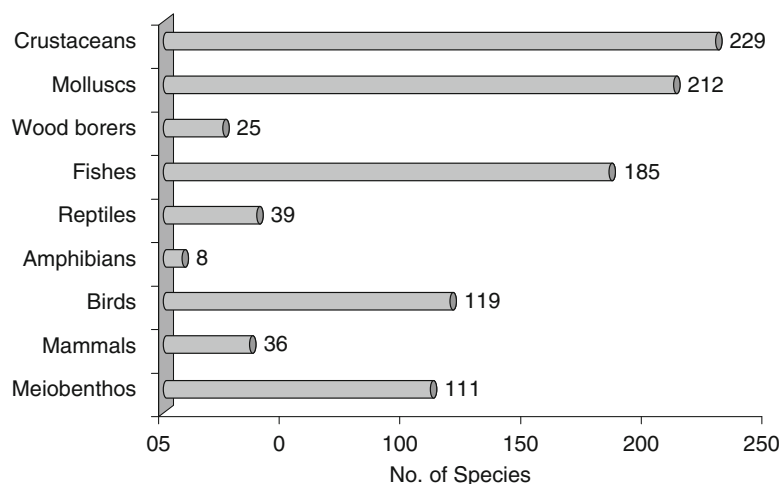
There are eight important lagoons along the east coast; they are the Chilika, Pulicat, Pennar, Bendi, Nizampatnam, Muttukadu, Muthupet. Chilika is the largest brackish water lagoon and Pulicat is the largest saltwater lagoon on the eastern seaboard of India. Chilika lagoon (Fig. 7) is spread over 1,100 km² while Pulicat lagoon is spread over 350 km².

There are nine important lagoons along the west coast of India (Fig. 1). They are Vembanad, Ashtamudi, Paravur, Ettikulam, Veli, Murukumpuzha, Talapady lagoon of the Bombay coast, and the Lakshadweep lagoons (Kavaratti and Minicoy). Vembanad and Ashtamudi are the largest coastal and backwater lagoons found in Kerala.

Halophila spp., *Thalassia* spp., *Cymodocea* spp., *A. marina*, *Acanthus* spp., *Xanthium* spp., *Acacia* spp., *Gracilaria*, *Asterionella* spp., *Enteromorpha* spp., are some of the floral species found in the lagoons of India.

The benthic fauna of lagoons constitute various species of, foraminifera, nematodes, gastrotrichs, oligochaetes,

Indian Ocean Coasts, Coastal Ecology, Fig. 5 Faunal diversity in mangrove habitat. (Compiled from Untawale et al. 2000)



Indian Ocean Coasts, Coastal Ecology, Fig. 6 Coastal lagoon at Vadhawan



polychaetes, calanoids, amphipods, isopods, decapods, tanaids, and molluscs. Among all the lagoons, Chilika has the richest biodiversity. Ecologically, it is endowed with a wealth of flora and fauna (Fig. 8). A total of 788 faunal species has been reported from the Chilika (Ingole 2002). The majority of these are aquatic and almost 29% (225) are fish species (Fig. 8). About 61 are protozoan species, 37 nematodes, 29 platyhelminthes, 31 polychaetes, 58 decapods and brachyuran crabs, 37 amphibians and reptile, 136 molluscs, 18 mammals, and 156 bird species.

Chilika Lake contributes to a major portion of the fish catch in the region. The lagoons provide an excellent opportunity for aquaculture and get a good foreign exchange. The rich demersal fishery (especially of prawn, crabs, and molluscs) supports over 80,000 fishermen from 122 villages. The

lake supports one of the largest populations of waterfowl during winter season. The area is known as an ideal habitat for crocodile, dolphin, and a variety of birds. An area of 15.53 km² of this lake has been designated as wildlife sanctuary. However, Chilika has been facing some natural and manmade problems, particularly frequent shifting of the mouth region, reduced seawater inflow, siltation, and encroachment. With a rise in human population in the lagoon periphery, pollution from domestic sewage, pesticide, agriculture, chemical, and industrial effluent have become major threats to the lagoon ecosystem. The lagoon was rapidly shrinking at the rate of about 2 km² per year. According to Ingole (2002) fish catches which used to be about 8,500 tonnes per year in 1980s were dwindling during 1985–2000.

Indian Ocean Coasts, Coastal Ecology, Fig. 7 Chilika lagoon (Orissa, India)

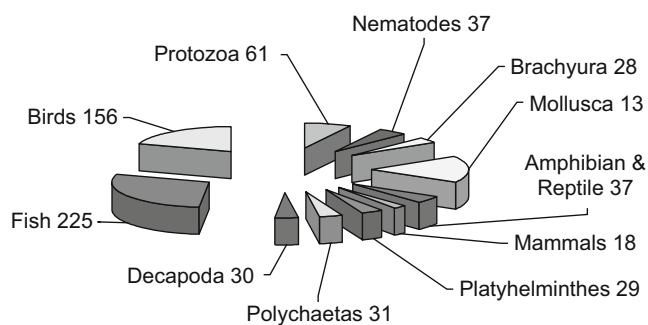


However, due to the timely action taken by the Chilika Development Authority (CDA, Government of Orissa), the lake environment is being restored under the “Chilika Development Plan.” Substantial increases in the fish production (11,989 tonnes) during 2001–02 (Ingole 2002) clearly demonstrated the efforts of CDA toward the sustainable development and conservation of this important ecosystem. Recently, due to the rapid recovery of the lake ecosystem, the CDA Authority has been awarded the International *Ramsar Award* and Chilika is also included in the list of *Ramsar Sites*.

Estuaries

Estuaries are an integral part of the coastal environment. They are the outflow regions of rivers, making the transitional zone between the fluvial and marine environment. Estuaries are the focal point of studies, and activities. Fourteen important estuaries have been reported along the east coast and west coast of India (Anon 1997; Qasim 1999). Hooghly, Rushikulya, Godavari and Krishna, Edaiyur-Sadras, Araniar, Ennore, Cooum, Adyar, Uppanar, Vellar, Kollidam, Kaveri, Agniyar, Kallar along the east coast, and Ashtamudi, Korapuzha, Beypore, Periar, Kadinamkulam, Vembanad, Netravathi and Gururpur, Gangolli, Pavenje, Kali, Narmada Amba, Purna, Mandovi, and Zuari estuary on the west coast of India.

Biodiversity in the estuaries is very impressive. Some floral species are *Oscillatoria* spp., *Enteromorpha* spp., *Spirogyra* spp., *A. marina*, *Excoecaria* spp., and *Sonneratia* spp. Various species of polychaetes, crustaceans, molluscs, echinoderms, and fish are the faunal component of estuaries.



Indian Ocean Coasts, Coastal Ecology, Fig. 8 Biodiversity of Chilika Lake ((No. of species); Ingole 2002)

Estuaries are semi-enclosed water bodies and thus, they provide a natural harbor for trade and commerce. They are also effective nutrient traps and provide a vital source of natural resources for people and are used for commercial, industrial, and recreational purposes. They also act as nursery grounds for a variety of shrimps and finfish and are the best settling places for clams and oysters.

Indian estuarine ecosystems are deteriorating day by day through human activities, and the dumping of an enormous quantity of sewage into the estuary has drastically reduced the population of spawning fishes. It has also caused considerable ecological imbalance and resulted in the large-scale disappearance of the flora and fauna. Introduction of untreated municipal wastewater and industrial effluents into these water bodies has led to serious water pollution including heavy-water pollution, which becomes bio-magnified and reaches people through the food chain.

Mudflats

Mudflats develop in sheltered places of the intertidal area. Twice each day, water flows in and out with the tides, filling or draining the flat. The mudflat receives nutrients from the tidal flow and the nearby marsh, particularly as it decays. This means that mudflats have rich plant and animal communities.

They are important as sedimentation areas and provide a rich source of organic material for the endo- and epibenthic community. Because of this high availability of organic substances, oxygen content in the pore water is rather low and may limit chemical and biological degradation processes.

Phytoplankton and zooplankton are abundant, as are mud snails. Filter-feeding animals such as oysters and clams live in mudflats because of the availability of plankton. Fish and crabs move through the flats at high tide. Birds and predatory animals visit tidal flats at specific times for feeding.

Sandy Beaches

The sandy beaches seem to be barren. There is no lush growth of macro-algae and apart from sea grasses there is no obvious plant life. Few animals live on the surface and most of them live below the sand. The organisms found here have suitable burrowing mechanisms, which may take the form of proboscis, parapodia of a polychaeta, and foot of molluscs.

Vast stretches of sand are seen along the east and west coast of India forming the boundary between land and sea. The beaches are subjected to the forces of waves, tides, currents, and winds. Sandy beaches along the Gujarat coast are limited between muddy and rocky shores. Although sandy strips along rocky cliffs are also observed along the Maharashtra and Goa coast, beaches along the central west coast of India are of sandy nature. The sandy shores of Goa and Karnataka (Fig. 9) are of limited width, while Kerala has extensive sandy beaches interspersed with coastal lagoons. Tamil Nadu has sand strips along the deltaic shores and rock-bound beaches. Beaches along the Andhra Pradesh coast are of limited width interspersed by the rivers Godavari, Krishna, and their tributaries. Orissa has extensive sand strips at Konark—Puri. The coast of Andaman and Nicobar Islands have sand strips interspersed with bluffs, rocks, or shingle along the coastline. The Lakshadweep atolls have long stretches of coralline sandy beaches with unique vegetation.

Macrofaunal species (benthic organisms having body size >0.5 mm) such as *Donax incarnates*, *Donax spiculum*, *Donax faba*, *Donax scortum*, *Suneta scripta*, *Mactra* spp., *Paphia malabarica*, *Bullia melanoides*, *Umbonium* spp., *Oliva* spp., *Emerita holthuisi*, *Eurydice* spp., *Gastrosaccus* spp., *Ocypode ceratophthalma*, *Ocypode macrocera*, *Ocypode platysus*, *Dotilla intermedia*, *Glycera alba*, *Lumbriconereis latreilli*, *Onuphis eremita* are some of the common species found on the sandy beaches of India.

Sand Dunes

The term “sand dune” reflects the image of vast amounts of shifting sand, barren of plants, and hostile to human habitation. Hot and dry winds shape and arrange the sand in geometric and artistic patterns. Dunes are of two types. The first type is found in the extremely dry interior desert such as Rajasthan in India and the other type is the coastal sand dune, which occur along the coast of India.

Dunes are composed of wind-blown sand. Fore-dunes are built up at the back of the beaches on the crest of berms and dune ridges where vegetation or other obstacles trap wind-blown sands. During periods of coastline advancement, successful dunes may develop to form a series of parallel dunes (Fig. 10).

The common vegetation found in the fore-dune are *Hydrophyllax maritima*, *Ipomoea pes-caprae* (Saurashtra), *Canavalia maritime*, *Cyperus arenarius*, *I. pes-caprae*, *Launea sarmentosa*, *H. maritima* (Orissa), *H. maritime*, *I. pes-caprae*, *C. maritima*, *C. arenarius*, *I. pescaprae*, *L. sarmentosa*, *H. maritime*, *Sporobolus virginicus* and *Zoysia matrella* (Andhra Pradesh), *I. pes-caprae*, *L. sarmentosa*, *Dactyloctenium aegypticum*, *C. arenarius* (West Bengal), *I. pes-caprae*, *Spinifex littoreus* (Goa). In the parabola dune the following vegetation have been recorded along the west coast of India: *Halopyrum mucronatum*, *Borreria articularis*, *Lotus garcinii*, *Asparagus dumosus*, *Enicostema hyssopiflorum*, *Peplidium maritimum*, *Cassytha filiformis*, and along the east coast of India *Euphorbia rosea*, *Geniospermum tenuiflorum*, *Phyllanthus rotundifolius*, *S. littoreus*, *Goniogyne hirta*, *Perotis indica*, *Brachiaria reptans*, *Elusine indica*, *Rothia indica*, *Trianthema pentandra*.

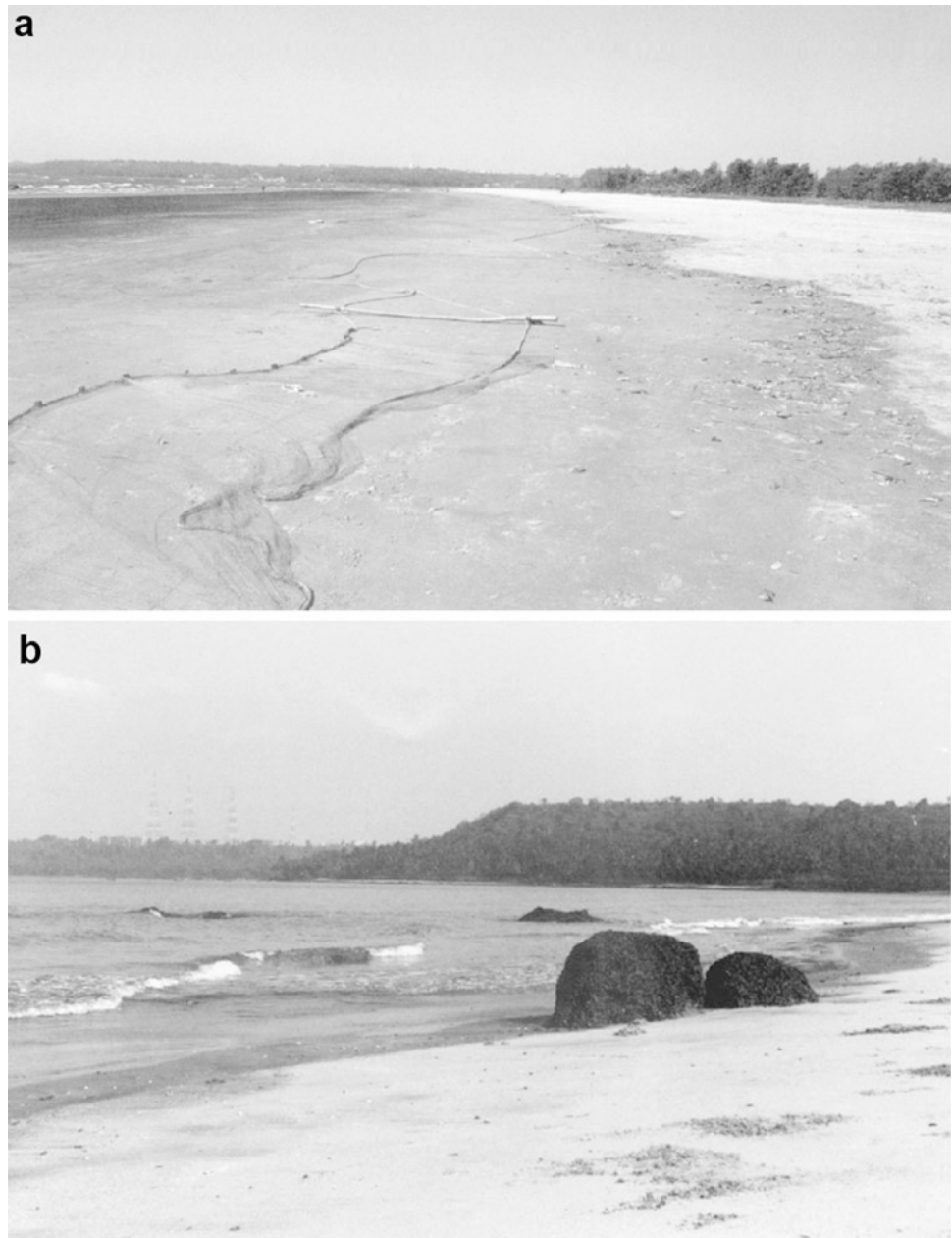
Sand dunes throughout the world have been recognized for their ecological significance. The coastal dune vegetation acts as shelter belts protecting the inner land. When the dune system and the vegetation are destroyed for short-time benefits in the name of development, disaster occurs. Floods and cyclones on the east coast of India are eye-opening examples. The cause and effect of destruction are not short-termed or locally limited.

Rocky Shores

In contrast to sandy beaches where many individuals live unseen in the soil, many of the plants and animals of the rocky shores are conspicuously displayed. Another feature of the rocky shore is the distinctly noticeable zonation of plants and animal communities.

The rocky coast shows a richer fauna than that of the sandy beach and is varied in composition. The rock and crevices give shelter to numerous crabs, molluscs, and fish. *Grapsus grapsus*, *Tectarius trochoid*, *Littorina scabra*, *Littorina angulifera*, *Littorina undulat*, *Trochus* sp., *Cellana radiata*, *Planaxis sulcatus*, *Thais bufo*, *Drupa* sp., *Hemifusus pugulinus*, *Perna viridis*, *Nereis* spp., amphipods, isopods,

Indian Ocean Coasts, Coastal Ecology, Fig. 9 Sandy beaches of Goa (Central west coast of India)



Holothuria, Sea Urchin, nudibranch, *Aplasia* spp., are some of the common fauna of the rocky shores. Common flora of the rocky shore are different species of *Porphyra*, *Gracilaria*, and *Enteromorpha*, *Padina*, *Ulva*, *Gelidium*, *Sargassum*, and *hypnea*, *Chaetomorpha* spp.

Sea Grass

Sea grasses are submerged flowering plants (angiosperms) that have adapted to life in the sea. They differ from what we refer to as “sea-weed,” in that they are plants with vessels and well-defined root and shoot systems. Sea grasses have been able to successfully colonize the marine environment because of five properties:

- the ability to live in a salty environment;
- the ability to function normally when fully submerged;
- a well-developed anchoring system;
- the ability to complete their reproductive cycle while fully submerged;
- the ability to compete with other organisms under the more or less stable conditions of the marine environment.

Sea grass habitats are mainly limited to the mudflats and sandy regions from the lower intertidal zones to a depth of ca. 10–15 m along the open shores and in the lagoons around islands. The major sea grass meadows in India occur along the southeast coast (Gulf of Mannar and Palk Bay) and along a

Indian Ocean Coasts, Coastal Ecology, Fig. 10 Typical sand dune found along the east coast of India



number of the islands of Lakshadweep in the Arabian Sea and of Andaman and Nicobar in the Bay of Bengal. The largest area (30 km²) of sea grass occurs along the Gulf of Mannar and Palk Bay, while it is estimated that ca. 1.12 km² occurs in the lagoons of major islands of Lakshadweep. A total of 8.3 km² of sea grass cover have been reported from the Andaman and Nicobar Islands (Jagtap et al. 2003).

The sea grasses of India consists of 14 species belonging to 7 genera. The Tamil Nadu coasts harbor all 14 species, while 8 and 9 species have been recorded from Lakshadweep, and Andaman and the Nicobar group of Islands, respectively. The east coast supports more species compared to the west coast of India. The main sea grasses are *Thalassia hemprichii*, *Cymodocea rotundata*, *Cymodocea serrulata*, *Halodule uninervis* and *Halophila ovata*. Species such as *Syringodium isoetifolium* and *Halophila* spp., occur in patches as mixed species. Gulf and bay estuaries mostly harbor low numbers of species, dominated by *Halophila beccarii* in the lower intertidal region and by *Halophila ovalis* in the lower littoral zones.

Sea grass beds are:

- Major primary producers (food manufacturers) in the coastal environment. Their high primary productivity rates are linked to the high production rates of associated fisheries;
- Stabilizers of bottom sediments; they provide protection against erosion along the coastline;
- Important nutrient sinks and sources, that is they help in the recycling of nutrients.
- With root-like stems, which extend horizontally under the sea bottom, sea grasses act to stabilize the sediment. These

sediments, that would otherwise settle on coral and prevent contact with sunlight, tend to accumulate and become trapped in the sea grass. Turtle grass, the most common type of sea grass thrives in areas that are protected from wind-driven current and surf. The broad leaves of turtle grass act as huge filters, removing particles from the water and depositing them as fine sediment. These sediments often contain organic matter, which contribute to the high productivity of this habitat. For this reason, the sea grass habitat attracts various species of fish, conch, lobster, turtles, and manatees for feeding and breeding. Numerous species of reef fish use sea grass as a protective nursery, hiding amid the grass from predators. Moreover, adult fish that hide in the coral reef during the day and venture out at night to feed, take advantage of the rich source of food that exists in the sea grass.

The natural causes of sea grass destruction in India are cyclones, waves, intensive grazing, and infestation of fungi and epiphytes as well as “die-back” disease. Exposure at the ebb tide may result in the desiccation of the bed. Strong waves and rapid currents generally destabilize the meadows causing fragmentation and loss of sea grass rhizomes. The decrease in salinity due to extensive freshwater run-off also causes disappearances, particularly of the estuarine sea grass bed.

Anthropogenic activities such as deforestation in the hinterland or mangrove destruction, construction of harbor or jetties, loading and unloading of construction materials as well as anchoring and moving of boats and ships, dredging and discharge of sediments, land filling and untreated sewage disposal are some of the major causes of sea grass destruction of India.

As of yet, there is no specific legislation protecting sea grasses, although generally the Fisheries Department is responsible for this habitat. Green Reef recommends that in addition to controlling the use of pesticides and decreasing run-off, dredging activities need to be strictly monitored and limited. Sea grass beds should be thought of as an indication of the health of the ecosystem; when they begin to disappear, we know there is trouble.

Coral Reefs

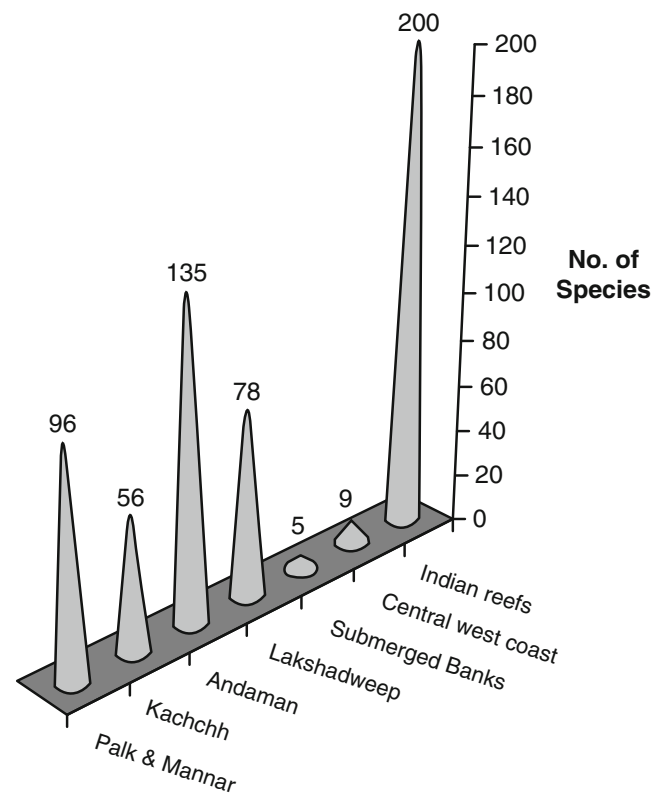
Coral reefs are among the most biologically productive, taxonomically diverse, and aesthetically important living organisms among all the aquatic ecosystems.

The major coral formation in India is around the Lakshadweep (816 km²) and Andaman Islands (960 km²), as well as in the Gulf of Mannar (94 km²) and the Gulf of Kachchh (406 km²). Reefs in the Gulf of Kachchh, Gulf of Mannar, and Andaman and Nicobar Islands are mostly of the fringing type, with a few platform, patch and atoll reefs, and coral pinnacles. Lakshadweep Islands on the contrary are mostly atolls with a few coral heads, platform reefs, and sand cays. Coral, though rare in occurrence, are reported at many locations along the west coast.

About 44 species of scleractinian coral and 12 species of soft corals occur in the Gulf of Kachchh (Fig. 11). The submerged reefs of these areas harbor 18 species of stony coral and have 45% coverage. *Acropora*, *Porites*, *Pseudosiderastrea*, and *Favia*, and one species of soft coral *Juncella juncea* are common. Two species of hard coral (*Pseudosiderastrea tayamai* and *Porites lichens*) have been recorded in patches along the central west coast. Large well-developed hard colonies have also been sighted at Colaba near Mumbai.

Malvan is considered to have the richest marine biodiversity along the central west coast. *Porites*, a stony coral is most common in Malvan waters. Nine species including a rare coral species belonging to the genera *Coscinaraea*, *Favites*, *Goniastrea*, *Synaraea* and *Pseudosiderastrea*, *Cyphastrea*, *Turbinaria* have been recorded in this region. Angria bank—a submerged reef off Ratnagiri has stony corals. A few patches of scleractinian corals have been reported in subtidal waters off the Goa coast (Rodrigues et al. 1998). Five coral species have been recorded from Gaveshani Bank off Malpe along the Karnataka coast (Qasim and Wafar 1979; Wafar 1990). *Pocillopora eydouxi* has recently been reported from Vishakhapatnam on the east coast.

About 96 species belonging to 36 genera of Scleractinian and 7 species of Ahermatypic corals are reported from Palk Bay and the Gulf of Mannar (Fig. 11). *Montipora* and *Acropora* represent 40% of the species in the Palk Bay. Thirty species of stony corals are reported from Manauli reef, of which the dominant corals are *Acropora*, *Porites*, *Goniastrea*, *Favia*, *Pocillopora* and *Montipora*. Krusadai island has a



Indian Ocean Coasts, Coastal Ecology, Fig. 11 Coral distribution along the Indian coast

well-developed coral reef and sustains 96 species belonging to 26 genera of hermatypic corals.

Massive occurrences of corals provide much needed protection from waves to the coastline; and coral productivity yields a multitude of flora and fauna dependant on the coral ecosystem. Coral reefs also provide opportunities for skin diving, under-water photography, sport fishing, and shell collection, thus, providing a vital stimulus to the tourist industry. Coral reefs are often exploited for calcium carbonate—a raw material for many lime-based industries. The fishery resources of the reefs are extremely rich and diversified. They are also exploited for their beautiful associated fauna such as molluscs and ornamental fish.

Due to population pressure, most of the coral reefs have become extremely vulnerable to industrial development and pollution along the coastline. Unless protection is offered to the coral reefs, most of them will diminish in size in the future and ultimately die. The reefs that will probably flourish are on the atolls of Lakshadweep and some of the islands of Andaman and Nicobar Islands.

Marine Protected Areas

There are 26 Marine Protected Areas (MPAs) in India comprised of national parks and wildlife sanctuaries declared as coastal wetlands; especially mangrove, coral reefs,

Indian Ocean Coasts, Coastal Ecology, Table 2 Marine national parks and sanctuaries along the Indian coast (Anon 2002)

States	Location	Name of the biosphere
Gujarat	Gulf of Kachchh	National Marine Park, Bird Sanctuary, and coral reef
Maharashtra	Mahaim (Mumbai) Malvan	Natural Mangrove Park Marine Park (coral reef)
Goa	Chorao island	Bird Sanctuary
Lakshadweep island	Kawaratti, Agatti, Minicoi islands	Marine Park (coral reef)
Tamil Nadu	Gulf of Mannar	National Marine Park, Sanctuary, and coral reef
	Muthupet	Reserve Mangrove Forest
Andhra Pradesh	Coringa	Wildlife Sanctuary
Orissa	Nalaban (Chilika)	National Marine Park and Bird Sanctuary
	Bhittarkanika	Wildlife Sanctuary (Turtle project)
West Bengal	Haribanga	National Marine Park and Bird Sanctuary (Tiger project)
Andaman and Nicobar Island	Wandoor	National Marine Park

and lagoons, under Wildlife (Protection) Act, 1972. These 26 MPAs are located in 7 coastal states and 2 union territories and cover 13% of the total coastal wetland area of the country. Some of these MPAs are shown in Fig. 1 and Table 2. The Gulf of Kachchh Marine Sanctuary and Marine National Park, the Gulf of Mannar National Park, and the Wandoor Marine National Park (Andaman Islands) have been established primarily to protect marine habitats.

Management of Coastal Ecology

As shown in Fig. 12, there is great demand for industrial development along the Indian coast. This in turn amounts to the increase in pressure on the coastal zone due to concentration of population, development of industries and ports, discharges of waste effluents and municipal sewage, and spurt in recreational activities which have adversely affected the coastal environment. Coastal resources are affected by activities far distant from the coast (viz., deforestation, damming of rivers, bunding/barraging of the coastal water bodies, river sand mining, discharge of pesticides, heavy metals, domestic and factory wastes, garbage, and other substances that are harmful to the coastal areas. Lagoons, estuaries, wetlands, and nearshore shallow water areas are particularly vulnerable to these activities. Considering the urgent need for protecting the Indian coast from degradation, the Government of India enacted “Coast Regulation Zone (CRZ) Act, 1991.” The area influenced by tidal action up to 500 m from High Tide

Line (HTL) and the land between the Low Tide Line (LTL) and the HTL has been declared as Coastal Regulation Zone (CRZ).

Mangroves and Coral Reef Ecosystem

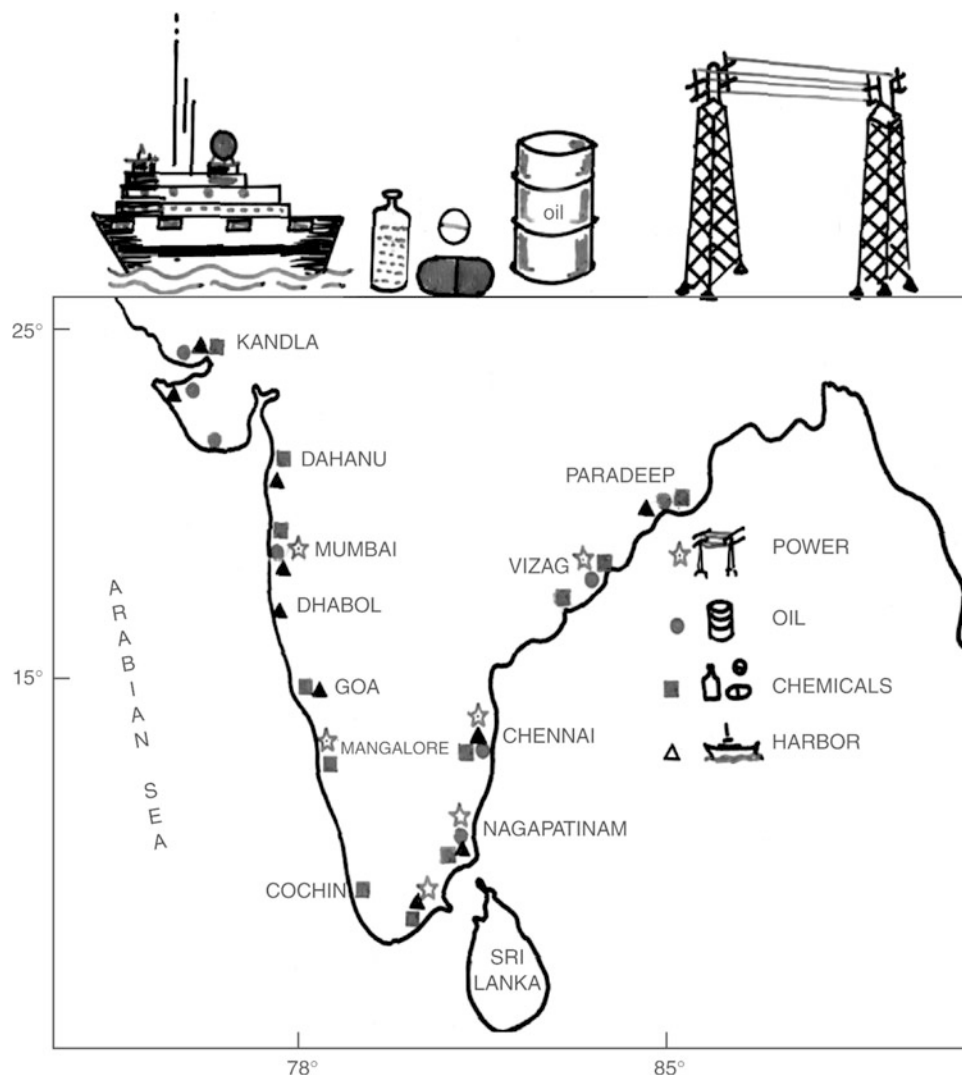
The mangroves are constantly threatened by increasing anthropogenic pressures such as indiscriminate cutting, reclamation mainly for agriculture and urbanization, fuel and construction, and overgrazing by domestic cattle. Mangrove ecosystems along the west coast, particularly in the States of Gujarat, Maharashtra, and Kerala, have been degraded to a large extent. However, West Bengal (Sundarbans), Orissa, and the Andaman and Nicobar Islands still form the best mangrove ecosystems of India. Considering various estimates for the mangrove cover in the country, 30% of the mangrove area was reclaimed for different anthropogenic activities during the period 1975–90 (Jagtap et al. 1993).

Realizing the importance of mangroves and coral reefs, the Government of India initiated efforts for their conservation and management. They were declared as ecologically sensitive areas under the Environment (Protection) Act, 1986, banning their exploitation, and followed by a CRZ Notification 1991 prohibiting development activities and disposal of wastes in the mangroves and coral reefs. The Ministry initiated a plan-scheme on conservation and management of mangroves and coral reefs in 1986 and constituted a National Committee to advise the Government on relevant policies and programs. On the recommendations of this committee, 15 mangrove areas in the country have been identified for intensive conservation (Anon 1997, 2002; Jagtap et al. 2002).

Considering the importance of coral reefs and the factors responsible for their deterioration, Andaman and Nicobar Islands, Lakshadweep Islands, Gulf of Mannar, and Gulf of Kachchh have been identified for conservation and management. Efforts have been initiated to establish Indian Coral Reef Monitoring Network (ICRMN) to integrate various activities on coral reefs through national and international initiatives.

Institutions of database networking and capacity building and training on coral reefs have been identified. With the establishment of the National Mangrove Committee (NATMANCOM), attempts are being made to protect, conserve, and restore the mangrove habitats. Mangrove regions in the country have been categorized presently under the Ecologically Sensitive Zone; vide CRZ Act of the country. As per the CRZ Act, no development in the mangroves or in the vicinity is allowed prior to an environmental impact assessment (EIA) and clearance from the Ministry of Environment and Forests (MoEF), Government of India. Few of the mangrove regions in the country have been conserved as Biosphere reserves for germplasm and wildlife sanctuaries.

Indian Ocean Coasts, Coastal Ecology, Fig. 12 Patterns of industrial development along the Indian coast



Shrimp culture Activities

The aquaculture industry is growing at a faster rate than many of the sectors of the coastal zone in India. Socially, its product is seen as a currency. However, if aquaculture expansion is not regulated, its long-term consequences will be felt in the quality of water bodies. In fact some adverse effects of shrimp culture are already seen along the east coast of India, particularly along the coast of Andhra Pradesh.

Sandy Shores

Mining of beach sand is a widespread activity in many of the coastal areas. Though sand is an important constituent of construction activities, it has to be borne in mind that it is these sand deposits that provide the natural protection to the coast from erosion. Storms, waves, currents, and wind temporarily displace vast quantities of beach sand that is then held in storage as sand bars. These sand bars then become the protectors of the coasts against those forces, which finally

return the sand to the beach. Thus, removal of sand from any part of the beach can aggravate the erosion and recession of the beach front altogether.

For the above reasons and also for the sustainable management of the coastal resources, many countries are now developing Coastal Zone Management (CZM) strategies, and some have already begun to adopt such programs. The CZM has to manage, develop, and conserve natural resources and, while doing so, it has to integrate the concerns of all relevant sectors of the society, economy, and prosperity.

A major thrust in implementing the CRZ notification is in the conservation of coastal resources, to achieve their sustainability, and long-term protection of its natural assets. The criterion for sustainable use is that the resource shall not be harvested, extracted, or utilized in excess of the quantity that can be produced or regenerated over the same period. It is important to learn the acceptable limits of coastal

environmental degradation and the limits of sustainability of coastal resources. Hence, in order to achieve the goals set forth in the notification, it is imperative to have first-hand information on the present land use practices and availability of the resources in the coastal zone. Contribution No.3897 of NIO, Goa.

Cross-References

- [Aquaculture](#)
- [Bioconstruction](#)
- [Coral Reef Coasts](#)
- [Estuaries](#)
- [Human Impact on Coasts](#)
- [Indian Ocean Coasts, Coastal Geomorphology](#)
- [Indian Ocean Islands, Coastal Ecology and Geomorphology](#)
- [Mangroves, Ecology](#)
- [Mangroves, Geomorphology](#)
- [Muddy Coasts](#)
- [Vegetated Coasts](#)
- [Wetlands](#)

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Indian Ocean Coasts, Coastal Geomorphology

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The coastal geomorphology of the Indian Ocean coast, with special reference to coasts of Pakistan, India, Sri Lanka, Bangladesh, and Myanmar, is mainly governed by the processes associated with monsoons.

Pakistan

The coastline of Pakistan, from the Iranian border on the west to the Indian border on the east is about 990 km long. This coastline is one of the active tectonic regions. The coast here is associated with a narrow continental shelf, except off Indus delta. The coast of Pakistan is divided into the Makran coast, Las Bela coast, Karachi coast, and Indus Delta coast.

The Makran coast, with approximately 473 km length, from the Iranian border to Ras Malan, consists of long sandy beaches associated with either wide coastal plains or valleys landward. These plains and valleys are interrupted by uplifted marine terraces at places. Also the Makran hill ranges, which lie about 32 km from the coast, become part of the coast at Ras Malan with massive headlands. Spits and bars are common seasonal morphologic units along sandy beaches. At places, well-developed beach ridges are seen. Dasht is the only river which brings a small quantity of sediment from land to the Arabian Sea along this coast.

The Las Bela coast, with about 260 km length, extends from Ras Malan to Ras Muari. The Ras Malan range is made up of sandstone and shelly limestone and presents gorges and cliffs as high as 600 m, which drop directly to the sea. Followed by this, on the east, is the Las Bela plain. The coast here consists of a series of beach ridges, sand dunes, bars, tidal flats, and lagoons with mangroves (Bird and Schwartz 1985). Between the Ras Malan ranges and the Las Bela main valley notable mud volcanoes are present, the largest among them is called Chandragup (Snead 1964). On the eastern side of the Las Bela valley, promontories of limestone are present. Here marine terraces at different elevations have wave cut sea caves and blow holes (Snead 1966). Along this coast, the Hab

River joins the Arabian Sea at its mouth compound bars, shallow lagoons, and sandy beaches are present.

The Karachi coast, about 48 km in length from Ras Muari to Clifton beach, consists of low rocky cliffs and sandy beaches of almost equal length. Marine terraces, sea caves, and arches are common in the sandstone and shale rocks. Sandy barrier beaches, spits (the longest one 15 km in length), shallow lagoons, tidal flats, and salt evaporation ponds are common along the sandy beaches.

The Indus delta coast is about 200 km long with uniform landforms namely large tidal channels with mudflats in between, barrier bars and spits with hooks, and beaches, and a few small mangrove shrubs. The sand bars and delta channels are dynamic in nature as they change their morphology due to tidal currents, waves and channel floods. The coastal area in this region is very flat and therefore up to 6.5 km from the coastline it is submerged during high tide. The river Indus brings a large quantity of sediment from land to sea and joins the Arabian Sea along this coast.

India

Most of the early literature on Indian coastal geomorphology was essentially of a descriptive nature based on the nature, location, and relationships of the landforms and sea level. Ahmad's (1972) was possibly the first and only book on coastal geomorphology of India, and contains data collected from large-scale maps and inferences drawn on the nature of the coasts. In addition, there are some isolated studies by Vaidyanathan (1987), Baba and Thomas (1999). The Space Application Centre (SAC 1992) has carried out a comprehensive study on the coast using LANDSAT and IRS data. The information on Indian coastal geomorphology presented in this article is based on these and many more isolated published studies.

India has about a 7,500-km long coastline. The coastline of India has been undergoing morphological changes throughout the geological past. The sea level fluctuated during the period of last 6,000 years and recorded marked regression during the period between 5,000 and 3,000 years before present (Rajendran et al. 1989). The present coastal geomorphology of India has evolved largely in the background of the post-glacial transgression over the preexisting topography of the coast and offshore (Baba and Thomas 1999).

The major rivers that cut across the coast and bring large quantities of water and sediment to the coast from Indian continent are the Ganges, Brahmaputra, Krishna, Godavari, and Cauvery on the east coast, and the Narmada and Tapi on the northwest coast. In addition, there are about 100 smaller rivers, these also supply considerable quantities of water and sediment. While larger rivers have well-developed deltas and estuarine systems, almost all the small rivers have estuarine

mouths with extensive mud flats and salt marshes and some of them with estuarine islands.

The continental shelf of India is very wide on the west coast with about 340 km in the north, tapering to less than 60 km in the south. The shelf is narrow along the east coast. The coastline on the west receives southerly winds that bring high waves during the monsoons (June–September). The east coast generally becomes active during the cyclones of the northeast monsoon period (October–November). The tidal range varies significantly from north to south. It is around 11 m at the northwest, 4.5 m at the northeast and around 1 m at the south.

Considering geomorphic characteristics, the Indian coast is divided into two categories, namely coasts on the west coast of India and coasts on the east coast of India. The coast on the west coast of India differs from the east in that there are practically no deltas on the west coast. The coastline here is modified by headlands, bays, and lagoons at irregular intervals. There is distinct evidence of the effect of neotectonics in some sections (Vaidyanathan 1987). The east coast on the contrary is known for the number of deltas especially along the northern portion, West Bengal and Orissa coast. Deltas in the southern portion have helped in recognizing ancient channels, ancient beach ridges, former confluences, and strandlines.

West Coast of India

Though there are a large numbers of small rivers bringing enormous quantity of sediment to the Arabian Sea along the west coast, deltas are not formed, possibly due to the high-energy condition of the coast. Beach morphological changes along the west coast are controlled by the southwest monsoon. The maximum morphological changes occur during early monsoons (June–August). During this period most of the material is transported to the offshore and some along-shore. Most of the material appears to be returned again during the fairweather season.

The west coast of India is further divided into the Gujarat coast, Maharashtra, Goa and northern Karnataka coast, and southern Karnataka and Kerala coast based on their geomorphological distinctions. The coastal area of Gujarat is the largest in the country with about 28,500 km². The coast in Gujarat, from west to east, varies from a deltaic coast, the irregular drowned prograded coast, the straightened coast, the spits and cusped foreland complex, and the mudflat coast. The Gujarat coast is further divided, from west to east, into five regions, namely Rann of Kutch, the Gulf of Kutch, the Saurashtra coast, the Gulf of Khambat, and the South Gujarat coast, based on coastal geomorphic characteristics.

The Rann of Kutch remains saline desert for the larger part of the year and is further divided into the Great Rann and the Little Rann. On the west of the Great Rann of Kutch is the area of the lower Indus deltaic plain which is characterized

by tidal creeks and mangroves. The coastline in the Gulf of Kutch has extensive mudflats and is highly indented with a number of cliffed rocky islands (Baba and Thomas 1999). Migration, joining of different creeks, reorientation of tidal current ridges, and regression of the sea are seen and are related to tectonically activated lineaments. The coast here is fringed by coral reefs and mangroves. Algae, salt marsh, dunes, and salt pans are very common. The Saurashtra coast has numerous cliffs, islands, tidal flats, estuaries, embayments, sandy beaches, dunes, spits, bars, bays, marshes, and raised beaches at some places. The coast, in Gulf of Khambat is indented by estuaries and consists of mudflats, dunes, and beaches. Here mudflats are seen at different levels and paleo-mudflats have been related to regression. The south Gujarat coast is relatively uniform and is indented by a series of creeks, estuaries, marshes, and mudflats. The Gujarat coast, from Great Rann to the south Gujarat coast, presents evidence for both emergent and submergent coasts.

The Maharashtra, Goa, and northern Karnataka coasts are characterized by pocket beaches flanked by rocky cliffs, estuaries, bays, and at some places mangroves. Beaches in southern Goa and some places along northern Karnataka, however, are long and linear in nature with sand dunes. The Mandovi and Zuari estuarine system in Goa is the largest in this part of the coast. Mudflats are found mainly along estuaries and creeks. Rocky promontories on the Maharashtra coast are made up of Deccan basalts whereas in the south they are mainly of granite gneisses. A number of raised platforms can be seen all along the coast. There are a few islands along the southern parts of this coast near Karwar. This coastal stretch is typical of a cliffy coastline with raised platforms and strong evidence of a submergent coast. The beaches in Goa and northern Karnataka are well-studied and classified as stable beaches with seasonal morphological changes and annual cyclicality (Nayak 1993).

The southern coast of Karnataka is characterized by long linear beaches, estuaries, spits, mudflats, shallow lagoons, islands, and a few patches of mangroves. Satellite image studies revealed northward shifting of the mouth of estuaries along this coast (SAC 1992). Beach erosion is severe in some areas along this stretch. The Kerala coast is known for the presence of laterite cliffs, rocky promontories, offshore stacks, long beaches, dunes, estuaries, lagoons, spits, and bars. Using Landsat images, three sets of sand dunes have been identified. The mud banks are unique transient nearshore features appearing during monsoons (Mathew and Baba 1995) at Kerala. They are unique phenomenon occurring at particular locations along the Kerala coast during the southwest monsoon season, which act as natural barriers to coastal erosion. Along the coast, sand ridges, extensive lagoons, and barrier islands (700 landlocked islands) are indicative of a dynamic coast. About 420 km of the 570 km coastline is protected by seawalls and about 30 km of the coast is

undergoing severe erosion. Maximum loss of material has been reported along the southern sections. The predominant southwest wave approach during monsoons, result in northerly littoral drift with varying speed. Some parts of the Kerala coast are known for rich heavy-mineral deposits. The characteristic coastal geomorphology provides an ecosystem, which supports both agriculture and fisheries. Evidences of both emergent and submergent coasts are available for the southern Karnataka and Kerala coast.

East Coast of India

The deltaic systems of the east coast experience the high sedimentation rate and periodic cyclones which result in extensive floods.

The east coast in the south, along Tamil Nadu and Pondicherry, is straight and narrow except for indentations at Vidyaranya. The major landform along this coast is the presence of a large delta formed due to the Cauvery River and its tributary system. The other landforms are mudflats, beaches, spits, coastal dunes, rock outcrops, salt pans and strand features. At a few places mangrove systems, and at Gulf of Mannar and Rameshwaram fringing and patchy reefs, are seen. Deposition and erosion have been reported at different beaches along this stretch. Rich heavy-mineral deposits have been reported at Muttam–Manavalakuruchi.

The coastline of Andhra Pradesh, mainly the deltaic coast, is 640 km long and comprised of bays, creeks, extensive tidal mudflats, spits, bars, mangrove swamps, marshes, ridges, and coastal alluvial plains. Inundations are seen in the extreme south of the Andhra Pradesh coast, that is, in the saltwater lagoon of Pulicate lake and also between the Godavari and Krishna deltas. The Kolleru lake, situated in the interdelta, formed due to coalescence of the deltaic deposits of the rivers and later it cut off from the sea (SAC 1992), is shrinking on the northern side. The deltaic and southern coasts are rich in agriculture and aquaculture production. The deltaic coast is well vegetated with mangroves. The Pulicate lake has extensive tidal flats and a 12 km long spit. In the north, residual hills and ridges are seen close to the sea. Rocky outcrops and bay beaches are seen here. Storm wave platforms, sea caves in rocks, and cliffs are common coastal features in the north. A critical examination of the relief chart off the region around the Krishna River confluence has indicated the presence of extensive banks in the shelf zone (Varadarajulu et al. 1985). The islands of the Krishna delta front are intertidal and submerged to a large extent during spring tide. The Krishna delta front has been growing through spits and barrier bars (SAC 1992).

The Orissa coast is a site of deposition formed and controlled by the Mahanadi and Brahmani–Baitarani deltas. Mudflats, spits, bars, beach ridges, creeks, estuaries, lagoons, flood plains, paleo-mudflats, coastal dunes, salt pans, and paleo-channels are observed all along the Orissa coast. The

Chilka lagoon is the largest natural water body of the Indian coast. The inlet mouth of Chilika lake is exposed to high annual northward littoral drift and observed to migrate about 500 m northward per year (Chandramohan et al. 1993). The width of beaches at Orissa vary. Littoral transport of sediments in the coastal region is a strong process. The coast is also exposed to severe cyclones. Turbidity in the nearshore as well as in the estuarine region is very high. Progradation of the coastal region in the north of the Devi estuary, and drifting of beaches has been observed. The Bhitarkanika and Hatmundia mangrove reserves are as extensive as 190 km². Gopalpur is rich in heavy minerals. Prominent and well-developed sand dune deposits containing monazite, zircon, rutile, ilmenite, and sillimanite occur along the southern coast of Orissa.

The West Bengal coast represents a typical deltaic strip with almost a flat terrain. The Hooghly and its distributaries form the conspicuous drainage system and forms an estuarine delta. The major geomorphic features are mudflats, bars, shoals, beach ridges, estuaries, a network of creeks, paleo-mudflats, coastal dunes, islands like sagar and salt pans. The Sundarbans, one of the largest single block of halophytic mangrove systems about 1,430 km², of the world need a special mention.

Sri Lanka

Sri Lanka has a coastline of about 1,700 km including that of the Jaffna lagoon. It is a tectonically stable tropical island consisting mainly of Precambrian rocks, and in the northwest, Miocene limestones, and Quaternary sediments. The central portion of the island is a highland surrounded by lowland coasts. Two-thirds of the island's coastline consists of sandy beaches bounded by Precambrian headlands (Swan 1979). The remaining one-third of the coastline, in the northwest and north, consists of sedimentary rocks. Beach material is predominantly terrigenous. Coastal dunes occur along some sections depending on prevailing energy conditions.

The continental shelf between the Gulf of Mannar and Pak Strait in the northwest and north, respectively, is considerably wide across to India. Elsewhere it is narrow.

The coastline of the island is affected by northeast and southwest monsoons. Wave energy is relatively low in the north and northwest because of shallow seas and barriers. In general, however, beaches are open to seasonal strong wave action. In the north and northwest where energy is low, sheltered lagoons with mangroves, estuaries, barrier beaches, spits, and tidal flats are common. Corals forming fringing and small barrier reefs are also seen. Beaches here are narrow and composed of coarse calcareous material. In the Pak Bay and Gulf of Mannar many depositional morphological units are seen. They include, intertidal barriers, multiple sand

bars, dunes, and in the southern part of the Gulf of Mannar a stable sand spit growing toward the northwest. Net sediment movement toward the north along the west coast causes this spit to maintain a stable sand body. From this spit to Colombo in the South, important morphological features that are seen are relict beaches, sand dunes, flood plains, deltas, lagoons, and swamps backed by raised beaches. Dune deposits here overlie limestone. From Colombo, further to the south, lateritized Precambrian rocks form promontories. A sandstone reef offshore, opposite Colombo, acts as a barrier to incoming large waves and the supply of sand material. Raised beaches up to 8 m above sea level are seen at Colombo. Further south the coastline is smooth and sandy, with bays and headlands, backed by raised beaches, flood plains, swamps, and laterite terraces. Along the southwest coast, wave energy is high and sand supply is poor and therefore the coast is undergoing severe coastal erosion. Yun-Caixing (1989) studied coastal erosion and protection using remotely sensed data between Colombo and the southernmost point of the island. The southernmost part of the island consists of low platforms of resistant granitic rocks. The coast here is indented and morphologic units seen are promontories, cliffs, barrier beaches, lagoons, and swamps.

The east coast adjacent to the southern tip, consists of wide coastal plains and low coastlands. Headlands are spaced far apart, and behind long barrier beaches are lagoons, and estuarine deltas. Sand-rich rivers traverse this sector. This change in coastal morphology is in response to a change in geological structure (Cooray 1967). Further north along the east coast, there are two linear submarine structures, namely the Great and the Little Basses (reef) ridges. These ridges are composed of calcareous sandstone (Throckmorton 1964). The landforms of bedrock and sand dunes are replaced by broad flood plains, river terraces, and lagoons, further north small barrier beaches are present. A series of large lagoons which are interconnected are called Batticaloa lagoon, a major feature along east coast. Further north, the coastline is made up of bays and headlands of coral, backed by beach ridges and lagoons. Estuaries, deltas, lagoons, and bay-head barrier beaches are common features along the coast. Along the northeast, bays and headlands backed by raised beaches, lagoons, and low residual rises are the common morphological features. Old beach deposits and dunes are seen, which are rich in ilmenite and rutile minerals.

Bangladesh

The coastline of Bangladesh is around 654 km long from the Indian border in the west to the Myanmar border in the east. This excludes tidal channels and delta estuaries. If estuaries, islands, and tidal channels are included it is more than 1,320 km long. The Bangladesh coast is divided into four

parts from west to east; Sundarbans, cleared Sundarbans, Meghna, and Chittagong. Except for the last one Chittagong, the coastline is low, swampy, and rapidly changing and composed of sediments of the Quaternary period in large alluvial basins. The source is from two vast river systems, the Ganges and the Brahmaputra.

The Sundarbans are thick mangrove and nipa palms swamps, with a total distance of about 280 km (about 195 km in Bangladesh) from the Hooghly River in India to the Tetulia River in Bangladesh. About 68 km long, sundarban forests have been cut and destroyed. This area, is presently, being used for extensive farming. Tidal estuaries, flat marshy islands, creeks and channels, banks of soft muds and clays with thick mangrove and nipa palms are characteristic features of the Sundarban coast. It represents the older deltaic plain of the Ganges with the presence of old beach ridges in the western swamps. The Meghna is a single main channel which after collecting water and material from the Ganges, Brahmaputra, and Meghna rivers, joins the Bay of Bengal. The characteristic feature is the series of extensive shoals called the Meghna flats developed at the mouth of the River Meghna. These shoals are barren mud and sand bodies. This strongly supports a drowned coastal region.

The geomorphic history of the deltaic plain, which includes Sundarbans and Meghna, is continuous shifting of the river course. In recent times, the Ganges has shifted to the east, resulting in the Meghna as a major course. The shifting is explained as tectonic by Morgan and McIntire (1956). Sediment supply and tectonic history at the delta with reference to the last glacial period is explained by Chowdhury (1996).

The Chittagong coast extends 274 km between two rivers, namely River Feni in the north and River Naf on the Myanmar border. Small beaches and broad sand flats between headlands along this coast are the common features. There are many islands and shoals found along this stretch of the coast.

Myanmar

The Myanmar coast is about 2,300 km long from the Bangladesh border to the border of Thailand. The coast is divided into three parts namely the Arakan, Irrawaddy, and the Tenasserim.

The Arakan coast runs parallel to a mountain chain of strongly folded Mesozoic and Tertiary rock. Near the Bangladesh border, the coast is elongated with steep-sided rocks and islands, but further south the coast consists of estuarine channels, mangrove forests, patchy coral reefs, and islands. The coast is an example of an emerged coast with many raised beaches and old sea cliffs. Another significant feature of this coast is the presence of mud volcanoes

which form temporary islands. With wave action, coming in slowly, they transform to shoals.

The Irrawaddy delta coast runs west to east, and is a large delta with deposition of silt and sand. The delta features a number of shoals, estuarine distributaries, channels, and mangrove forests. From the delta region a large volume of sediment is shifted to the east to the Gulf of Martaban by southwest waves during monsoons.

The Tenasserim coast is composed of rocky promontories, valleys, estuaries, mangrove-fringed creeks, and sand spits. Estuarine lagoons and bays are silted up and transformed into mangrove swamps and saline marshy lands. Beaches are rich in heavy minerals, namely ilmenite and monazite. Some beaches are also backed by coastal sand dunes.

Cross-References

- [Barrier Island Landforms](#)
- [Coral Reef Coasts](#)
- [Desert Coasts](#)
- [Indian Ocean Coasts, Coastal Ecology](#)
- [Indian Ocean Islands, Coastal Ecology and Geomorphology](#)
- [Mangroves, Geomorphology](#)

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Indian Ocean Islands, Coastal Ecology and Geomorphology

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Geographically, the Indian Ocean islands (Fig. 1) range from oceanic to continental, geologically from volcanic, limestone, granite, metamorphic to mixed, and physiographically from low to high. Most of these types of islands, though, are not sharply separated.

Oceanic islands are those considered never to have been part of, or connected with, any continental landmass. Their biota is commonly poor in diversity, with unbalanced or uneven representation of taxa, compared with those of continents or continental landmasses. Chagos archipelago, Diego Garcia, and Cocos-Keeling are some such examples. Continental islands may vary in dimensions from subcontinental sizes down to small rocky outposts, the essential characteristics being their continental type rocks and their history showing a former land connection to an adjacent continent. Madagascar, the Malay archipelago, Seychelles, Sri Lanka, and Indonesian islands are examples of continental islands. Seychelles in the north-west Indian Ocean is an extreme case that is totally isolated today, but in Mesozoic and possibly early Tertiary time it was connected to Madagascar. Volcanic islands are rather small (1–100 km across) but often very high, ranging in elevation from 500 m to 3,000 m. They occur generally in irregular clusters, in sub-rectangular patterns or in long lines. Coral islands appear either as an accumulation of coral sand and gravel on the surface of coral reefs or as a slightly emerged limestone platform of formerly live coral not more than a few meters above mean low water. Barrier Islands are constructed entirely by the terrigenous or bioclastic sands from barrier beaches and are built up by longshore drift, probably first as offshore bars, and gradually gaining size later by eustatic oscillations, dune building, and colonization by vegetation.

Because of the very high number of the islands within some island groups in the Indian Ocean it would be difficult to describe them all. Instead, salient features of major groups are given below (see also Table 1).

Western Indian Ocean

Gulf of Kachchh Islands

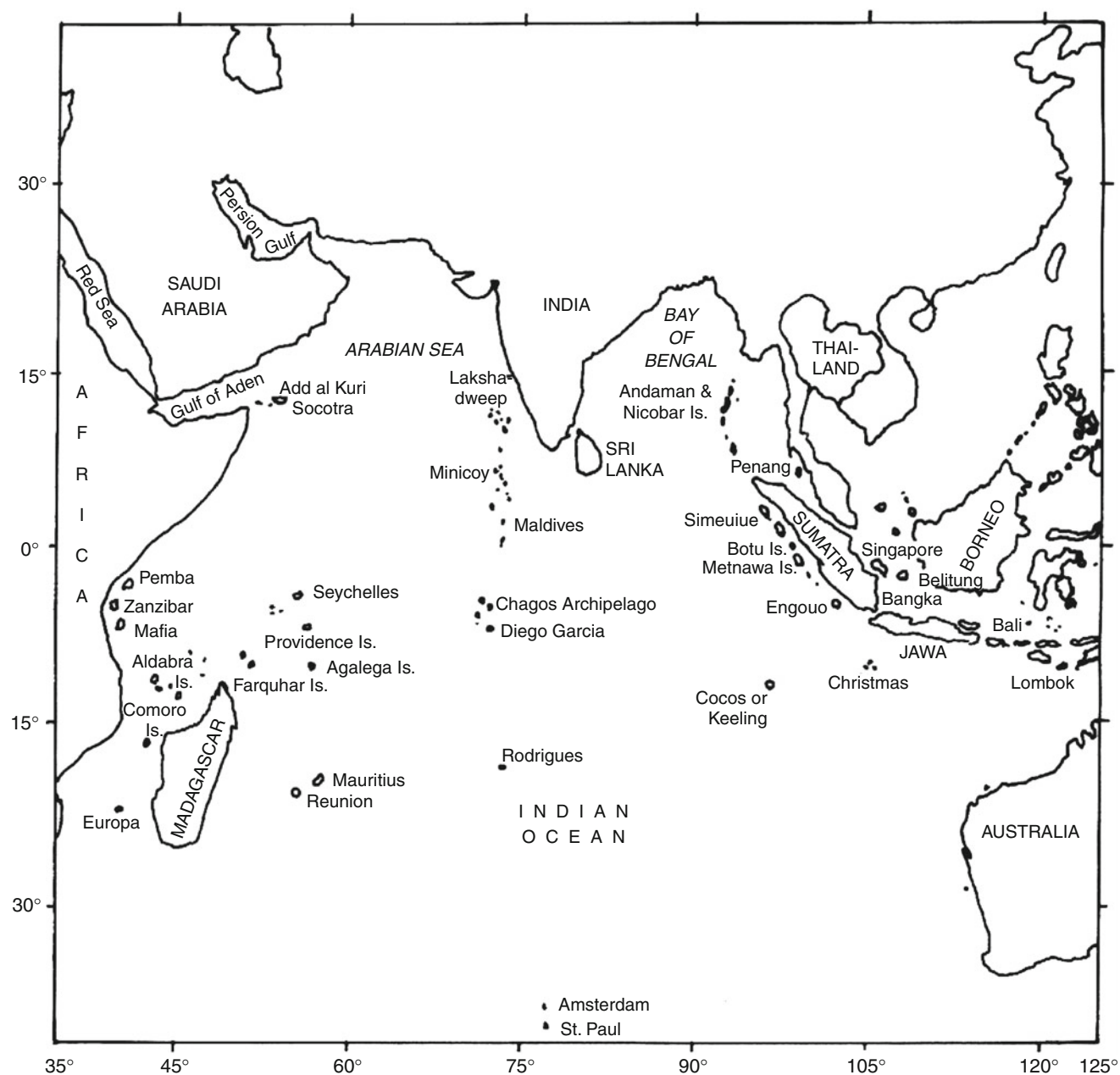
The 42 islands of the Gulf of Kachchh ($22^{\circ}15'N$ – $23^{\circ}40'N$; $68^{\circ}20'E$ – $70^{\circ}40'E$) are the northernmost coralline or sandstone based islands in India. Almost uninhabited, the vegetation inland consists only of shrubs. Several of the islands have dense mangrove patches on the coast, 34 islands have fringing reefs (often called as patch reefs) confined to intertidal sandstones or wave-cut, eroded, shallow banks. The region is tectonically unstable and evidence of uplift can be seen in the form of raised reefs near the mouth of the Gulf, not far from extant islands.

The coastal geomorphology and the fauna and flora of the islands are influenced considerably by the sediment depositional regime, high velocity tidal currents (up to 5 knots), and a large range in environmental parameters (e.g., temperature 15° – $30^{\circ}C$, salinity 25–40). The extreme conditions also limit coastal biodiversity to 37 species of corals and a smaller number of other invertebrates. However, algal growth along these coasts can be substantial at certain times of the year. The mangroves already constrained by high salinity and high tidal exposure also have been heavily impacted due to felling for fuel and fodder. Areas around some of the islands have earlier been good pearl oyster and chank fishing grounds, and one of the islands is even called Chank island. However, overexploitation has decimated both these fisheries.

Laccadive: Chagos Ridge

Lakshadweep islands. These are the northern-most islands of the Laccadive–Chagos ridge (9° – $12^{\circ}N$; 72° – $74^{\circ}E$). Located about 200–400 km off the southwest coast of India, this part of the ridge comprises of 12 atolls, 3 reefs, and 5 submerged banks. Of the 36 islands on the atolls, only 10 (Minicoy, Kalpeni, Andrott, Agatti, Kavaratti, Amini, Kadamat, Chetlat, Kiltan, and Bitra) are inhabited. The northernmost Bitra Island is the smallest inhabited island in India. Among these, Minicoy is separated from the rest by the 9° channel. It is culturally and linguistically closer to the Maldivian islands.

Basically coralline, and no more than a few square kilometers in area, all these islands are low-lying with profuse coral growth all around. The only cultivated plant is coconut, besides a few vegetable and horticultural plants introduced from the mainland. The coast toward the lagoon is sandy and habited by sand dune flora *Spinifex* and *Ipomea*. The seaward shore is rocky and typical of all oceanic atolls, with a steep drop in profile. Radiocarbon dating of the storm beach



Indian Ocean Islands, Coastal Ecology and Geomorphology, Fig. 1 Major islands of the Indian Ocean

at Kavaratti island gave an age of about 6,000 BP indicating their recent origin. Some of the uninhabited islands are only sand cays; one of them is an important nesting ground for seabirds.

The littoral and sublittoral fauna and flora have been studied reasonably well. The known biodiversity status is as follows: hard corals – 104 species, soft corals – 37 species, fishes – 163 species, invertebrates – about 2,000, and algae – 119 species. The bleaching event of 1998 has, however, caused a serious reduction in coral biodiversity. Shore

erosion and silting are additional causes for loss of coral cover and reduction in species abundance.

Maldivian islands. The double chain of Maldivian islands (7°N – 0.5°S , 73°E) is the largest part of the Laccadive-Chagos ridge that extends southwards from India to the center of the Indian Ocean. The more than 1,200 islands, clustered in 19 groups of atolls, are entirely low-lying. The geologic history of the island chain is a complicated picture of sea-level changes, reef and carbonate platform development, and erosional events.

Indian Ocean Islands, Coastal Ecology and Geomorphology, Table 1 Indian Ocean Islands

Name of island	Position	Location	Geomorphology	Major disturbance
Barrow islands (Middle Boodie, Pasco & Double islands)	20°40'–20°58'S, 115°18'–115°30'E	West Australia (56 km off Pilbara coast)	Fossiliferous Miocene limestone	Oil fields
Christmas island	10°35'S, 105°35'E	290 km south of Indonesian islands	Guano deposits	Anchorage and phosphate loading dock
Cocos-Keeling islands	12°00'S, 96°56'E	South east Indian Ocean	Atoll	Storms, earthquakes, red tides
Dampier archipelago (six main and many small islands)	20°20'S–20°50'S, 116°20'E–117°10'E	Eastern Indian Ocean (northwest Australia)	Igneous rock of Archean age	Dredging for shipping
Houtman Abrolhos islands (Total 35 islands)	28°15'–29°00'S, 113°30'E–114°05'E	Eastern Indian Ocean		Oilfield
Kimberley coast reefs Broome, Wyndham, Holothuria and Lacepede reefs, Adele	17°58'S, 122°14'E 15°28'S, 128°06'E 16°52'S, 122°08'E 15°31'S, 123°09'E	Northwestern coast of Australia	Steep rocky shores	Iron ore mining
Monte Bello islands and Lowendal islands	20°20'S–20°33'S, 115°27'E–115°37'E	West northwest of Dampier	Vacant Crown islands	Tourism and oil exploration activities Reef damage by <i>Acanthaster planci</i>
Muiron island and Ningaloo island	21°40'S, 114°2'E	Eastern Indian Ocean	Limestone islands. Part barrier and part fringing reefs	Overexploitation of marine resources Petroleum tenements
Rowley shoals Mermaid reefs Clarke reef Imperieuse reef	17°7'S, 119°36'E 17°10'S, 119°20'E 17°35'S, 119°56'E	Northwest Australian Shelf	Atoll	Fishing, anchorage damage
Scott reef and Seringapatam reef	14°5'S, 121°51'E 13°40'S, 122°00'E	–	Atolls	Exploitation of molluscs and holothurians
Bahrain Mubarraq, Sitra and other islands	26°N, 50°E	Persian Gulf, Saudi Arabia coast	Limestone and deserts	Oil and gas
Lampi island (30 islands)	10°50'N, 98°10'E	Near Burma	Coral reef	Undisturbed
Moscov island	13°50'N–14°20'N	Burma Sea	Rocky shoreline	Illegal logging, collection of turtle eggs
Chagos archipelago Blenheim reef Diego Garcia	5°12'S, 72°29'E 7°20'S, 72°25'E	East Chagos Archipelago South of Chagos	Atoll Atoll	Undisturbed Military presence, dredging and blasting
Egmont atoll (five islands)	6°40'S, 71°20'E	West of Chagos	Atoll	Undisturbed
Peros Banhos atoll	5°20'S, 71°55'E	Northwest of Chagos	Atoll	Undisturbed
Salomon atoll	5°20'S, 72°15'E	North of Chagos	Atoll	Undisturbed
Speakers bank	5°32'S, 72°25'E	North of Chagos	Submerged atoll	No manmade changes
Victory bank	5°32'S–72°15'E	–	Submerged atoll	Undisturbed
Comoros island	12°S, 44°E	North of Mozambique	Four high islands, fringing reef	Dredging
Mayotte barrier reef	12°30'S, 45°10'E	Southernmost islands of Comoros group	Extinct volcano surrounded by reef	Siltation from erosion, mining of coral rocks
Andaman islands	10°30'N–14°N, 92°E–93°E	Bay of Bengal	Emerged part of mountain chain	Massive siltation due to deforestation

(continued)

Indian Ocean Islands, Coastal Ecology and Geomorphology, Table 1 (continued)

Name of island	Position	Location	Geomorphology	Major disturbance
Nicobar islands	6°30'N–9°30'N, 93°E–94°E	Bay of Bengal	Emerged part of mountain chain	Overexploitation of corals and shells for ornamental use
Lakshadweep islands	9°N–12°N, 72°E–74°E	Arabian Sea	Atolls	Dredging, mining, siltation, and <i>A. planci</i> infestation
Indonesian islands	0°50'–1°25'N, 131°16'E	Irian Jaya	Coral island	
Kepulauan Aru	1°1'N–134°10'E	Irian Jaya	Coral island	
Pulau Mapia	0°25'N–130°08'E	Irian Jaya	Coral island	
Raja Ampat (proposed wildlife reserve)	2°30'N–130°50'E	Irian Jaya	Coral island	
Sabuda-Tataruga				
Karimunjawa(proposed wildlife reserve)	5°50'N–110°20'E	Java	Coral island	
Pulau Seribu (strict nature reserve)	5°26'–5°37'S, 106°24'–106°37'E	Java	Coral island	Fishing activities
Karimata	1°30'S, 109°00'E	Java Kalimantan	Coral island	
Pulau Maratua(Karang Muarasproposed strictnature reserve)	1°50'–2°10'N, 118°30'–118°55'E	Java Kalimantan	Coral island	
Pulau Sangalaki(Marine park)	2°05'N–118°15'E	Java Kalimantan	Coral island	
Pulau Semama(wildlife reserve)	2°05'N–118°15'E	Java Kalimantan	Coral island	
Aru Tenggara(proposed marinereserve/ marine park)	6°35'S – 7°11'S, 134°12' – 135°E	Moluccas	Coral island	Fishing and exploitation of marine resources
Kepulauan Kai BaratTayandu (proposedmarine multiple use reserve)	5°30'S, 133°0'E	Moluccas	Coral island	
Pulau Angwamase	7°55'S, 131°20'E	Moluccas	Coral island	
Pulau Banda(marine park)	4°30'S, 130°E	Moluccas	Volcanic origin	
Pulau Pombo(marine park)	3°31'S, 128°22'E	Moluccas	Coral island	Overexploitation
Pulau Renyu-PulauLucipara (proposedstrict nature reserve)	5°40'S, 127°50'E	Moluccas	Coral island	Overexploitation
Komodo National Park	8°35'S, 119°30'E	Nusa Tenggara	Rocky island	Dynamite fishing
Pulau Rakit	8°35'S, 118°E	Nusa Tenggara	Coral island	
Pulau Satonda	8°10'S, 117°45'E	Nusa Tenggara	Volcanic	
Kepulauan Peleng	1°50'S, 123°14'E	Sulawesi	An archipelago	
Kepulauan Sangihe	3°45'S, 126°35'E	Sulawesi	Islands and reefs	
Pulau Kakabia	6°55'S, 122°30'E	Sulawesi	Islands and reefs	
Pulau Pasoso	0°10'S, 119°45'E	Sulawesi	Limestone island	
Pulau Smalona	5°10'S, 119°10'E	Sulawesi	Coral islands	
Selat Muna	5°05'S, 122°05'E	Sulawesi	Coral islands	
Spermonde islands	5°30'S, 119°10'E	Sulawesi	Coral islands submerged reef, patch reefs	Dynamite fishing, sedimentation
Tiga island	3°50'S, 123°15'E	Sulawesi	Islands and reefs	
Togian island	0°10'N–0°40'S, 121°32'E–122°12'E	Sulawesi	Extended mountains surrounded by limestone	Dynamite fishing

Tukang Besi	5°35'S, 123°50'E	Sulawesi	Archipelago		
Muara Siberut(five islands)	1°30'S, 99°25'E	Sumatra	Islands with coral reefs		Overfishing
Pulau Weh	6°0'S, 95°30'E	Sumatra	Coral reefs		
Sheedvar island	27°0'N, 53°E	Iran	Coral reefs		
Pulau Paya (group of islands)	6°02'N–6°05'N, 99°54'–100°04'E	Malaysia	Coral islands		Fishing activities
Pulau perak	4°41'N, 98°56'E	Malaysia	Conical rocky islands		
Pulau Tiga	5°44'N, 115°40'E	Malaysia	Volcanic island		Legal protection and management
Sembilan islands	2°03'N, 100°33'E	Malaysia	Volcanic island		//////
Pulau Bohey dulongPulau BodgayaPulau Tetagan	4°38'N, 118°46'E	Malaysia	Volcanic island		Domestic pollution, overexploitation of marine fauna
Pulau Sipadan	4°05'N, 118°40'E	Malaysia	Coral reef		
Pulau gayaPulau SapiPulau MamutikPulau ManukanPulau Sulung	6°04'N–5°55'N, 116°E	Malaysia	White sandy beaches with rocky interruption		Coral mining, overexploitation of reef fauna
Maldives (a group of atolls)	7°N–0°30'S, 73°E	Central section of Chagos archipelago	Atolls		Growing urbanization, domestic pollution
Mauritius	20°S, 58°E	180 km northeast of Réunion island	Remains of old emerged coral reefs		
Cargados Carajoshoals	16°23'S, 59°27'E	North northeast to Mauritius	Coral islands		Exploitation of marine products, Guano mining
Île Plate	19°53'S, 57°39'E	–	Volcanic rock		Dynamiting, fishing, anchoring, and boat grounding
Rodrigues	19°42'S, 63°25'E	North eastern Mascarene island	Volcanic island		Public interference
Mozambique Ilhas da Inhaca e dosPortugueses(Inhaca islands)	26°S, 33°E	Southern most island of Mozambique	Pleistocene dune rock–		Goat overgrazing denudation, public interference
Primeira and Segundoislands	16°S–17°S, 38°E–41°E		–		
Quirimba islands	10°45'S–12°42'S, 41°E		–		
Réunion islands					
Réunion	21°7'S, 53°32'E				
Europa and Basses delndia	22°20'S, 40°20'E	In Mozambiquechannel	Atolls		Undisturbed
Iles Glorieuses	11°30'S, 47°20'E	100 km northwest of Europa	Atolls		Undisturbed
Tromelin	15°52'S, 54°5'E	390 km east of Madagascar	Atolls		Undisturbed
Saudi Arabia					
Al Wajhto Qalib	26°16'N, 36°28'E	Northern Red Sea	Rocky island		Undisturbed
Farasan archipelago	17°40'N, 42°10'E	Red Sea	Coral islands, swamps, and reefs		Undisturbed
Seychelles	5–10°S, 45–56°E	NE of Madagascar			
Platte island	5°50'S, 55°E		Granitic and coralline islands		–
Doivre island	5°50'S, 55°E				

(continued)

Indian Ocean Islands, Coastal Ecology and Geomorphology, Table 1 (continued)

Name of island	Position	Location	Geomorphology	Major disturbance
Aldabra island	9°25'S, 46°25'E	North of Mozambique channel	Atoll, limestone	–
Bird island	3°43'S, 55°13'E	Northern edge of Seychelles	Sandy clay and calcareous	–
Dennis island	3°48'S, 55°40'E	–	–	Phosphate mining
Cousin islands	4°20'S, 55°40'E	Southwest Seychelles	Granite island	Tourism
Curieuse island	4°05'S, 55°43'E	–	–	–
Singapore	1°20'N, 103°50'E	–	–	Growing urbanization
Pulau Hantu	–	–	–	–
Pulau Sulong	–	–	–	–
Pulau Salu	–	–	–	–
Sri Lanka	8°N, 82°E	16 km from Trincomalee	Extended part of Indian mainland (limestone-granite)	Mining for lime, dynamite fishing, tourist pressure, overexploitation of marine fauna
Sudan				
Mukkarwar island	20°50'N, 37°17'E	North of port Sudan	Well-formed reefs	–
Sanganab atoll	19°45'N, 37°25'E	Red Sea	Atoll	<i>A. planci</i> infestation, overexploitation Fishing, tourism pollution, and human interference
Suakin archipelago	19°14'N, 37°51'E	–	–	–
Sultanate of Oman				
Daymaniyat islands (10 islands)	23°45'N, 58°10'E	Gulf of Oman	Corals	Litter, <i>A. planci</i> infestation
Tanzania				
Latham island	6°50'S, 39°50'E	64 km off Dar es Salaam	Atoll	Undisturbed
Mafia island	7°40'S, 40°40'E	South of Dar es Salaam	–	–
Maziwe island	5°30'S, 39°5'E	Northern Tanzania	–	–
Thailand				
Koh Lam	12°55'N, 100°47'E	Northern Gulf of Thailand	Coral reefs	Tourism and pollution
Koh Sak	–	–	–	–
Koh Krok	–	–	–	–
Koh Phuket	8°N, 98°20'E	Andaman Sea	Rocky extended mountains	Extensive sedimentation, dredging, mining

As would be expected in the case of small islands on the atolls, the coastal ecology of the islands is reef-dominated. The reefs, though principally atolls, have the unusual features of broken rims that consist of numerous patches or faroes, many of them with islands, and the presence of lagoonal islands which are simply knolls with their emergent surfaces capped with vegetation.

Biological and ecological information on Maldivian islands is rather poor, with stress on only some groups. About 200 species of corals under 60 genera have been recorded so far. No comprehensive checklist of other groups of coastal marine fauna exists; however, the cowry shells and groupers (40 species) are important components of reef biodiversity. Similarly, descriptions of zonations of the reefs are known from some islands but detailed studies of the ecology, either at community or at species level, are scarce.

Islands of Chagos archipelago. The Chagos archipelago (5°–8°S, 71°–73°E), the southern part of the Laccadive–Chagos ridge, consists of five coral atolls with islands, besides several reefs that are partially or wholly exposed at low tide. The five atolls are: Great Chagos Bank, Peros Banhos, Salomon, Egmont, and Diego Garcia. The number of islands on these atolls varies from 4 in Diego Garcia, to 24 in Peros Banhos. The total land occupied by these islands is about 40 km².

Most of these islands are located on the atoll rims with elevation of no more than 2–3 m. Raised reefs with small, uplifted, and vertical cliffs rising to over 6 m occur in two atolls – southern Peros Banhos and northwestern part of the Great Chagos Bank. Isotopic dating of fossil corals in the emerged beach rock of the islands and some extant corals gave an age not more than 5,200 years BP, indicating that all these islands are relatively recent in origin.

Being coralline islands, the coastal morphology is typically characterized by vast and profuse reef growth. Mangroves and associated flora are absent. The coastal and inland flora consists primarily of native vegetation (*Tournefortia*, *Scaevola*, and *Casuarina*), disturbed by coconut plantations in inhabited islands. None of the 250 species of flora are endemic. Two faunal groups – birds and turtles – are important biological components. The islands provide nesting grounds for over 50 species of seabirds and several species of green and hawksbill turtles.

Seychelles

A total of about 42 granitic and 74 coralline islands, spread over 5°–10°S and 45°–56°E, with a total land area of 455 km² comprise the Seychelles. The inner Seychelles islands to the north are granitic, remnants of ancient Gondwanaland, rugged, mountainous, and rise up to 1000 m in the Morna Seychellois on Mahé island. Coralline islands and atolls of the Amirantes, Farquhar, and Aldabra groups, spreading westwards and southwards from the granitic group, constitute the

second group. They are composed of numerous low islands or atolls in several clusters, each located on top of volcanic structures of various sizes. Principal granitic islands in the Seychelles archipelago are Mahé, Praslin, Silhouette, La Digue, Curieuse, Felicité, North Island, St. Anne, Providence, Frigate, Denis, Cerf, and Sea Cow island. Among the coralline islands and atolls, Alphonse, Bijoutier, St. François, St. Pierre, Astove, Assumption, Coetivy, and Aldabra are the major ones. Some of the coralline islands, though relatively low-lying by comparison with granitic islands, are often taller, reaching as much as 8 m above sea level. These include Aldabra, Assumption, Astove, Cosmoledo, and St. Pierre. These limestone atolls have formed on top of volcanic structures and rise from water depths of over 2,000 m. The high limestone islands are also characterized by terraces that reflect changes in sea levels during the last glacial cycle. The Amirantes island group is the second largest group after Inner Seychelles and comprises 10 islands and atolls and several shoals and submerged reefs. The major atolls in the Farquhar (or Providence) group are Farquhar and Providence, each rimmed with several islands. St. Pierre Island in this group is a circular, uplifted atoll, with coastal cliffs rising to 10 m. In the Aldabra group, Aldabra and Cosmoledo atolls have several islands on their rims, whereas Astova and Assumption are elevated atolls: in the Assumption island, reef rocks rise to 7 m above sea level, with dunes on the east and south rising to nearly 30 m.

Coastal geomorphology of all the islands is characterized by the presence of reefs. Three major types – fringing, platform, and atoll reefs – are recognized. The granitic islands have well-developed fringing reefs. Among the outer islands, several are raised platform reefs (e.g., Assumption and St. Pierre) and others are atolls (e.g., St. Joseph, St. François).

There are pronounced differences in the coastal ecology of these islands. The northern islands lie in the path of the east-flowing Equatorial Counter Current whereas the southern islands lie in the path of the west-flowing South Equatorial Current. Besides, the granitic islands receive more rainfall and are often forested. As a result, the nutrient regimes in the coastal waters are distinctly different between these two groups. The high nutrient levels around the granitic islands favor dominance of crustose coralline algae and frondose macroalgae whereas nutrient-poor waters around the coralline islands support a hermatypic coral dominance. The raised reefs were used as nesting sites by seabird colonies and have been extensively mined for guano.

Description of coastal ecology has primarily been with reference to coral reefs, though not all have been studied as extensively as Mahé or Aldabra islands. The coral diversity is more or less same between granitic and outer islands, the former with 51 genera and the latter with 47 genera. The total species count is 161, and this does not include those of *Acropora*. Though the coasts with well-developed reefs can

be expected to sustain a high faunal and floral diversity, there is still poor documentation from most islands. The recorded forms include 128 species of marine caridean shrimps, 49 species of brachyuran decapods, 150 species of echinoderms, 450 species of molluscs, about 1,000 species of fish, 8 species of sea grasses and 4 species of turtles. Among these the gigantic land tortoise of Aldabra atoll and the double coconut, *coco de mer*, of Mahé Island are unique.

Mascarene Islands

Mauritius. Mauritius (20°S, 58°E) represents the southern part of the Mascarene Plateau, which is an arcuate series of banks extending for 2,000 km from the Seychelles Bank. The Mauritius island is volcanic in origin and is composed of olivine basalt and doleritic basalt.

Mauritius, along with several small adjacent islands, spreads over 1,865 km². The northern part of the island is a plain while the center is a plateau rising to a peak height of 826 m at Piton de la Rivière Noire and bordered by low mountain remnants of a large volcano. The south of the island is largely mountainous.

The crenulated coastline, exposed to varying wave activity, extends to about 200 km. The southwest coast is made up of basaltic rocks while carbonate sands, the bulk of which is coral debris, largely cover the remaining parts. A large submarine platform with extensive fringing coral reefs that cover three-fourth of its coastline surrounds the island. There are also the remains of old, emerged coral reefs, recalcified to varying degrees, indicating recent uplift.

Sugarcane and tea cultivation are the major revenue sources for the islands. Conversion of forest lands to plantations has, however, reduced the original forest cover to less than 1%. The other important revenue sources for the island are coastal fishing and tourism. Reef fisheries yield about 200 tons per year and reef tourism caters to more than 300,000 visitors a year.

Considering that the entire coastline is reef-rimmed, the corals- and reef-associated fauna and flora are important components of coastal biodiversity. The reefs cover an area of about 300 km², with a maximum reef width of 4 km. Mauritius reefs are notable for the absence of reef flats; consequently, sediments accumulate in the lagoon providing a favorable environment for sea grass growth. A total of six sea grass species – *Thalassiodendron ciliatum*, *Syringodium isoetifolium*, *Halophila ovalis*, *Halophila stipulacea*, *Halodule universis*, and *Halodule* sp. – are known from Mauritius.

A total of about 186 species of corals belonging to more than 50 genera have been reported from the Mascarene archipelago, with 75% of these recorded in Mauritius reefs. The fish diversity, with 263 species, is also high. The molluscan fauna is another important biological constituent of Mauritius reefs. A detailed survey has revealed the presence of more than 3,500 species, with approximately 10% of them being

endemic. These include the Imperial Harp shell *Harpa costata* and the cowry *Cypraea mauritiana*, besides species like *Clanculus mauritianus* and *Bursa bergeri*. Diversity of marine macro algae is also high: 127 species, mainly red and green algae, have been recorded from the littoral zone.

Higher freshwater flux, more siltation, and high humidity favor the growth of mangroves on the northeast and east coasts of Mauritius. Dense mangroves dominated by *Rhizophora mucronata* cover an estimated area of 20 km².

Rodrigues. Rodrigues island (area 110 km²; 19°42'S, 63°25'E) and two cays at 10°24'S and 56°38'E (known as Agalega island) are part of the Republic of Mauritius. Rodrigues, the smallest of Mascarene islands, is of volcanic origin and consists of subhorizontal basaltic flows. The north-eastern part of the island is mountainous but not very tall, the peak height not exceeding 400 m.

Like Mauritius, Rodrigues island is also reef-rimmed, with a wide expanse of reef platform extending without a break for 90 km around the island, providing a fringing reef cover of about 200 km². Presence of reef flats, with a width ranging from a low of 50 m in the east to as much as 10 km in the west distinguishes Rodrigues from Mauritius. The reef flats also provide habitat for large sea grass beds, though the species diversity is much less; only two species – *H. ovalis* and *Halophila balfouri* are known from Rodrigues. Muddy accumulations, hence development of mangroves, are rare.

Coral species diversity is high, with a similar number of species as in Mauritius. Information on other coastal marine fauna and flora are scarce; however, the small islands around Rodrigues are important nesting sites for brown noddy, lesser noddy and white tern.

Réunion. Réunion (21°7'S, 53°32'E) is the most south-westerly of the Mascarene islands. Covering an area of 2,512 km², Réunion is a large Hawaiian-type volcano that includes an older part, the massif of Piton des Neiges (peak height 3,069 m), incised by three large cirques, and an active volcano to the southeast. Several small islands – Tromelin to the north of Réunion, Europa and Bassas de India atolls in the Mozambique Channel, Juan de Nova off the west coast of Madagascar, and Iles Glorieuses to the north of the Mozambique Channel, are the other island dependencies of Réunion. These are relatively very small. Tromelin is no more than a cay of 1.1 km², Juan de Nova is a raised fossil reef of 9.6 km² and the Grande Glorieuse Island covers only 3.9 km².

The coast of Réunion is generally rocky, with low cliffs cut in lavas. Sectors of low coast correspond with the three large depositional cones built below the three cirques and show pebbly beaches while the rare sandy beaches are related to embryonic fringing reefs. Because of the relatively young age, reefs are less developed, discontinuous, narrow with their widest part no more than 550 m, and cover only an area of 7.3 km². In contrast with Mauritius and Rodrigues, reef platforms are abundant on Réunion but muddy accumulations

and mangroves are totally absent. Lack of organic and terrigenous material also limits the diversity of sea grasses and their extent: only one species, *H. stipulacea*, that too not in abundance, has been known from Réunion.

Coastal marine fauna and flora have been better inventoried in Réunion than in the other two islands. This includes 40 species of dinoflagellates, 150 species of macro algae, 120 species of corals, 90 species of hydroids, 2,500 species of molluscs, and more than 650 species of fish. Coastal fisheries, however, are relatively less developed, with no more than 1,500 tons yield per year, of which only 100–150 tons are truly reef fishes. Tourism, likewise, is also less developed compared with Mauritius.

Madagascar

Madagascar is a fragment of a lost continent (Lemuria) and this severance from the ancient land mass is evident from the sheer drop of mountain into ocean depths of 3,000 m or so, especially on the eastern side. The island's rocks, volcanic structure, besides the subsoil formed of granites, gneiss, and crystalline schists, warm water springs and frequent earthquakes also provide evidence for this origin. Paleontological evidence for this comes from the remains of many large prehistoric birds and even the present day fauna of the island has a special individuality. The large number of endemic flora also confirms that Madagascar is an island that has been long since isolated from other regions. Madagascar is also one of the largest islands in the world (5,87,000 km²) with a coastline of about 4,000 km. The east and west coasts are asymmetrical in physiography. The east coast presents an almost unbroken appearance, with few bays and indentations. The continental shelf here is narrow and coral reefs and mangroves are poorly developed. The west coast, on the other hand, has a broad continental shelf and has the majority of the island's reefs and mangroves.

Reefs and mangroves are important coastal ecosystems of Madagascar. The reefs cover an area of 200 km². Most of the west coast has large tracts of tidal marshes (4,250 km²) of which 3,200 km² are populated by mangroves. Sea grass beds are also extensive on the west coast. Emergent fossil reefs up to 10 m above present sea level are found in the far northwest coast. A barrier reef, 10–16 km offshore, also exists at the edge of the continental shelf.

There is an extensive bibliography of the various coastal and marine fauna and flora, synthesized from numerous studies of many French scientists. Most of the information is from Toliera and Nosy-Bé and biodiversity of the whole island could be still higher than what is known – 200 species of corals, 1,500 species of fishes, 28 species of sponges, 227 species of echinoderms, 1,158 species of mollusks, 779 species of crustaceans, 121 species of worms, 182 species of ascidians, 108 species of algae, besides 5 species of turtles and 32 species of mammals.

Comoros Islands

The Comoros archipelago consists of four major islands – Grande Comore, Anjouan, Moheli and Mayotte – at the northern end of the Mozambique Channel (12°S, 44°E). All these islands are of volcanic origin and mountainous, and are surrounded by numerous coral-line and granitic islets. The Grand Comore is the largest island among these, with an area of 1,131 km². While three of these islands have fringing reefs, Mayotte is surrounded by a 140 km long barrier reef lying 13–15 km offshore. The coastal features include the mangroves, which, in some islands, are expanding due to influx of terrestrial sediments from hillsides.

Among these islands, only Mayotte has been studied to some extent. These studies are essentially related mainly to the description of the barrier reef and its faunal and floral composition, since it is one of the few barrier reefs in the world, and the best developed in the Indian Ocean. The reef, which had a good live coral cover, was heavily impacted during a bleaching event in 1983 related to El Nino.

Islands Off Tanzania

The islands off the coast of Tanzania are Mafia, Pemba, Unguja (Zanzibar), and those of Songo Songo archipelago. Mafia island (7°40'S, 40°40'E), along with the four small adjacent islands, are continental islands off Rufiji delta. The coastline consists of vast stretches of sheltered and exposed fringing reefs and mangroves besides beds of sea grasses, algae, and soft corals. The Unguja island, slightly larger than the Mafia island, as well as the Mnemba island lying to the north, also have coral development all around the coast. The Pemba island is 62 km long and 22 km wide, with a reef area of 1,100 km² along its coastline. All these islands show evidence of several terraces, along with indication of a relatively recent subsidence. Another feature common to these is the 3,300–3,400 m of marine sediments, ranging from Miocene to Cretaceous, underlying them.

The Songo Songo archipelago (8°30'S, 39°30'E) consists of a 7 km long island, with four smaller islands in the vicinity. As with other islands off Tanzania, these islands support some of the largest expanses of shallow water coral reefs, with the estimated reef cover of about 40–50 km². No other remarkable coastal features are known from these islands.

Islands Off Mozambique Coast

The islands off the Mozambique coast are grouped into Quirimbas archipelago, Primeiras and Segundas archipelago, Bazaruto archipelago, and Inhaca and Portuguese islands. Besides these, the Mozambique, Goa, and Cobras islands are located just 4 km off the mainland coast.

The Quirimbas archipelago (10°45'–12°42'S) comprises a 200 km chain of 32 islands along with numerous reef complexes. The Primeiras and Segundas archipelagos,

located at 16°12'–17°17'S consists of 10 islands and two reef complexes. The Bazaruto archipelago consists of five islands located between 21°30', and 22°10'S. All these islands are small, with the largest no more than 25 km² in area and all lie close to the coasts.

Coastal morphology of these islands is composed of grasslands, scrubs, and mangroves, with varying degrees of development in the different islands. All of these, however, support good fringing reef growth. As with most islands, it is the coral fauna and fish that were widely studied. About 50 genera of reef building corals and 300 species of fish are known from Quirimbas archipelago. A total of about 155 molluscan forms, with 6 endemic species among them, have been reported from Bazaruto archipelago. The western coasts and the area between the islands and the mainland coast have good sea grass beds.

Socotra Island

The Socotra (8°N, 53°E), off the mouth of Gulf of Aden, is an island that has survived the subsidence of the great primeval continent, which embraced present-day Africa, the Middle East, southern Asia, and the Northwestern part of the Indian Ocean. It is a fairly large island (3,582 km²) with its mountainous interior rising to 1,520 m. The coastline is varied, consisting partly of low-lying plains and partly of steep limestone cliffs, edging an undulating plateau (500–600 m high) that covers much of the island.

Bahrain

This consists of a group of low-lying islands, largely of limestone outcrop and desert, off the Saudi Arabian coast. Ranging from a rocky out crop (Jidda Island) to the large Bahrain Island (660 km²), these are low and sandy islands, except for clusters of barren rocky hills in the center.

Eastern Indian Ocean

Gulf of Mannar Islands

A chain of 20 islands (8°45'–9°16'N, 79°4'–79°29'E) constitutes the coralline islands of the Gulf of Mannar between southeast India and Sri Lanka. None of these Islands is inhabited. Spreading over not more than 2 km² individually, most of these islands have only shrubs as vegetation and occasionally some patches of mangroves. The fringing reef growth is profuse all around the islands. The sea grass beds associated with the reefs have been important feeding grounds for the Dugong species.

Geologically, these islands are connected with those of northern Sri Lanka through a series of shallow banks called Adam's bridge between Rameswaram in India and Talaimannar in Sri Lanka. As a consequence, there is a good similarity in island geomorphology, coastal ecology, and fauna and flora among these islands.

Sri Lanka

Sri Lanka lies off the southeast tip of India between 6° and 10°N and 80° and 82°E (65,610 km²). The island is basically a central mountain mass of Pre-Cambrian crystalline rocks ringed by a broad coastal plain. The highest point is the Pidurutalagala peak (2,700 m). The plains are fairly level in the north but the extensive soft limestone deposits are broken elsewhere by outcrops of the main rock core. The coastal region is characterized by fringing reefs, shallow lagoons, marshes, and many sandy bars, especially in the north. Though not a very large island, three distinct climatic features are evident: the low country Dry Zone that receives low rainfall, the East Coast Plains that experience one monsoon, and the Low Country Wet Zone which receives rainfall from both the monsoons.

Coastline features vary considerably, from sandstone to granite, but detailed studies are scarce. Along the west coast, coral growth is mainly on ancient sandstone and on the east coast, it is on gneiss or granite outcrops. Fringing reefs are found only along 2% of its 1,585 km coastline and not all of these are comprehensively mapped. Surveys have mainly been carried out for reef-building corals and fishes. A total of about 183 species of corals and over 350 species of fishes have been recorded. Though a number of faunal groups occur in these reefs, their systematic records are not known. Mangroves occur along the southwest, north-west, and northern coasts. The total area covered by mangroves is 36 km². *Rhizophora*, *Avicennia*, *Excoecaria*, *Lumnitzera*, *Aegiceras* and *Sonneratia* are the mangrove genera recorded.

Extensive damage to coastline habitats has been recorded where reefs and mangroves are abundant. The reefs are mined for lime manufacture and the mangrove wood is used to fuel the limekilns. Often coastal forests are also exploited for use as fuel in the limekilns. Though detailed information on the impacts are not available, damages to coastal habitats at local scales are considerable.

Andaman and Nicobar Islands

These are the emerged parts of a mountain chain that stretches from the Arakan Yoma in Myanmar to the islands of Indonesia. Spread meridionally between 6° and 14°N, and between 91° and 94°E in the Bay of Bengal, these islands number more than 500, of which only 38 are inhabited. All the islands are mountainous, sedimentary in nature, and have fringing reefs towards the east. The total area covered by these islands is 8,293 km². The Andaman group of islands is separated from the Nicobar group by the 10° channel which has a heavy tidal flow and difficult to navigate with conventional crafts. As a result, the biogeography of the Andaman has more of Malay affinities whereas that of the Nicobar has Indonesian affinities.

Most parts of these islands are covered with thick forests and the low-lying areas are covered with mangrove swamps. Biodiversity of the islands is quite high, with an abundance of

corals, fishes, algae, turtles, and the unique saltwater crocodile. Avifauna is more endemic in nature, with distinct local species of eagles, parakeets, and orioles. The islands have a few land mammals like deer and elephants, which were introduced from the mainland India.

Indonesian Islands

Indonesia is an island nation of 13,700 islands having a coastline of about 60,000 km. The islands form a region of tectonic instability, marked by frequent earthquakes and volcanic eruptions. These tectonic movements have also shifted out of the sea some of the numerous coralline reef formations along the island coast.

Typically tropical in climate, the larger islands (e.g., Java, Sumatra, Borneo, and Sulewasi) have varied coastal geomorphology ranging from mangrove-bordered shores through estuarine deltas to coral reefs. Borneo is the third largest island in the world and lies between the Sulu Sea, Java Sea, and South China Sea. It is densely forested, with extensive swampy lowlands in the southern and southwestern coastal areas.

Smaller islands are more coralline in nature. The Indonesian region is known for the highest diversity of corals and mangrove species, in the latter case practically all known mangrove species from the new world are present here.

Cocos or Keeling Islands

These islands (12°S, 96°56'E), numbering 27 and covering a total area of 30 km², lie about 960 km southwest of Sumatra (not to be confused with Cocos Island, a small uninhabited island of 26 km² area off Costa Rica). The largest among them is the West Island, with an area of 3.2 km². These are extremely coralline islands but also with a luxuriant growth of land vegetation that includes coconut palms, sugarcane, and banana. No quantitative accounts of other fauna are available but the coconut eating crab *Birgus latro* is an interesting species from these islands.

Christmas Island

Christmas island, with an area of 128 km², and lying 650 km south of Java Head, is a strongly uplifted island surrounded by high cliffs of coral limestone. It is well-known for its extremely rich phosphate deposits and the coconut crab *B. latro* (another Christmas island is an atoll in the Line islands, central Pacific).

In summary, the major features of Indian Ocean islands are:

- Wide range in size, from sand cays to some of the largest islands in the world.
- Coralline origin in most of the oceanic regions and granitic or sedimentary origin in coastal regions.
- Coasts characterized by fringing reefs in almost all cases, and mangroves and other wet lands in most others.
- Endemism in some islands, with native flora and fauna.

- Most of the smaller islands are uninhabited. Where settlements have taken place, marked erosion in biological diversity and resources are noticeable.

Suggested further reading on this subject may be found in the bibliography listing.

Cross-References

- Atolls
- Clifed Coasts
- Coral Reef Coasts
- Coral Reef Islands
- Coral Reefs
- Coral Reefs, Emerged
- Indian Ocean Coasts, Coastal Ecology
- Indian Ocean Coasts, Coastal Geomorphology
- Mangroves, Ecology
- Mangroves, Geomorphology
- Mining of Coastal Materials
- Rock Coast Processes
- Small Islands
- Tors

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Ingression, Regression, and Transgression

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Definition

A transgression is a landward shift of the coastline while regression is a seaward shift. The terms are applied generally

to gradual changes in coast line position without regard to the mechanism causing the change. In addition, these terms usually are applied to changes over periods greater than 100 years as can be expected to be recorded by facies distributions in the geologic record or the stratigraphic interpretation of seismic reflection. “Transgressions” and “regressions” are commonly used, for example, to refer to coast line changes due to glaciations, which cause both eustatic sea-level changes and subsidence or rebound. Of particular significance is the “Holocene transgression” which corresponds to a eustatic rise in sea level of between 100 and 130 m and between 18,000 and 6,000 year bp. The terms also have been applied, however, to changes occurring over shorter time scales in, for example, Lake Chad.

“Ingression” refers to the advance of marine conditions into more-or-less confined areas, like the drowning of a river valley (Schieferdecker 1959, terms 1260 and 1840, as cited in Jackson 1997) or to the infiltration of water to an interior low-lying area of land creating an inland body of water. In the former application, at least, neither mechanism nor time period is implied, although the tendency seems to be to use “ingression” to refer to more rapid, if not catastrophic, transitions rather than to more gradual changes. An example might be the marine invasion of a glacial lake by breaching of a morainal barrier during post-glacial, sea-level rise. A rise in relative sea level will also raise the water table in coastal aquifers causing the appearance of lakes and ponds, but the same could be accomplished by changes in recharge (precipitation-evapotranspiration).

Facies Relationships and Implications

Regressions or transgressions can be due to any combination of (1) eustatic sea-level rise or fall; (2) subsidence or uplift; and (3) sedimentation or erosion. In sedimentology, transgressions are also referred to as “retrogradation” in which a depositional environment due to the rate of creation of accommodation space exceeds the sediment supply (Curry 1964; Nichols 1999). This results in a landward shift of coastal facies belts, although it is possible that the advance of the marine conditions had occurred so quickly that sediment deposition could not remain in equilibrium with changing conditions. In this case, relict, subaerial sediments may be found drowned in place or shallow water deposits covered discontinuously by deep-water facies. Transgressions are usually associated with a rise in relative sea level due to eustatic sea-level rise and/or coastal subsidence. Even during periods of stable sea level or slowly rising sea level, however, erosion of coastal deposits can result in a transgression.

Regressions are also referred to as “progradations” in which the sediment is supplied at a higher rate than the potential space for sediments to be accommodated is created.

In the geologic record, this can be preserved by up-column transitions to distinctly shallow water facies. A regression is usually associated with a falling sea level, but even in the face of a stable or slightly rising local sea level, the shoreline may still be displaced in a seaward direction by the rapid deposition of coastal sediments. Progradation is, for example, associated with delta formation. The relationship between the rate of sea level changes and the rate of deposition or erosion was discussed by Curry (1964).

Onlaps and Offlaps

Global sea level changes are recognized in seismic sequence stratigraphy. Depositional sequence boundaries are identified by lateral terminations of strata and incorporated into a curve of eustatic sea level. Onlaps are overlapping relationships of shallower water sediments over deeper-water sediments, which progressively pinch out toward the margins of a sedimentary basin. “Marine transgressive sequence” is used as a synonym for onlaps. Offlaps occur when progressively younger sediments have been deposited in layers offset seaward, often associated with an upward coarsening.

The Anthropocene defines an epoch characterized by human impacts. On the appropriate time scale for transgressions, future forecasts of sea-level rise continue to be developed under a range of climate change scenarios (Horton et al. 2014). Curry, accordingly, came to refer to the occurrence of a shift in the shoreline independent of the evidence found in the geologic record for these changes.

Further suggested readings are also included in the following bibliography.

Cross-References

- [Changing Sea Levels](#)
- [Coastal Sedimentary Facies](#)
- [Sea-Level Indicators, Geomorphic](#)
- [Sequence Stratigraphy](#)

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Isostasy

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Numerous observations point to a complex and changing relationship between land and sea surfaces throughout geological time. In some localities elevated coral reefs, wave-cut rock platforms, and molluscs embedded in their original marine sediments attest to past sea levels having been higher than present. At other sites, drowned forests and submerged sites of human occupation point to sea levels having been locally lower than present. These observations represent a measure of relative sea-level change which can involve a land-movement signal as well as an ocean-volume signal. The indicators of submerged or elevated coastlines therefore point to one of three occurrences: land has moved up or down, ocean volumes have changed, or both have occurred simultaneously.

Tectonic process operating within the earth have caused uplift and subsidence throughout the Earth's history, resulting in relative sea-level change on a wide range of spatial and temporal scales. They include uplift and subsidence at convergent plate margins where the relative sea-level change is usually episodic and abrupt but cumulative over long periods of time resulting in, for example, the marine mollusc beds high in the Andes of South America that were first described by Charles Darwin. The tectonic processes also include slower and longer-duration events such as the initiation of continental rifting and sea floor spreading with the concomitant changes in the displacement of water by the developing ocean ridge system. Long-term thermal contraction of the cooling outer layers of newly created ocean crust at the ocean ridge results in sea-floor subsidence, creating basins into which sediments accumulate, thereby magnifying the subsidence. Large volcanic edifices stress the earth and cause more local subsidence and deformation of the earth's surface in the vicinity of the load.

At the same time that the tectonics events shape the earth's surface and shift the relative positions of land and sea surfaces, ocean volumes also change, largely because of

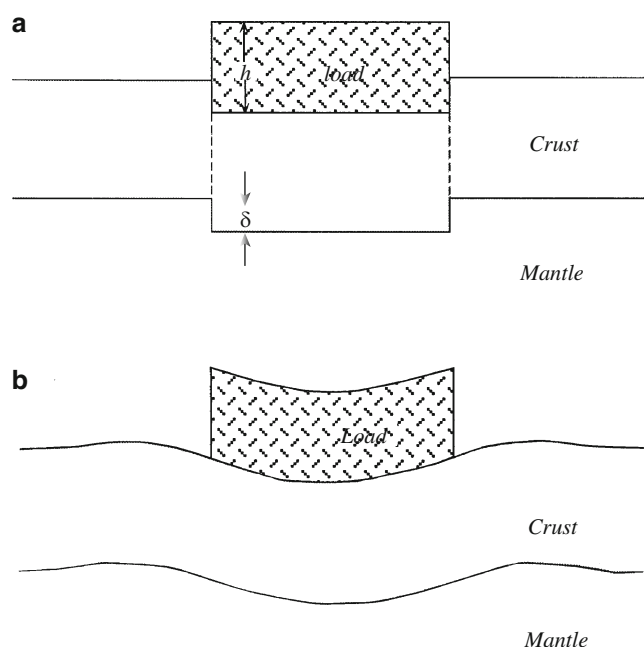
climate-driven changes in the extent of glaciation of the planet. During extended cold periods large ice sheets form, extracting water from the oceans and lowering sea levels. As the climate warms up sufficiently to melt the ice sheets sea levels again rise. Such glacial cycles have occurred at intervals throughout much of the earth's history but they have been most significant during recent times, the Quaternary period, for which the record has not yet been wholly overprinted by the subsequent tectonic and land-shaping events.

The combined result of the tectonics and glacial cycles is a sea-level signal that has varied significantly in time as well as being geographically variable. The record of this variability is, however, far from complete, and to be able to model and predict the migration of coastlines, an understanding and separation of the underlying causes of sea-level change is essential. Isostatic processes are key elements in this understanding and separation.

The Isostatic Process

Isostasy is the tendency of the earth's crust and lithosphere – the upper, effectively elastic layer of the earth – to adjust its vertical position when loaded at its surface by, for example, ice, water, volcanos, or sediments. For this purpose, the earth can be represented to a good approximation as a spherically symmetric body with a fluid core of about 3,400 km radius. The upper layer is called the lithosphere and includes the crust. Its thickness is typically between 50 and 150 km, varying with the tectonic history of the region. The lithosphere is characterized by being relatively cold and to behave elastically when subjected to load stresses below a critical failure limit. The mantle, between the lithosphere and core, is at a temperature that is relatively close to the melting point of terrestrial materials. As a result the mantle flows viscously, with characteristic relaxation times of 10^4 – 10^5 years, when subject to non-hydro-static stress. It is this zonation of a “rigid” lithosphere over “viscous” mantle that gives validity to the isostatic models.

The simplest representation of isostasy is by “local” response models which are statement of Archimedes' principle: a load of heights h , density ρ , placed on the earth's surface results in a subsidence of the underlying surface of $\delta = h\rho/\rho_m$, where ρ_m , the density of the mantle, exceeds the density of the lithosphere (see Fig. 1a). This model assumes that the crust or the lithosphere has no shear strength (or has failed under the load) and overlies a fluid mantle. This model, while unrealistic in many respects, is nevertheless useful for estimating magnitudes of crustal deflection beneath loads. For example, under a 3 km thick ice sheet the crust is predicted to deflect by about 1 km. A more reasonable model is one in which the load is supported by both the “elastic” strength of the crust-lithosphere and by the buoyancy forces at the base



Isostasy, Fig. 1 Models (a) local isostasy (b) regional isostasy. In (a) the load is supported by the buoyancy force at the base of the crust or lithosphere, whereas in (b) the load is also supported by the elastic stresses created in this layer. As the load diameter in (b) increases the isostatic response at the center of the load approaches that of local isostasy

of the layer (Fig. 1b). In this model, the mantle also behaves as a fluid and it provides a reasonable description of the earth's response to loads with time constants that are longer than the relaxation times of the mantle. These models have been extensively used to represent the response of the earth to sediment loads or to volcanic loads. They are usually referred to as regional isostatic models.

When the load duration is of the order 10^3 – 10^5 years any load-generated stresses that have propagated into the mantle will not have relaxed and the viscosity of the mantle must be taken into account. In these cases the isostatic models are usually represented by an elastic layer over a viscous or viscoelastic halfspace or, in the case of global problems, by spherical shell models of an elastic lithosphere over a viscoelastic mantle and fluid core. Both lithosphere and mantle may be represented by some degree of layering in physical properties (elastic moduli, viscosity, and density). Formulation of these spherical response models are well developed and solutions for the surface deformation under complex surface load geometries exist. Figure 2 illustrates an example of surface deformation where a large-diameter axis-symmetric ice sheet has been instantaneously removed. The rheology (viscosity structure) of the planet is realistic (see Fig. 6, below) and the results indicate that the crustal readjustment

continues for thousands of years after the unloading is completed.

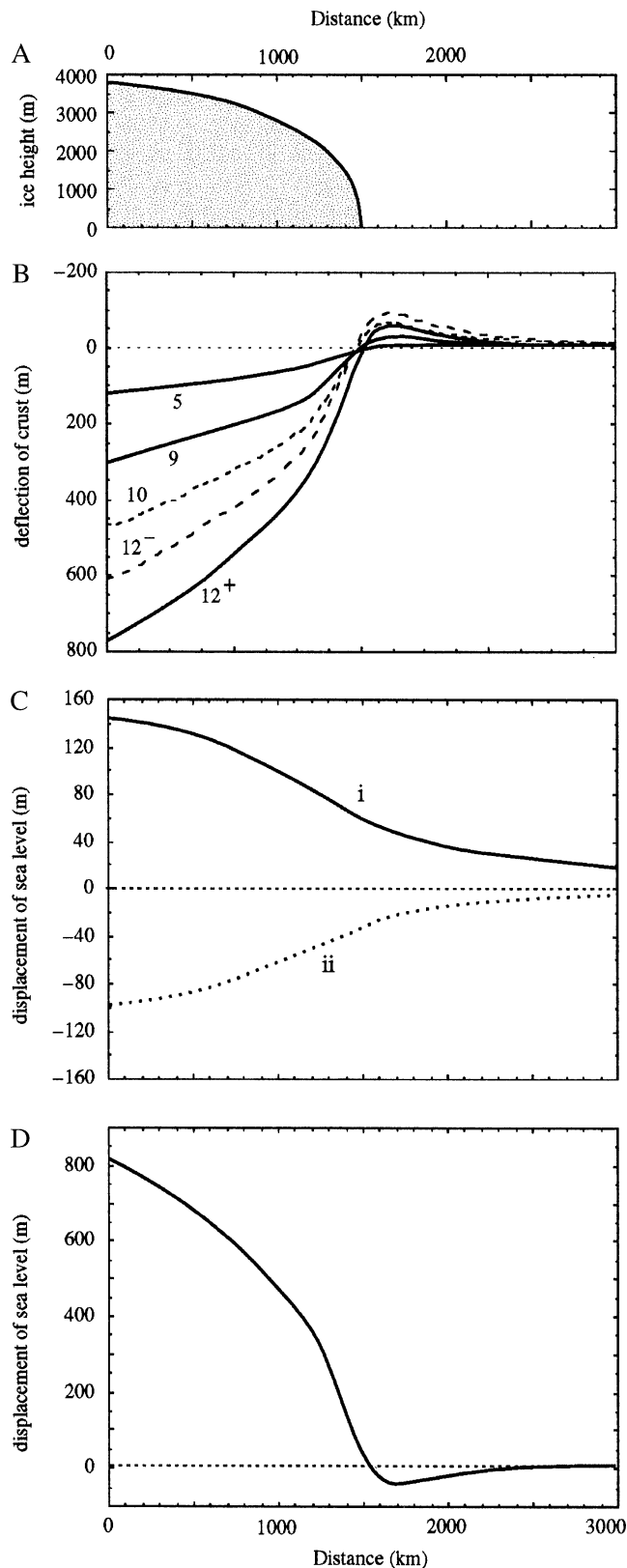
In addition to the surface deformation, the gravity field of the planet also changes under the load: the shape of the envelope containing the mass is modified by the deformation and material is redistributed within this envelope. At the same time there is a redistribution of the material on the surface: sediments are transported from mountains into basins, or the meltwater from land-based ice sheets flows into the oceans. Surfaces of constant gravitational potential – surfaces on which the gravity vector is everywhere perpendicular – therefore, change with time as the load and planetary response evolve. One such equipotential is the geoid, the shape of the ocean. (If the ocean is not an equipotential surface then the gravity vector has a component along the surface and ocean currents result until an equilibrium state is reached; thus in the absence of winds and other perturbing forces, the ocean will be an equipotential surface. This is called the geoid.) Figure 2 illustrates the change in the equipotential surface resulting from the unloading. It includes a contribution from the surface load itself – the ice “attracts” the ocean water and pulls the ocean surface up around it (curve *ii*) – and a contribution from the earth's deformation (curve *i*). The illustration is for the period while the ice is intact and when melting starts both curves will evolve with time.

The example in Fig. 2 illustrates that relative sea-level change resulting from the removal of the ice sheet contains several elements. The crust is displaced radially, the ocean surface is deformed by the redistribution of surface and internal mass, and water is added to the ocean. The rebound resulting from the melting (or growth) of the ice sheet is referred to as glacio-isostasy. The water added to (or withdrawn from) the oceans has its own isostatic effect and is referred to as hydro-isostasy.

The combined glacio–hydro-isostatic processes are of global extent. The melting of an ice sheet in one location modifies sea level globally, not just by changing the amount of water in the ocean but because of the planet's isostatic response to the changing surface load of ice and water. Other loading processes, such as by sediments or volcanic loads, are usually more local in their consequences. Also, these tectonic process generally occur on longer time scales so that the mantle response can usually be approximated as a fluid, and the local or regional isostatic models are mostly appropriate.

Glacio-Isostasy

Ice sheets represent surface loads that reach radii in excess of 1,000 km and thickness approaching 3 km. These loads are large enough to deform the earth and to produce substantial



Isostasy, Fig. 2 (continued)

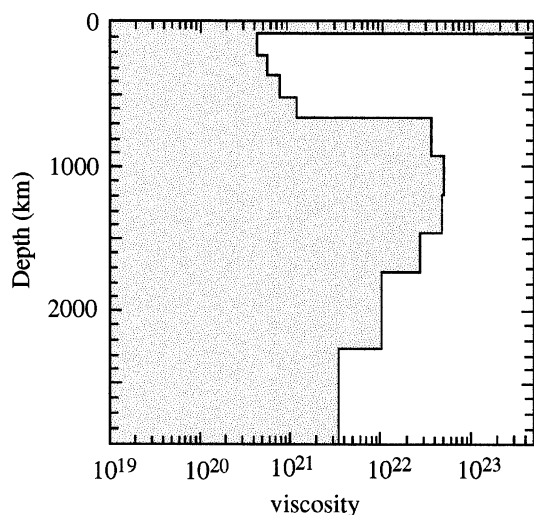
changes in sea level as illustrated in Fig. 2. Glacio-isostasy is the major cause of sea-level change in areas of former glaciation. When a large ice sheet melts the rebound of the crust is of larger amplitude than the rise in sea level resulting from the addition of the meltwater to the oceans (typically 120–130 m, see Fig. 8 below) from all of the ice sheets. If ΔV_i is the change in volume of ice on land and A_o the area of the ocean, then this second signal is

$$-\frac{\rho_i}{\rho_w} \int \frac{1}{A_o(t)} \frac{d}{dt} \Delta V_i(t) dt$$

where ρ_i , ρ_w are the densities of ice and water, respectively, and both A_o and ΔV_i are functions of time. This contribution is referred to as the ice-equivalent sea-level change.

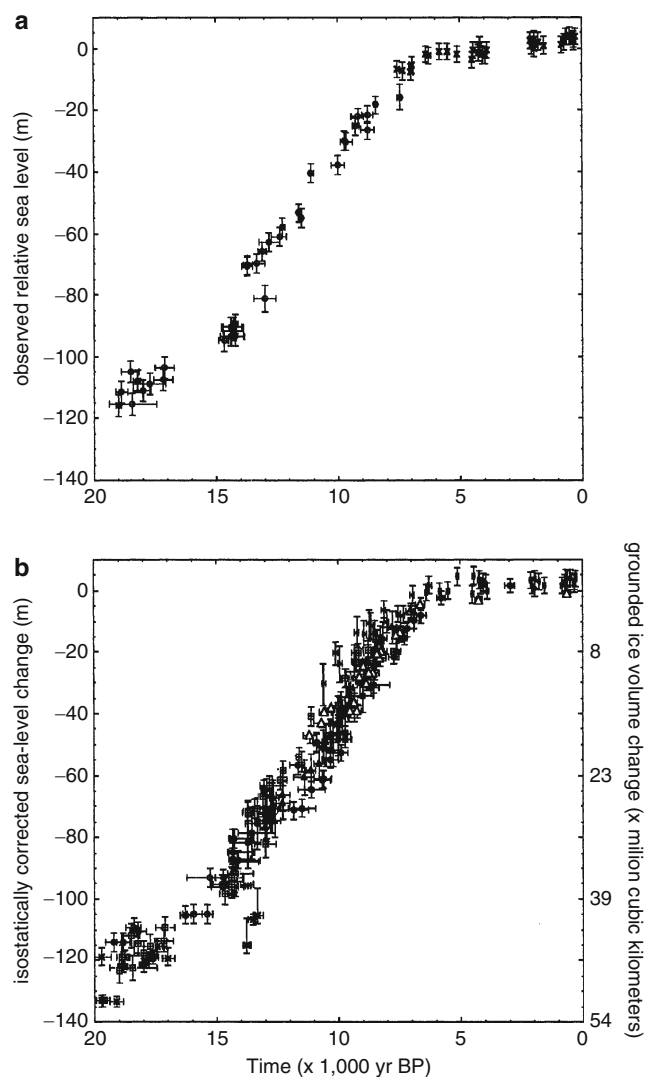
Because of the viscosity of the mantle, the crust continues to rise long after the ice has vanished and sea level appears to have fallen since deglaciation. This is seen in the Gulf of Bothnia and northern shores of the Baltic Sea, as well as in the Hudson Bay area of northern Canada. For these locations near former centers of glaciation the rebound signal dominates and the observed sea-level curves are characteristic relaxation curves (although only the post-glacial part of the change is recorded) (Fig. 3 (Angermanälven)). Near the ice margins the rebound is reduced in magnitude and may become comparable to the rise resulting from the increase in ocean volume. Now the time dependence of the sea-level change becomes more complex, with its character depending on the relative importance of the two contributions. In Fig. 3 (Andøya), for a site just within the ice-sheet margin, the rebound initially dominates but later, because of the melting of other and distant ice sheets, the ocean volume increase becomes the dominant factor and sea level rises until a time when all ice sheets have melted. The remaining signal is a late stage of the relaxation process and sea levels continue to fall up to the present.

Isostasy, Fig. 2 (a) Radial cross section of axisymmetric ice sheet. (b) Deformation of the earth's surface under the ice that has loaded the earth for 20,000 years (curve 12+). At 12,000 years ago the load is removed instantaneously. The initial response is elastic (curve 12-) and this is followed by viscoelastic creep, the surface being shown at 10,000 years (10), 9,000 years, and 5,000 years ago. (c) The gravitational attraction of the ice load, represented as the deflection of the geoid (i), and the change in geoid from the change in the planet's gravity due to the deformation of earth under the load (ii). The results are shown for a period before unloading starts. (d) The relative sea-level change, due to the combination of crustal deformation, change in gravitational attraction, and ocean volume change long after the load has been removed. The sea level is expressed with respect to its present position. If the coastline formed near the center of the load soon after the ice melted, it would now be at nearly 800 m elevation



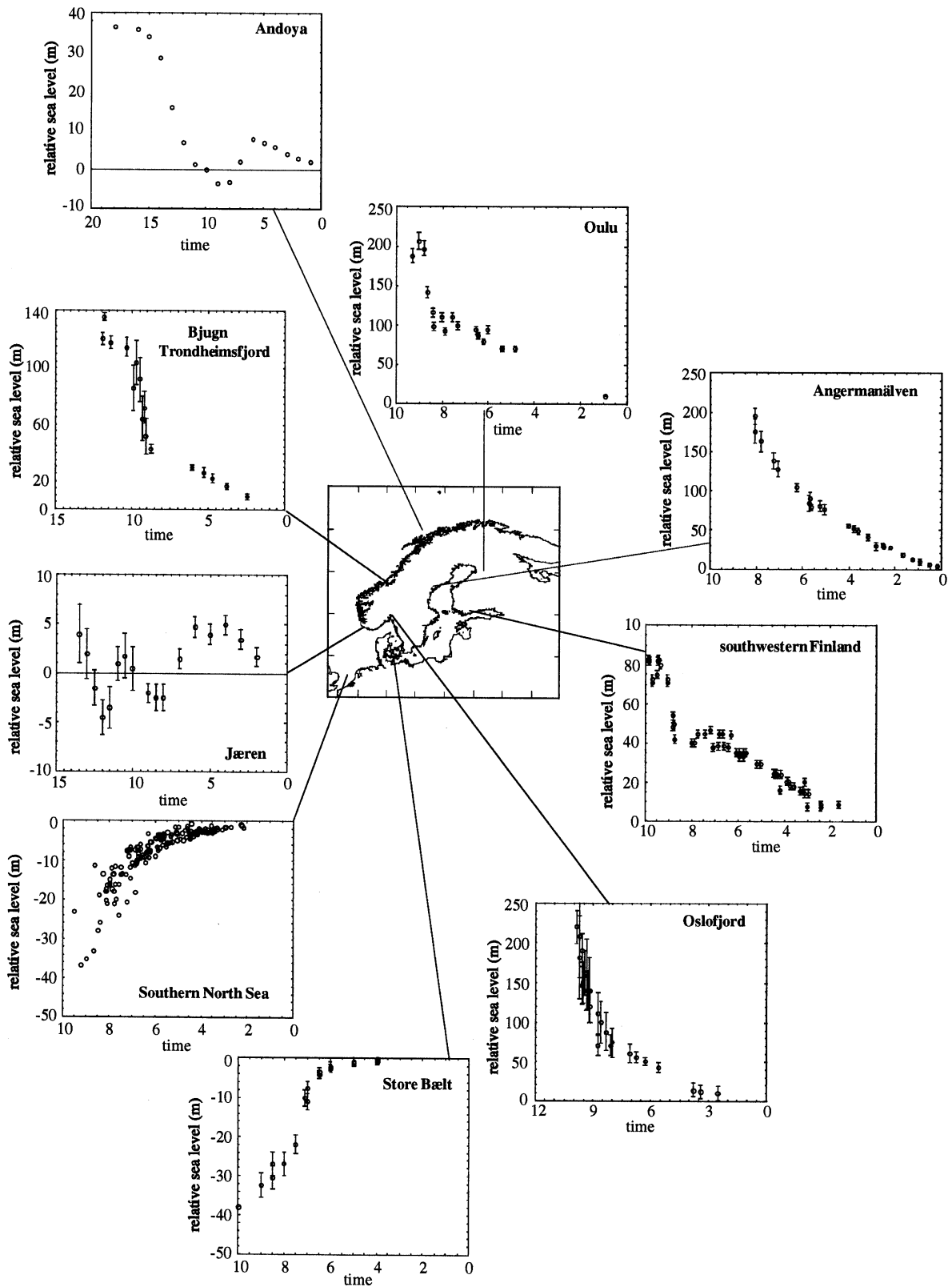
Isostasy, Fig. 3 Observed relative sea-level change for sites in Scandinavia illustrating some of the spatial variability in the response. The ice sheet covered all of Scandinavia and spread onto the German, Polish, and Russian plains. Retreat started at about 18,000 years ago and the final disappearance of ice occurred at about 10,000 years ago. The time scale used in these plots corresponds to the radiocarbon time scale which differs from a calendar time scale by about 10–15% for this interval ($1 \text{ C}^{14} \text{ years} \approx (1.1\text{--}1.15 \text{ calendar year})$)

The rate and magnitude of the sea-level change is a function of the earth's viscosity and the ice history: of the duration of the ice load, of its areal extent, and of its thickness. The importance of the rebound phenomenon is that it provides a means of estimating the earth's rheology: if climate models and geomorphological observations constrain the ice geometry through time, then observations of sea-level change provide a constraint on the mantle viscosity. If the ice models are not sufficiently well-known then it becomes possible to learn something about the ice sheets as well. Figure 4 illustrates observational results for sea-level change across Scandinavia. Here, the ice sheet reached its maximum at about 20,000 years ago and most melting occurred between about 16,000 and 10,000 years ago. As the ice retreated, coastlines formed on the emerging land providing a comprehensive description of the rebound across northern Europe. The rebound did not cease at the time melting ceased and coastlines have continued to retreat in formerly glaciated regions up to the present. This can be seen in tide gauge records across the Baltic, with present sea-level falling locally at rates approaching 1 cm/year in the northern part of the Gulf of Bothnia. Figure 5 illustrates the rate of crustal rebound and to obtain relative sea-level change these values must be increased by about 1–1.5 mm/year. Coastlines here continue to retreat despite other factors that may contribute to an increase in global ocean volume.



Isostasy, Fig. 4 Schematic contributions to sea level change from the glacio-isostatically driven crustal rebound and increase in ocean volume from meltwater. (a) For a location near a former center of glaciation where the rebound (i) exceeds the rise in sea level (ii) from the added meltwater, (iii) is the total change. (b) For a location near the former ice margin where the two contributions are of comparable magnitude but of opposite sign. (c) For a location beyond the ice margin where the crustal uplift is replaced by subsidence. The effect of the water load (ii) now is important as well as the ice-load effect (i). The meltwater contribution is given by (ii) and the total change by (iv)

Glacio-isostasy does not cease at the ice sheet margins. Because the mantle flow generated by the changing surface load is constrained within a deformable shell, when some areas are depressed under a growing load others are uplifted. The latter areas form a broad zone or swell around the area of glaciation of amplitude that may, depending on the size of the ice sheet, reach a few tens of meters. When the ice sheet melts this peripheral swell

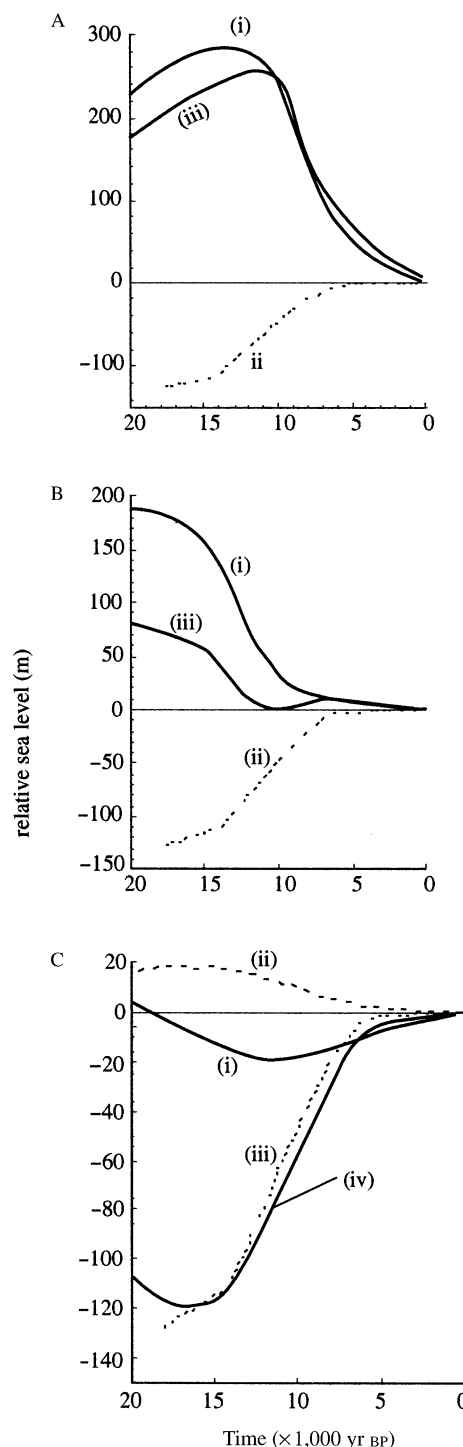


Isostasy, Fig. 5 Present rates of crustal uplift (in mm/year) of Scandinavia based on rebound models and on observed rates from tide gauges across the region. (From Lambeck et al. 1998; with permission of Blackwell Publishing)

subsides and for island or coastlines on it sea level will be seen to be rising at a rate that is over and above the ice-volume equivalent contribution (Fig. 3 (Store Bælt)). Beyond the Scandinavian relic ice margins this occurs in areas of the North Sea and as far away as the western and central Mediterranean and here the sea level continues to rise even when all melting has ceased. Beyond the North American ice sheet this zone of recent crustal subsidence and marine flooding occurs as far away as the southern USA and Caribbean.

Observations of sea level within and beyond the former ice margins provide the principal source of information on mantle viscosity. A typical result for northwestern Europe is illustrated in Fig. 6 where the rebound phenomenon provides a good constraint on the viscosity of the upper mantle. The main features of the viscosity profile include a lithosphere of thickness 65–75 km, a relatively low value for the viscosity of the mantle immediately below the lithosphere, and increasing viscosity with depth, particularly at a depth of about 700 km. Analyses for different regions produce comparable results although actual values for the viscosity and lithospheric thickness may differ because of the possibility that the rheology is laterally variable. The determination of such variability is one of the important research areas in glacio-isostasy.

While the glacio-isostatic models are well understood, one of the key limitations of their application is the inadequate knowledge of the former ice sheets. The ice margins at the time of the Last Glacial Maximum, some 20,000 years ago, are usually well-defined by geomorphological markers but the timing of their formation is not always known. This occurs particularly where the ice margins stood offshore and left few datable traces of both the time of their formation and of their retreat. Also, the ice thickness cannot usually be inferred from observational evidence alone and is inferred instead from glaciological and climate models. The sea-level observations can nevertheless help constrain the ice models in important ways. Thus, the total ice volumes in the models for all the major ice sheets must yield a global sea-level curve that is consistent with the changes observed far from the ice sheets (see section “[Hydro-Isostasy](#)”). Also, details in the ice models can also be derived from the sea-level data from sites within and near the former ice margins. The shape of the sea-level curve from a near-margin site (Fig. 3) changes quite rapidly with distance from the former ice margin, with the signal evolving from that for a central-load site to that for a site on the peripheral swell, and observations across the margin can constrain the former ice distribution within the ice-marginal region. One of the more recent research directions in glacio-isostasy is the use of this sea-level and crustal-rebound evidence to improve models of the ice sheets during the last deglaciation phase.



Isostasy, Fig. 6 Profile of mantle viscosity (in units of Pa s) inferred from glacial rebound analysis of European sea-level data. (G. Kaufmann, with permission)

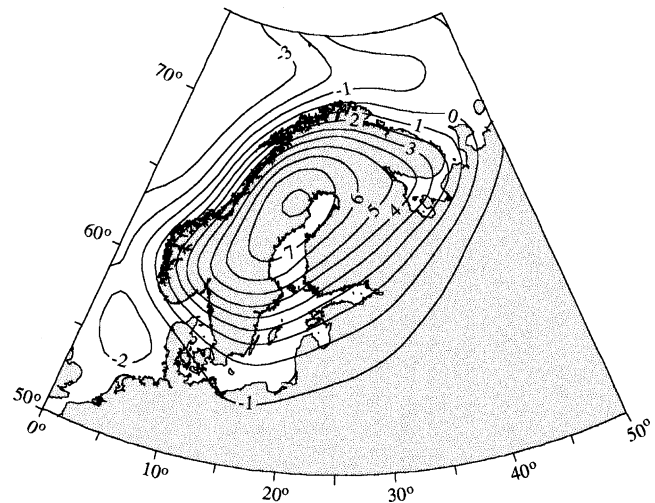
Hydro-Isostasy

As ice sheets melt, the additional water entering the world's oceans loads the sea floor, load stresses are propagated through the elastic lithosphere into the mantle, the newly

stressed mantle material flows toward unstressed regions and the sea floor subsides. The shape and holding-capacity of the ocean basin is thereby modified and the ocean water is redistributed, changing sea level. This adjustment of the earth under the time-dependent water load and the concomitant sea-level change is referred to as hydro-isostasy. Since the onset of the last deglaciation, sea levels have risen on average by about 120–130 m and the additional load has been sufficient to modify the shape of the earth. This is a result of the long wavelength nature of the water load. Loads of dimensions less than the thickness of the lithosphere are supported mainly by the strength of the lithosphere and the resulting surface deformation is small. But large-dimension loads effectively see through the lithosphere and are supported by the much more ductile mantle which flows even under small changes in the stress field.

At continental margins the hydro-isostatic deformation of the earth's surface describes quite complex patterns because of the geometry of the load. The lithosphere acts as a continuous elastic layer or shell and the continental margin is dragged down by the subsiding ocean lithosphere but, because of the asymmetry of the load, not by the same amount as in mid-ocean. At the continental coastlines, therefore, the subsidence will be less than it would be in mid-ocean. At the same time, some of the mantle material flowing away from the stressed oceanic mantle flows beneath the continental lithosphere, causing minor uplift of the interior. The net effect of the ocean volume increase is a seaward tilting of the continental margin which will be seen as a variable sea-level signal across the shelf. This effect is clearly seen for tectonically stable continents that lie far from former ice sheets, as in the case of Australia. While the ice sheets are still melting the dominant sea-level signal here is from the increase in ocean volume and the glacio- and hydro-isostatic effects are second order. But when melting ceases the on-going isostatic effects come into their own. Now sea level appears to be falling at the coastal site as the ocean waters recede to fill the still-deepening ocean. In consequence, small sea-level highstands are left behind with peak amplitudes of 1–3 m occurring at the time global melting ceased (Fig. 7). Such highstands are common features along many continental margins and manifest themselves as relic shorelines or fossil corals above the present formation level or habitat. If the coast is deeply indented, sites at the heads of gulfs, being furthest away from the water load, experience greatest uplift while offshore islands experience least uplift. This differential movement provides a direct measure of the viscosity of the mantle across the continental margin.

Like the glacio-isostatic effect, the water load does not only deform the surface of the earth, it also results in a redistribution of mass and a change in gravity and in the shape of equipotential surfaces. The total sea-level changes associated with the hydro-isostasy include, therefore, both the

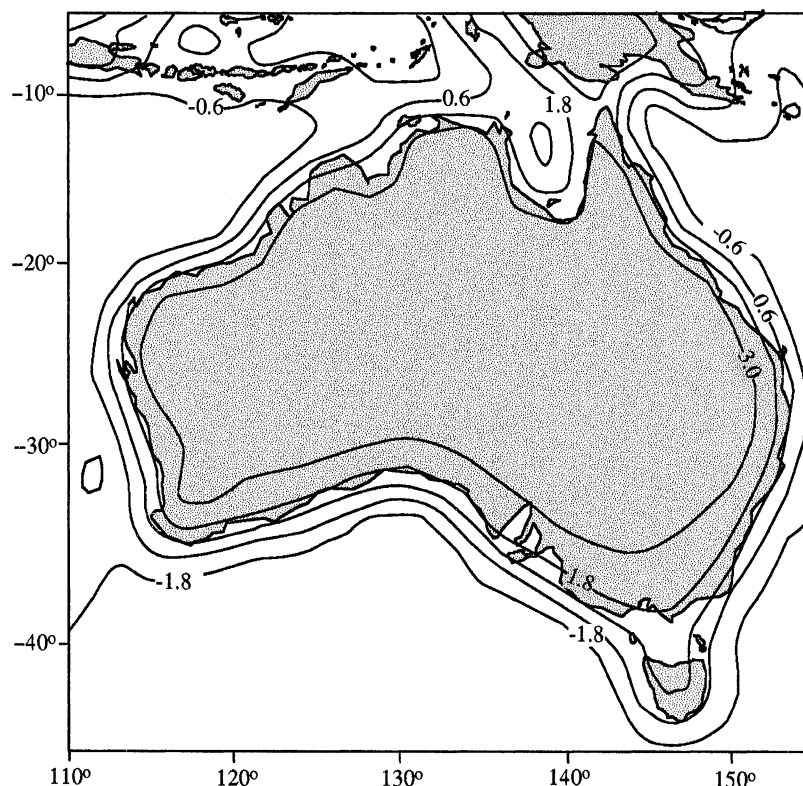


Isostasy, Fig. 7 Sea level at 6,000 years ago around the Australian margin illustrating the effect of hydro-isostasy as a tilting of the margins of the continents. Sea levels are present relative to present mean sea level. Contour intervals are 1.2 m. The present-day rate of change in mm/year given approximately by dividing the contour values by 6 and changing the sign, the resulting negative value for most locations indicating a fall in sea level from isostasy alone

crustal radial deflection and the associated geoid change. Also, the glacio and hydro-isostatic effects are closely linked when their cause is the deglaciation of the last ice sheets. Near the edge of the ice sheets, for example, the water is pulled up (Fig. 2) and the waterload is increased above what would result from a uniform distribution of the meltwater over the entire ocean. Here the hydro-isostatic signal is a function of the magnitude of the glacio-isostatic effects. Elsewhere, the broad zone of crustal rebound surrounding a large ice sheet may occur in an oceanic environment. Then, when the ice sheet melts this swell subsides, increasing the volume of the ocean basin, water is withdrawn from other parts of the ocean, and a further global adjustment of sea level occurs. Thus, the treatment of hydro-glacio isostasy requires a global and consistent formulation that ensures that these various interactions are included.

The hydro-isostatic signal is an on-going one even when major melting of the world's ice sheets ceased about 6,000–7,000 years ago. Thus sea-level change today will contain a small but not insignificant component of hydro-isostatic origin (cf. Fig. 7 for the Australian region). This signal must, of course be superimposed upon any other changes, including possible global warming signals. The results indicate that sea levels around the Australian margin are slowly falling under the combined glacio-hydro-isostatic response to the past melting of the large ice sheets (with the possible exception of Tasmania where the glacio-isostatic effect of Antarctic ice volume changes becomes significant, canceling out the hydro-isostatic signal such that little overall change now occurs). Similar isostatic effects will be present

Isostasy, Fig. 8 Sea-level change for the past 20,000 years. (a) A record of observed local relative sea-level change from Barbados and other Caribbean sites, and (b) isostatically corrected sea level from a number of sites distributed globally and combined into a single ice-equivalent sea-level curve. Scale on the right hand side gives the corresponding change in volume of ice on land and grounded on shallow sea floor



at all coastline, increasing in magnitude as the locality approaches the regions of former glaciation.

The importance of the sea-level observations far from the ice margins is that because the glacio-hydro isostatic effects are relatively small (10–15% of the total signal) they provide an estimate of the change in volume of the oceans when corrected for the isostatic effects: the observed sea level, less the isostatic correction yields the ice-equivalent sea level defined above and hence an estimate of the change in ocean volume ΔV_i . Several long records, extending back to the Last Glacial Maximum, of local sea-level change exist which provide evidence for the change in ice volume since this time. They indicate (Fig. 8) that maximum ice volumes globally were $(50\text{--}55) \times 10^6 \text{ km}^3$ greater than today but, they do not indicate necessarily where this extra ice was stored. To resolve that issue recourse to the study of the glacio-isostatic process from formerly glaciated regions is necessary.

Sediment and Volcanic Loading

Large accumulations of sediment occur along many of the continental margins reaching, in some instances, a thickness of 10 km or more. The rate of accumulation is usually slow and continuous, occurring over periods of tens of millions of years with the sources of sediments coming from continental interiors where tectonic processes have caused uplift and erosion processes have carried the sediments to the sea.

Examples include the Bay of Bengal, the northwestern margins of Europe, the eastern margin of North America, and the Gulf of Mexico. Thick accumulations of sediments are possible because of the subsidence of the lithosphere under the growing sediment load. With the above model of local isostasy an ocean basin of depth d_o can, with adequate sediment supply, lead to a maximum subsidence of $d_o \rho_s (\rho_s - \rho_m)$ where ρ_s is the density of sediments. This assumes that the basin is ultimately filled to sea level. For $d_o = 4 \text{ km}$, $\rho_s = 2.5 \text{ g cm}^{-3}$, $\rho_m = 3.5 \text{ g cm}^{-3}$ the maximum thickness of sediment that can be attained is about 10 km. However, in this case the deeper sediments will have been deposited in water depths initially of $d_o = 4 \text{ km}$, whereas the characteristics of the fauna preserved in the basin sediments usually indicate that deposition invariably occurred in relatively shallow waters. Isostasy alone, therefore, cannot produce thick sediment sequences but it does act as an amplifier of subsidence that is the result of other processes: in this case mostly the thermal contraction of ocean lithosphere as it cools from an initially hot layer formed at the ocean ridges and then moves away from the heat source.

On short time scales, sediment loading can lead to substantial coastal subsidence. This may occur in conjunction with deglaciation cycles where sediments are eroded from the continents during the deglaciation stage and delivered to coastal environments at some later stage. An example of such subsidence occurs along the US coast of the Gulf of Mexico, particularly for the Mississippi delta. Here, coastal

subsidence occurs at rates approaching 10 mm/year and are attributed in part to the isostatic response to recently delivered sediments, but also in part to the extraction of fluids from the sediments and the associated compaction. Here, as in most isostatic problems, several factors will contribute to the observed signal.

Volcanic loading of the crust provides another example of isostasy at work. Large volcanic complexes form on the sea floor, and elsewhere, because of upwelling convection currents in the mantle that lead to an injection of magma into the crust and ultimately onto the surface as volcanos. The mantle source regions for the magma appear to be long-lived and as the lithosphere moves over the earth's surface under the forces of plate tectonics, a trail of volcanos is left on the surface. The Hawaiian chain provides the type example. Other examples include the Society Island chain whose current center of volcanic activity lies to the east of Tahiti. The subsidence of the lithosphere beneath the volcano is adequately described by the regional isostatic model in which the load is supported by the elastic stresses within the lithosphere and by the buoyancy force at the base of the layer (Fig. 1b). Because of the elastic properties of the lithosphere small peripheral bulges, concentric about the center of loading, develop and any islands located in this zone at the time of volcano development are uplifted by some tens of meters. An example of this is provided by the uplifted atoll that forms Henderson island, southeast of Pitcairn island. This small island about 200 km from the volcanic island of Pitcairn, appear to have been uplifted some 20–30 m at the time of Pitcairn's formation, perhaps 700,000 years ago. The location of the zone of maximum peripheral uplift provides a measure of the flexural wavelength of the lithosphere, a parameter that characterizes the physical response of the layer to loading. With time, some relaxation of the loading stresses can be expected to occur within this layer such that the isostatic state evolves slowly from regional to local isostasy and that the volcano slowly subsides. Thus Tahiti, a relative young volcanic load of about 1–2 million years, may be subsiding at a rate of about 0.2 mm/year or less.

These examples of vertical movements driven by sediment or volcanic loading illustrate the interaction that occur between the various isostatic contributions to sea-level change. To estimate rates of tectonic uplift or

subsidence, heights of identifiable coastlines are measured with respect to present sea level. Thus, the fluctuations in sea level of glacio-isostatic origin must be known, but these fluctuations are inferred from the same observational evidence. An important research area is to develop methods for separating out these effects, through observational improvements and through improved modeling of the physical processes.

Suggested further reading on this subject may be found in the following bibliography.

Cross-References

- ▶ [Changing Sea Levels](#)
- ▶ [Coastal Changes, Gradual](#)
- ▶ [Coastal Changes, Rapid](#)
- ▶ [Coastal Subsidence](#)
- ▶ [Endogenic and Exogenic Factors](#)
- ▶ [Eustasy](#)
- ▶ [Geodesy](#)
- ▶ [Glaciated Coasts](#)
- ▶ [Ingression, Regression, and Transgression](#)
- ▶ [Paleocoastlines](#)
- ▶ [Sea-Level and Climate Change](#)
- ▶ [Sea-Level Fluctuations Over the Last Millennium](#)
- ▶ [Submerged Coasts](#)
- ▶ [Tidal Datums](#)
- ▶ [Uplift Coasts](#)

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Jet Probes

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Introduction

Jet probes are used for a variety of purposes (e.g., underwater cable routing, marine archaeology, coastal engineering) and are usually deployed in conjunction with other data-collection techniques such as hydrographic surveys (to determine water depths and map existing bottom conditions), subbottom profile survey (to identify near subbottom stratigraphy, 3–7 m depth), side-scan sonar survey (to identify morphological variations, and natural and man-made obstructions on the seabed), and vibratory coring (to acquire direct physical information of nearsurface sediments). Jet probe surveys acquire indirect physical information on subsurface lithology by surveying the thickness and stratigraphic layering of sedimentary covers on land or underwater. A jet of either air or water is used to penetrate the sand cover; the latter, however, is only applicable underwater (USACOE 2002).

Most jet probe surveys, in the service of coastal engineering for shore protection via beach renourishment, provide a rapid means for determining the nature of unconsolidated sedimentary deposits that occur underwater. Because jet probes have no cutter head and depend only on the power of a water jet to penetrate bottom sediments, they are restricted to use in shallow waters (i.e., the effective range of operating depths is usually from 1 m to about 30 m) that overlie unconsolidated (loose) sandy deposits. Clear water is desirable, but not essential, because it facilitates site location, maneuverability of jet probe equipment over the bottom, and visual estimation of turbidity plumes that are created by water-jetted penetration of

a pipe down through the sediment (CBNP 1995). Jet probing finds application in marine archaeology, geotechnical studies that feature searches of seabed deposits for beach-compatible sands that can be placed on degraded beaches, and geological investigations that attempt to determine the thickness of sand covers on the seafloor or on lakebeds. Although widely deployed in many different kinds of environments and for various applications by scientists and engineers, jet probing probably finds most extensive application in coastal sand searches (e.g., Meisburger and Williams 1981; Meisburger 1990; Keehn and Campbell 1997; Finkl et al. 1997, 2000, 2003; Andrews 2002) that require reconnaissance surveys of bottom types or verification of geophysical survey data (e.g., subbottom profiles, side scan sonar surveys).

Grab samples provide information about surficial seafloor sediments, whereas vibracore and jet-probe samples can penetrate down into the sediment layers. Vibracore samples are relatively inexpensive to obtain and can recover the long and relatively undisturbed cores that are required to assess the composition and grain sizes of the materials, as well as to establish the stratigraphy of the deposits (e.g., Meisburger and Williams 1981). Water jets are less expensive than cores (CBNP 1995; USACOE 2002), involving the water-jetted penetration of a pipe down through the sediment in order to determine the layering, as opposed to (undisturbed) core retrieval for splitting and analysis.

Marine Archaeology

This tool assists marine archaeologists in determining the nature or presence of materials or features that lie within or underneath bottom sediments (Anon 1996). On archaeological sites, the jet probe is manually driven through various non-consolidated sediments on the seabottom where the probing pipe goes through soft strata until it hits bedrock, a cemented stratum, compacted clay, or artifact. This tool provides information regarding the location and elevation of buried ancient waterline features (indicators of previous sea-level positions)

and other geomorphological data. Ultimately, the information enables the archaeologist to reconstruct shallow coastal-marine sedimentary environments, local surficial stratigraphic sequences, and other geological features that can then be dated and calibrated with archaeological finds.

Stratigraphic Studies

Coastal-marine stratigraphic studies often rely on a range of techniques that are used to compile various kinds of information, that is, related to layering of different kinds of materials on the seabed (e.g., Toscano and Kerhin 1990; Wells 1994). Data are commonly derived from several independent studies viz. surface sediment samples, vibracores, and seismic records to compile an assessment of Quaternary stratigraphy, as, for example, in the Paranaguá Bay Estuary in southern Brazil (Lessa et al. 2000). Estuarine environments often provide ideal conditions for jet probing because there is a range of unconsolidated materials related to coarse- and fine-grained facies. Fluvial- continental deposits often occur with paleo-valleys as substratum for more recent sedimentation. These kinds of estuarine environments are often characterized by the intercalation of transgressive-regressive mud and sand facies that can be effectively studied using jet probes in conjunction with other techniques.

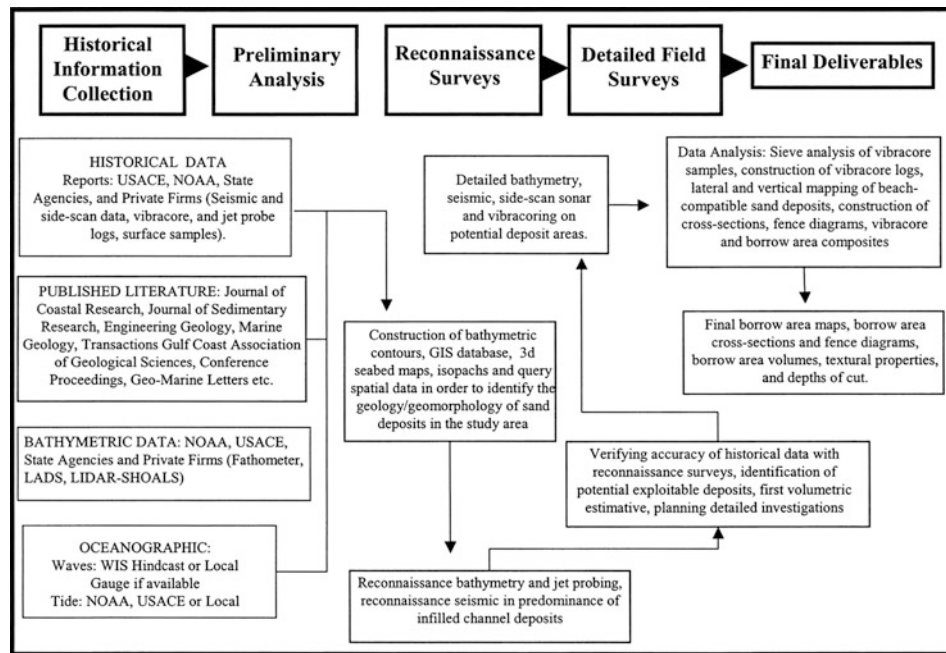
Underwater surveys of lakebeds often use jet probes to assist in reconnaissance verification of sedimentary bottom types, especially where sediment samples and grain-size analyses are eventually required. Lakebed studies often combine jet probing with underwater video investigation as independent lines of inquiry. Jet probe surveys to determine the thickness of sand cover are based on differentiation of the kinds of materials that are penetrated by the jet probing. On the American Great lakes, for example, the presence of diamictites (tills) that have been eroded from truncated drumlins to produce cobble-boulder lag deposits on the lakebed can limit the effectiveness of jet probes, as would any other substantial impediment to penetration of sedimentary layers (e.g., Stewart 2000).

Assessment of Sand Resources and Mining

Sandy shores occur along about 13% of the world's coastline (Coleman and Murray 1976) and it is estimated that today about 75% of these shores are eroding (Bird 1985). Beach erosion is thus a common problem along sandy coastlines and it is necessary to artificially renourish beaches because they provide natural protection from storms and have economic value (Finkl and Walker 2002). The location of materials that are suitable for beach renourishment becomes an issue for best management practices that have to consider

environmental concerns, methods of shore protection, storminess, and impact of exploration procedures to locate sand bodies on the seafloor. Even though sand sources differ from region to region around the world, there is a commonality to the need for good-quality sand and methods of looking for adequate longterm supply, as described, for example, by Anders et al. (1987), Conkright et al. (2000), and Walker and Finkl (2002). The salient problem then, is how to best locate sand sources that are appropriate for beach nourishment. Although inland sand sources are often suitable from a textural and compositional point of view for beach replenishment, their location away from the coast requires overland transport that can pose significant placement problems along the shore. Offshore sources of beach-quality sand are thus most often sought as geotechnical and economic reserves. Inner continental shelves host a range of coastal (e.g., beach ridges, dunes, nearshore bars, flood- and ebb-tidal deltas, estuarine sands) and marine sediments (e.g., shoals, banks, ridges, terraces, blanket deposits) as well as terrestrial deposits (e.g., glaciofluvial materials on valley floors, winnowed tills, coarse-grained alluvial terraces, and plains), all of which have been drowned and modified by rising sea levels during the Holocene (e.g., Toscano and York 1992). Offshore sands that are suitable for beach renourishment are a sought-after and coveted commodity because in many regions they are in dwindling supply (e.g., Freedenberg et al. 2000).

Advanced geophysical and geotechnical procedures are often backed up or verified by low tech efforts, such as jet probing, that are essential to the efficiency and economic success of offshore sand searches. General procedures for the exploration and development of borrows were summarized by James (1975) and Meisburger (1990), for example, who emphasized the value of collaborative approaches. Figure 1 shows the main sequential steps in modern data collection that are followed in sophisticated coastal sand searches that integrate diverse techniques. As shown in the figure, jet probe surveys are typically conducted at the reconnaissance level in conjunction with a suite of independent, but related, geophysical and geotechnical survey operations that provide specific kinds of information that collectively elucidate sediment thickness, lateral continuity, structural relationships, and composition. It is important to note that the first step in sand resource assessment is the review of historical data, an effort that is essential to proper appreciation of prior efforts and conclusions. Jet probe logs (Fig. 2) are archived because they contain useful information that may be required later. A typical jet probe log, as shown in Fig. 2, includes the usual kinds of locational information; date acquired, water depth, top and bottom divers, start- and end-times, etc. Notes in the log include important information that is related to the length of pipe, penetration depth, jet pump capacity, weather conditions, turbidity levels, and characteristics of the sand (grain size, percentage of silt content, color).



Jet Probes, Fig. 1 Flow diagram showing the organization, routing, and sequential application of coastal sand searches that are normally deployed on the inner continental shelf. Note that investigations begin with review of historical data, including proprietary reports and works in the public domain, and proceed to the construction of electronic databases that interface with GIS frameworks. Reconnaissance jet probing

assists in the verification of historical data and provides focusing criteria for conducting detailed surveys. After review of historical, laboratory, and field data, sand resources with the greatest potential for use as beach sediments are identified as borrow sites. Jet probe surveys provide critical information in the evaluation of offshore sand resources and help identify which deposits are exploitable

Sand searches for beach nourishment and protection commonly employ jet probe surveys (e.g., see Meisburger and Williams 1981; CBNP 1995; Walther 1995; Freedenberg et al. 2000; Finkl et al. 1997, 2000). Usually conducted as a reconnaissance field survey (cf. Fig. 1), the procedure is often misunderstood and the least utilized tool in sand search investigations. Usually deployed after preliminary assessment of historical data (e.g., geophysical and geotechnical information), comprehension of the regional geology and geomorphology, and computer aided analysis (including GIS summaries), jet probing should verify previously indicated field conditions. Jet probe surveys thus perform a valuable function in sand searches and their relevance and importance should not be underestimated as a time- and cost-saving effort.

Reconnaissance bathymetric and jet probe surveys are also used to verify hydrographic features with widely spaced bathymetric surveys, historical surface sand samples, jet probes, core sites, and other potential sand features in the study area. Reconnaissance bathymetric surveys groundtruth and verify the National Oceanic and Atmospheric Administration hydrographic data in selected areas of potential sand deposits. The reconnaissance bathymetry should be compared with historical bathymetry to identify areas where sand has accumulated by natural coastal processes or offshore dredge disposal. An example of reconnaissance jet-probe survey is

shown in Fig. 3 for a portion of the southwestern coast of Florida. Here, on the wide continental shelf of the eastern Gulf of Mexico, a range of sedimentary deposits overlie a karstified limestone peneplain that extends seaward from the Florida peninsula (Evans et al. 1985). Although the karst surface is somewhat irregular due to dissolution of the carbonate rocks, drowned valleys are infilled and planar areas are covered by blanket deposits and ridges.

Inlets along this coast, which produce deltaic deposits, show no strong regional trends and are stable in terms of channel width, length, geographic position, and orientation (Vincent et al. 1991; Finkl 1994). The low wave energy regime that influences sediment accumulation at inlets in this region enhances construction of large ebb-tidal deltas, which store enormous quantities of sand (Davis et al. 1993). Flood-tidal deltas along the west-central Florida coast are relatively inactive due to small tidal ranges, sheltered lagoons, and ebb-dominated inlets (Davis and Klay 1989; Finkl 1994). The wide Continental shelf offshore southwest Florida, described by Davis (1997) and which gently slopes seaward toward the central basin of the Gulf of Mexico, maintains shallow depths to 9 km offshore to the 10 m isobath. Shelf morphologies and coastal (inlet) morphodynamics impact spatia distributions of mineral resources (Wright 1995), large-scale coastal behavior (Short 1999), and barrier island evolution (Oertel 1979).

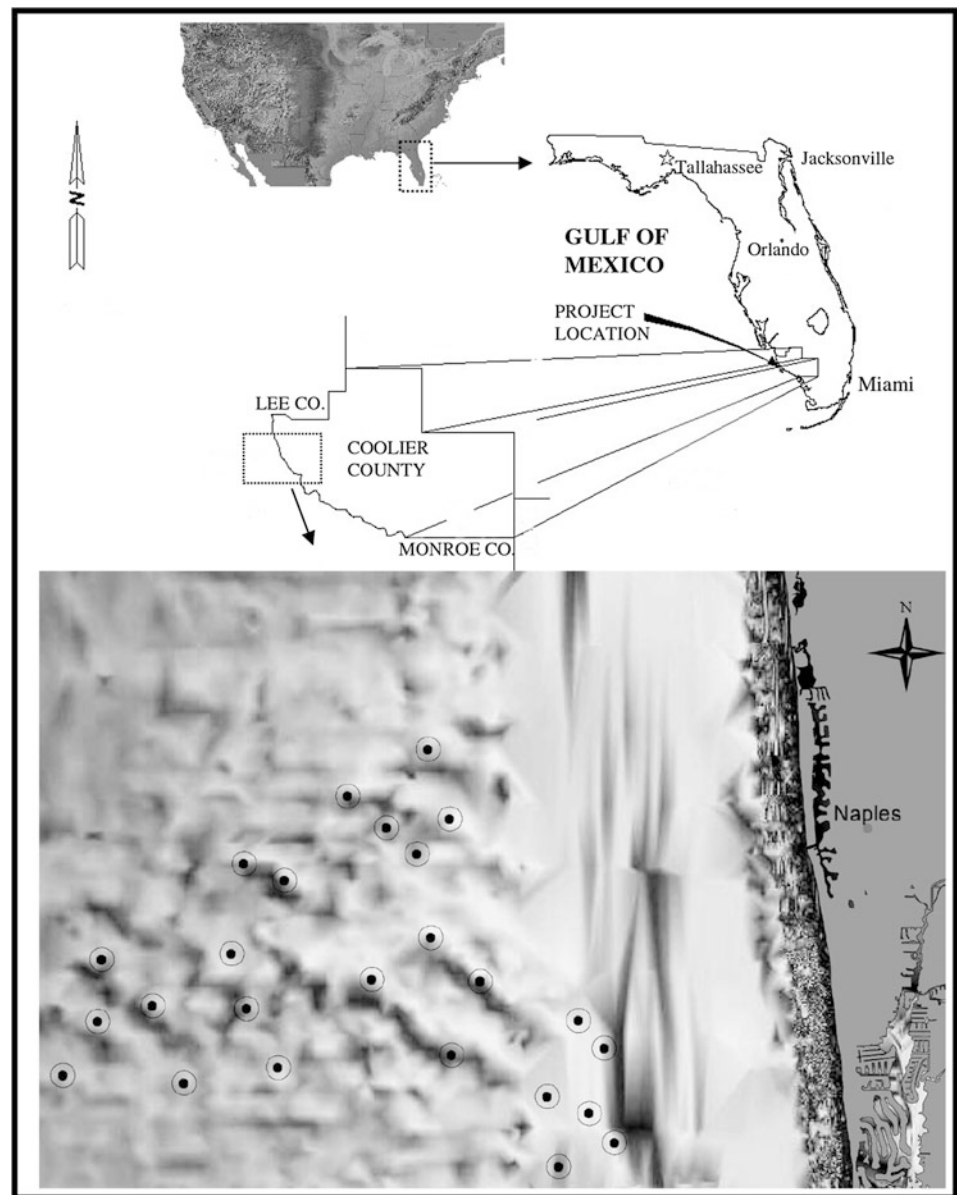
Jet Probes, Fig. 2 Example of a jet probe log showing the kind of information that is logged in verbal or numerical formats along with graphic displays of sediment composition. These digital logs are part of a GIS framework and the information contained in them can be queried for special purposes. Note that some information is back loaded into the logs because it is obtained subsequent to field logging. Granulometric analyses, for example, report median grain sizes for clasts (e.g., sand grains) and particulate matter (e.g., percent silt content)

JET PROBE LOG				
PROJECT: LONGBOAT KEY JET PROBES			JET PROBE: LBJP-02-14	
COORDINATES:	DATE: 03/27/02	WATER DEPTH: 24.8		
N = 1,132,829	START TIME: 1441	TOP DIVER: MTL		
E = 408,682	END TIME: 1447	BOTTOM DIVER: JW		
NOTES	ELEV.	DEPTH	SYMBOL	DESCRIPTION
LENGTH OF PVC PIPE: 20.0'	-24.8	0		SEA FLOOR
PENETRATION DEPTH: 18.0'		1		SAND (SW) With Some Shell Hash
JET PUMP TYPE: BRIGGS & STRATON 3.0 HP		2		
GAL/HR: 8460		3		
DIAMETER OF PIPE: 1.5"		4		
SUPPORT VESSEL: CPE II		5		
POSITIONING: DIFFERENTIAL GPS USCG BEACON		6		
NAVIGATION SYSTEM: "HYPACK"		7		
WEATHER:		8		
WIND:		9		
DIR: NORTHWEST		10		
SPEED: 10-15 Kt.		11		
WAVES:		12		
DIR: NORTHWEST		13		
HEIGHT: 1-3 Ft.		14		
CURRENT:	-37.8	15	SAND (SW) Fine Grained	
DIR: NORTH		16		
SPEED: SLIGHT		17		
		18		
SAND SAMPLES		19	REFUSAL	
TOP:-24.8 Ft., 0.17 mm, Silt: 1.59%		20		
Munsell Color: 5Y-7/1 Light Gray		21		
MID:-33.8 Ft., 0.14 mm, Silt: 2.07%		22		
Munsell Color: 5Y-7/1 Light Gray		23		
BOTTOM: -42.8 Ft., 0.15 mm, Silt: 2.46%	-42.8	24		
Munsell Color: 5Y-7/1 Light Gray				
TURBIDITY:				
TOP (0'-9'): Moderate				
BOTTOM: (9'-18'): Moderate				
DRAWN BY: MDA	CHECKED BY: JLA			
JOB NO: 8488.47				

Various types of sand ridges (linear accumulations of sand bodies) are common on inner shelves along many shores the world over (viz. Duane et al. 1972; Swift and Field 1981; McBride and Moslow 1991). These topographically positive sedimentary accumulations on the seafloor are recognized as relict sand bodies that formed in response to prior stillstands of mean sea level (MSL) when sea levels were lower than those of today. On the shelf off southwestern Florida, for

example, prominent seabed morphologies include linear sand ridges, some of which extend continuously for distances greater than 6 km. These deposits formed during the Flandrian Transgression (most recent Holocene trend in sea-level rise) (Davis 1997). Depressional (negative topographic) features are incised into the karst surface and some surficial marls. When the continental shelf was exposed to subaerial geomorphic processes during low stands of sea level, streams cut into

Jet Probes, Fig. 3 Jet probe location diagram showing bathymetry in terms of a graduated grayscale ramp so that sedimentary accumulations on the seafloor may be inferred from bathymetric highs. Jet probe locations, identified by the circled dots, are strategically placed to provide information related to the deposit thickness. Note the placement of jet probes on sand ridges



the karstified surface and persisted as valleys until sea level rose and they became infilled with recent marine and terrigenous sediments.

Figure 3 shows the distribution of reconnaissance jet probe locations of Naples, Florida. The jet probe locations are strategically placed on the basis of hydrographical, geophysical, geotechnical, and geological (including geomorphological) information (cf. Fig. 1) that indicates the presence of sand deposits. As illustrated here by the shaded bottom relief on the lower left side of the Fig. 3, caused by sedimentary accumulations on the seafloor, jet probes are sited on ridges, inter-ridge depressions, sand flats, and in other areas to verify thickness of sedimentary covers (or lack thereof). Emphasis must be placed on the

fact that jet probes are not randomly sited on the seabed just to see what is there; rather, siting is intelligently coordinated with all collateral data that is related to the nature of bedrock surfaces and sedimentary accumulations on the seafloor. Reconnaissance jet probing is a strategy that is conducted as part of an overall coordinated methodology to define the presence of beach-quality sands on the seafloor.

Jet probes are thus taken in areas that show promise for sand deposits and to confirm historical vibracore logs. Jet probes and surface sand samples provide an indication of the thickness and characteristics of the unconsolidated sediment layers. With two dive teams, consisting of a geologist and a support diver, generally 8-15 jet probes can be obtained

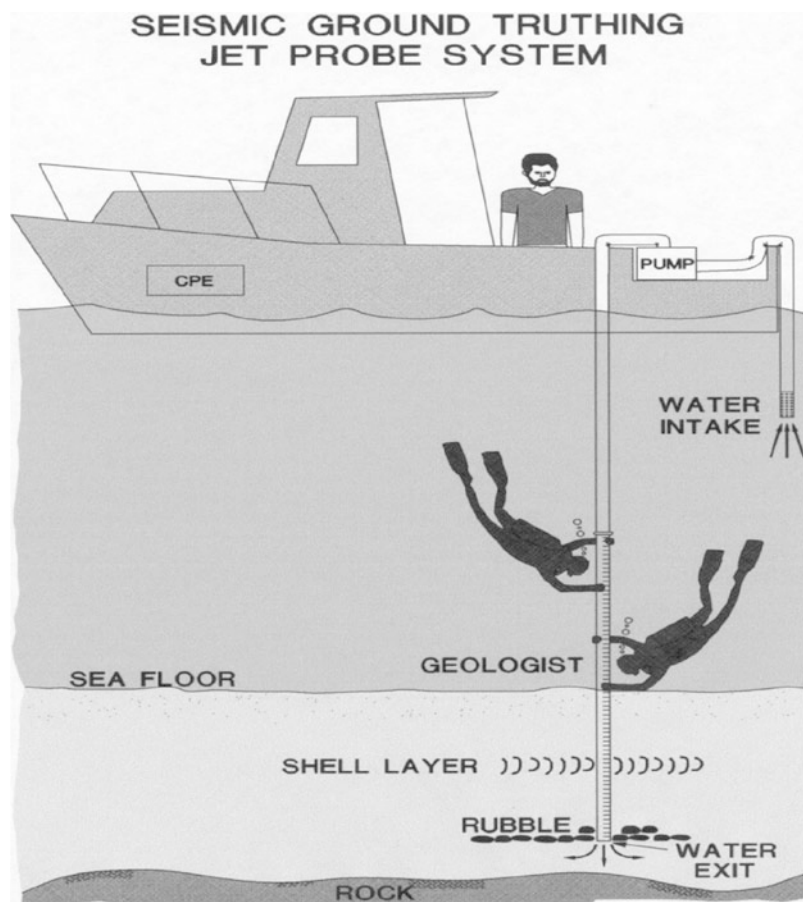
in a day depending on water depths, weather, and sea conditions (Andrews 2002).

Geologists who are proficient in SCUBA diving, operate the jet probe by penetrating a graduated 7 m water pressure pipe into the ocean bottom and making observations as it passes through the sediment layers. The geologist is on the bottom and the support diver stays at the upper end of the probe to hold it upright against the current (Fig. 4). The support diver also observes the turbidity level changes from above as silt is washed out of the probe hole (becoming suspended in the water column) during penetration of the seafloor. The geologist on the bottom observes the graduated scale on the probe and by the "feel" of the objects it encounters, makes mental notes of the depths of each change in texture, which are afterwards incorporated into the field log (cf. Fig. 2). An experienced diver-geologist can distinguish layers such as shell, rubble, sand, peat, clay, and rock. The probe is jettied to the total length of the pipe (usually 7 m) or until it encounters a layer that it is unable to penetrate. On the Florida Gulf coast, karstified limestone formations on the inner continental shelf (e.g., Evans et al. 1985; Hine et al. 1998) usually limit jet probe penetration because sand deposits are less than 7 m thick. Nearly contiguous offshore

sand ridges (described above), which are related to ebb-tidal deltas, and paleo barrier island, beach, and surf zone environments constitute the major source of beach renourishment sand on the central coast of west Florida. Jet probes, which are ideally suited to quickly and economically measure the thickness of thin sand deposits (i.e., <7 m in thickness), are therefore widely used in this geomorphic setting to determine the isopachs of shelf deposits that often occur in the form of sand sheets (shoals) or low ridges.

To obtain sediment samples from various depths, wash borings are obtained by the following methods. The geologist, who directs the jet probe into bottom sediments, takes two sample bags that are labeled "mid-depth" and "bottom of hole." The support diver, near the water surface, takes one sample bag labeled "surface sample." The probe is driven to its total depth of penetration, point of refusal (caused by hard layers, large floater, or bedrock) or maximum length of pipe. If that depth is 6 m, for example, the probe is pulled out and a second hole is probed to a depth of 3 m, 2–4 m up current from the first hole. The geologist pulls the first probe and the support diver signals the boat to haul the probe to the surface. The geologist takes a sample of the material that has formed a mound (spoil pile) around the probe hole and

Jet Probes, Fig. 4 Schematic diagram showing the procedure for jet probing bottom sediments on the seafloor. Note that the portion of pipe that penetrates into the sediments contains graduated marks so that sediment thickness can be accurately determined. The geologist-diver works at the lower level near the seafloor and is proficient at estimating the nature of the materials probed by the "feel" of the pipe as it penetrates to refusal or reaches the end of the pipe



places it in the “bottom of hole” sample bag. A subsample of the material forming a mound around the second (shallower) hole is placed in the bag labeled “mid depth.” The support diver, after the jet probe is hauled to the surface workboat, swims toward the bottom while moving against the current at about 2 m from the probed area (first two holes) and obtains an undisturbed “surface” sample from the bottom.

The subsamples removed from the washout mounds provide a representative bulk sample of the material that the probe passed through and which was jetted to the surface by the water pressure in the pipe. Materials comprising the washout mounds are deposited in the reverse order of the actual stratigraphic layers in the bottom sediments. Wash borings tend to have inherently low silt contents because the fine-grained particles, which have lower specific gravities than larger grains, tend to remain in the water column as suspensions. The denser grains thus settle annularly in a mound around the jet probe. Suspension of fine-grained materials (typically silt plus clay and possibly organics) produces turbidity clouds in the water, which are quickly dispersed by currents. It is essential for the near-surface support diver to estimate changes in the turbidity plumes issuing from the jet probe so that the presence of fine-grained sediments is not underestimated from inspection of the heavier wash borings that quickly settle out of the water column. With experience, estimates of fines at different depths can be surprisingly accurate. Even though these samples (spoil from jet probing and estimates of fines) are extremely useful in the selection of areas for additional investigation, they are not meant to supplement or replace vibracores when defining borrow sites.

Upon returning to the surface workboat, both the diver-geologist and support diver immediately relay their underwater jet probing observations (i.e., depths of penetration, nature of the materials in different layers, and levels of turbidity that were associated with different depths) to the second onboard geologist who records this information in a permanent logbook. The descriptions relayed to the logbook should also include information that is relevant to characterization of the seafloor surface viz. sand ripples, algae, sea grass, surface rubble, or other observations. This information is often used to assist in the interpolation of sidescan sonar data. The sand samples are cataloged and notes on the texture (grain size) and color are recorded. To prepare jet probe data for inclusion in reports, data that were recorded in the logbook are digitally entered into a jet probe log that is formatted in a manner similar to vibracore logs (see ► [Vibracore](#)). Sand samples are sieved to determine grain size and compared, in both wet and dry states, with a Munsell soil color chart. Representative samples are archived in small sample bags for presentation, reporting, and review. An example of data compilation for a jet probe survey is summarized in Table 1, which shows the classification of the jet probe, local relief of the

surrounding seafloor, penetration of the probe, grain size, turbidity, and other relevant observations. Classification of the jet probe is important to interpretations of the survey because a single probe does not determine the viability of a deposit. The classification reported here is not universal, just an indication of what kind of system might be devised to show the resource potential of a probed area. Categorization of the “area of influence” for a single jet probe is comprehended by the application of “buffers,” whereas multiple jet probe penetration defines a deposit. The buffer concept for jet probes represents an area that expands or contracts, depending on local sedimentary and geomorphological conditions. A sand sheet deposit will, for example, have a larger buffer zone around each jet probe because these kinds of deposits tend to be rather uniform over relatively large distances. The buffer around a probe on isolated sand ridges or in valley fills (i.e., drowned fluvial valleys, delta distributaries, tidal channels) will be a smaller zone because these kinds of deposits have limited lateral extents and conditions of sedimentation change in relatively short distances away from the probe. Local relief of the seabed in this area, increased by the presence of sedimentary bodies, is an indication of penetration depth for jet probes. Figure 5 demonstrates the observation with a fairly good correlation coefficient ($R^2=0.3935$). Once a survey is completed and the full range of parameters is appreciated and incorporated into an electronic database (see below), each jet probe is back classified so that it indicates the location of potential sand resources to be further investigated by refined geophysical (seismic and sidescan sonar) and geotechnical (vibracore) methods. Each jet probe is thus classified into one of five categories that range from unsuitable to a high potential for use. The categories are defined in Table 1 and it is important to note that application of the buffer concept in a spatial context on maps permits the recognition of sands (and the associated seafloor texture as seen in sidescan sonar images, three-dimensional bathymetric models, or isobathic expression of geomorphic units) that are potentially useful in beach replenishment projects. Grain size is determined by granulometric procedures from the subsamples collected by the geologist manning the jet probe. Turbidity is reported as estimated in the field and is a rough guide to the percent silt in the deposit (which is accurately determined later in the laboratory). Other observations included in Table 1 refer to the presence of rock fragments (e.g., limestone rubble, coral fragments), whether grain sizes fine or coarsen upwards or downwards, or any other property that should be noted.

Modern jet probe surveys are interfaced with advanced navigational software and differential GPS that make it possible to incorporate data into GIS database systems in such a way that reconnaissance-level surveys can be easily updated by new information and to facilitate efforts to groundtruth geophysical and geotechnical surveys. Table 1 is an example

Jet Probes, Table 1 Summary of field results for a jet probe survey off Charlotte County southwestern Florida

Jet Probe (#)	Category (rank) ^a	Relief (ft)	Penetration	Grain size (mm)	Turbidity (estimated)	Observations and notes
1	3	N/Av ^b	19	0.18	M to H ^c	Four feet of silty sand with clay balls on top/no refusal
2	4	N/Av	12	0.24	H; H to M	Three feet of sand with silt/clay on top
3	5	N/Av	14	0.42	L to M	Slightly fining upwards
4	5	N/Av	20	0.31	L to M	Slightly fining upwards/no refusal
5	4	5	5	0.23	L to M	Fining upwards (1 ft. layer), rock on bottom
6	4	5	7	0.35	H	Two feet of silty sands on top, 1 ft. rubble on bottom
7	3	3	4	0.23	M to H	Fining upwards, rock on bottom, not well-defined ridge
8	2	4	3.5	0.19	M to H	Homogeneous, rock on bottom
9	1	4	3	0.16	H	One-foot thick layer of silty sand on top
10	2	5	7	0.17	L to M	About 4% silt and very fine sand
11	3	4	3	0.33	M to H	Fining upwards, relatively thin
12	1	4	3.5	0.17	H	Silty sands, 0.5 ft. of rubble
13	2	5.5	4.5	0.17	M to H	One layer of silty sand on top
14	2	4.5	7.5	0.15	L to M	Fining upwards, very fine sediments
15	3	5	4	0.71	H	Fining upward, shell fragments
16	3	4.5	5	0.19	M	Finer than 0.2
17	2	5	8	0.14	H	Silty sand, too fine and H turbidity
18	4	5	9	0.16	M to H	Finer top, 0.5 mm visual estimate of bottom sand
19	3	6	16	0.14	M to H	Finer top, coarser on bottom (0.23 visual estimate)
20	2	1	3	0.22	H	Finer silty sand in top layer
21	4	6	8	0.23	M to H	Fining upwards, 1' silty sand on top, 0.5' rubble bottom
22	4	4	6	0.23	M to H	Fining upwards, at least 4 ft. of fine-medium sand
23	3	4.5	4	0.95	M to H	Shell fragments, some silt in top 2 ft
24	4	7	7	0.52	M to H; L to M	At least 4 feet of clean sand, silty sand in top 2 ft
25	3	4.5	6	0.17	L to M	Clean sediments but too fine
26	5	3.5	5	0.23	L to M	Five feet of clean sand
27	5	4	8	0.37	L to M	Eight feet of clean sand, fining upward
28	4	6	6	0.19	L to M; M to H	Somewhat finer-grained than neighbors
29	5	3	5	0.29	M; M to H	Four and one-half feet of clean sand
30	2	2	2	0.41	M to H	Missed the top of the ridge
31	3	3.5	3	0.28	M	Missed the top of the ridge
32	3	3	3	0.22	H	Limited penetration, high turbidity
33	3	4	4	0.19	M to H	Fine sand, M to H turbidity levels
34	1	1	0	-	-	Trough before reef gave a "false-ridge" impression
35	1	3	3	0.16	H	Fine sediments, high turbidity, limited thickness
36	1	2	2	0.22	H	Limited thickness, high turbidity
37	2	3.5	3	0.34	H	Limited thickness, high turbidity
38	2	3	5	0.2	M to H	Silty sand on top, relatively high silt %
39	3	3	4	0.54	M	Shell fragments, 3 ft. of clean sand
40	3	3	3	0.6	H	High turbidity, limited thickness
41	1	1	2	0.19	H	Trough after outcrop, limited thickness and silty sediments
42	3	3	4	0.6	M to H	Limited thickness, shell fragments, 1 ft. layer of rubble 3 to 4'
43	4	2.5	5	0.57	L to M	One foot of rubble from 4 to 5
44	5	3.5	6	0.5	M	Five and one-half feet of coarse gray sand
45	2	3	2	0.22	L to M	Missed depositional area
46	5	3.5	11	0.25	M	Homogeneous sediment distribution

(continued)

Jet Probes, Table 1 (continued)

Jet Probe (#)	Category (rank) ^a	Relief (ft)	Penetration	Grain size (mm)	Turbidity (estimated)	Observations and notes
47	4	N/Av	14	0.17	L	Coarsening upward
48	2	N/Av	15	0.13	L	Sediments <0.15, but coarsening upwards and low turbidity
49	2	N/Av	15	0.13	L	Sediments <0.15, but coarsening upwards and low turbidity
50	2	N/Av	15	0.15	L	Sediments <0.15, but coarsening upwards and low turbidity

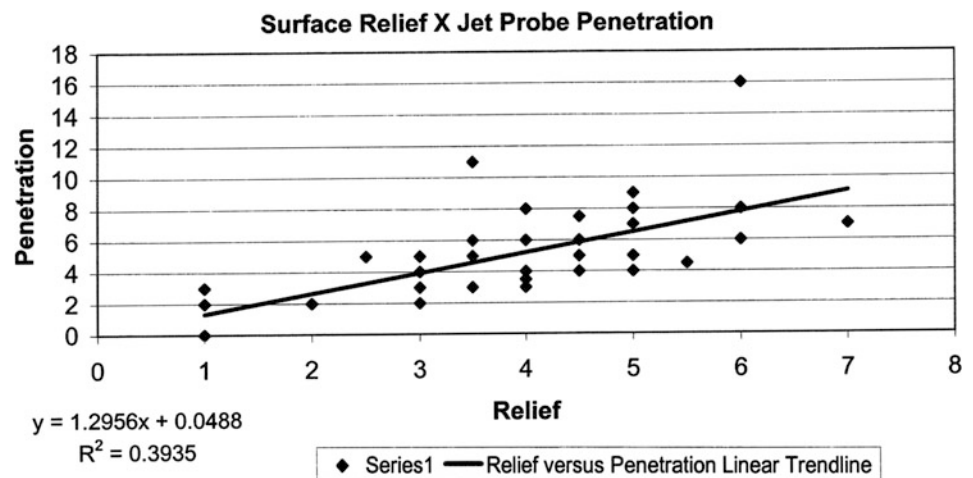
Table lists major criteria that are useful for the interpretation of sand deposits. Information that is summarized in tubular form assists in the identification of materials that are suitable for beach replenishment.

^aBuffers divide jet-probed sedimentary deposit thickness into four categories based on sand quality and dredging capabilities, as follows: (1) Unsuitable: Deposit is less than 0.5 m thick, or mean grain size <0.17 mm, or there are high levels of turbidity during jet probing. (2) Marginally useful: Deposit is less than 1 m thick, or mean grain size is <0.2 mm, or the deposit is thicker with larger grain sizes but there is high turbidity, or the presence of silty sands or rubble layers. (3) Conditionally usable: Deposit thickness is between 0.5 m and 1 m with relatively good quality sediments containing a mean grain size greater than 0.2 mm, but there are limiting factors such as limited thickness of sand bodies, or high turbidity levels, or good penetration but sediments analyzed >0.2 mm but visual description on other layers was <0.2 mm mean diameter of sand grains. (4) Potentially useful: Deposits thicker than 1 m but less than 1.5 m and sand grain sizes are between 0.2 mm and 0.25 mm; there is moderate to low turbidity. (5) High potential for use: Deposit is more than 1.5 m thick and sand grain size is more than 0.25 mm, there is moderate to low turbidity

^bN/Av = not available

^cThe terms low, medium, and high are relative estimates of silt content based on visual interpolation of turbidity plumes

Jet Probes, Fig. 5 Linear regression analysis for a subset of jet probes collected offshore Naples, Florida, showing that jet probe penetration in jet-increases creases with increasing local relief of sediments on the seafloor. The survey area included a series of sand ridges with intervening troughs. The troughs contained a thin veneer of sand (generally less than 0.5 m) over limestone bedrock whereas the sand ridges had a local relief up to at least 2.5 m (units on the graph are in feet)



of the kind of jet probe-related information that can be extracted from GIS databases or queried for specific purposes. GIS analysis rings also facilitate querying procedures that can locate potential targets for sand mining activities and that can also point to areas where sediment texture and compositional information is insufficient to make reliable conclusions as to the presence of quality-sand sources. On the basis of jet probe data and other information, specific sand deposits are identified for detailed field surveys. Summary reports are usually prepared in a composite GIS framework, that is, in an electronic database and maps that help estimate sand volumes and approximate costs for detailed investigation of each potential borrow area, based on characteristics such as grain size and distance from the beach nourishment site on the shore and dredging suitability.

Conclusion

Jet probes are used to obtain information related to surficial sediment thickness on land and in shallow coastal waters. On land, jet probes may be powered by air or water pressure forced through a length of pipe. Jet Probes represent a good low-cost survey method for reconnaissance surveys on the sea and lakebeds. They are applicable to archaeological investigations and stratigraphic studies of thin sedimentary sequences, but it is in the search for beach-compatible sediments on the inner continental shelf that they find greatest use.

As a coastal resource tool, jet probes are often underutilized because researchers tend to use more sophisticated survey methods in the belief that greater value is received from greater expenditure. Jet probes are, however,

an economical way of determining not only the thickness of sedimentary bodies but also their composition, grain size, compaction, and inclusions of rock fragments or other materials. When used collectively with a defined area as a specialized reconnaissance survey method, jet probes provide groundtruthing for geophysical, geotechnical, geological, and geomorphological interpretations of the seabed sediments. The main drawback for jet probing is that operators need to acquire sensitive skills for interpreting the “feel” of probe penetration. With some practice, however, geologist-driver operators can become proficient estimators of the various parameters that are normally associated with jet probe surveys. The most widespread application of jet probing is in coastal sand searches because increased knowledge of offshore sand resources is required for beach nourishment projects. Maximum water survey depths for jet probing are limited to about 30 m and the depth of penetration to the length of pipe that is easily handled underwater, usually about 7 m. For practical considerations, the minimum operating depth in water is about 1 m. As the search for sand resources intensifies, due to increasing erosion of beaches and coastal land loss on protective barrier islands and shoals, jet probes will increasingly serve as comparatively inexpensive procedures for evaluating seabed sediments on inner continental shelves.

Advanced geophysical and geotechnical procedures are essential for the accurate definition and location of sand resources on the inner shelf; however, these resources can be optimized if backed up or verified by low tech and less costly efforts, such as jet probing, that are essential to the efficiency and economic success of offshore sand searches. Advancements in positioning and navigation software and hardware that can be interfaced with GIS systems in the field permit analysis of spatial data associated with the jet probes in a timely fashion that increase survey efficiency and applicability. Although combinations of modern marine exploration techniques have contributed to the cost-effectiveness and success of sand search investigations, they are of reduced value if they are not accompanied by logical and rational planning of surveys in accordance with local geology and geomorphology.

Cross-References

- [Archaeology](#)
- [Coastal Sedimentary Facies](#)
- [Monitoring Coastal Geomorphology](#)
- [Nearshore Geomorphological Mapping](#)
- [Offshore Sand Banks and Linear Sand Ridges](#)
- [Offshore Sand Sheets](#)
- [Shelf Processes](#)
- [Shoreline and Coastal Terrain Mapping](#)
- [Surf Zone Processes](#)
- [Vibrocure](#)

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Karst Coasts

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Karst coasts are defined as coasts made of calcareous rocks and showing distinctive geomorphic characteristics related to their lithologic nature (Jennings 1985). In some cases, these original features are directly due to seawater action and marine weathering. Halokarst is a word sometimes used to designate such erosion forms (Fairbridge 1982). In other cases, karst coasts correspond to the exposure by coastal retreat or to the drowning by the postglacial transgression of subaerial or underground karst features, which are actually the result of limestone solution in continental environments (Trenhaile 1987).

Coastal Karstification

Seawater Corrosion

The word corrosion, introduced by A. Guilcher, includes different chemical, physicochemical, and biological processes operating on carbonate-rich rocks in coastal environments and resulting in specific erosional features.

The ability of seawater to dissolve calcium carbonate has long been a contentious issue. Nowadays, the most widespread opinion is that no real evidence favoring solution has been produced. It appears that coastal seawater is saturated or oversaturated with calcium carbonate and data-showing solution at night, through emission of carbon dioxide by green algae living in pools, is not wholly convincing. However, recent work seems to indicate that chemical erosion has been underestimated (Miller and Mason 1994). It is now proved that, at least in tropical environments, undersaturation

of inshore waters may occur at night with respect to calcite and at any time with respect to aragonite and high magnesian calcite, accounting for some 10% of the erosion in coralline limestones (Trudgill 1976).

Bioerosion (q.v.), a term for the removal of rock by the direct action of living organisms, is generally acknowledged to play the greatest role in the development of coastal corrosional features, not only in the tropics where an enormously varied marine biota live on calcareous substrates, but also in higher latitudes (Kelletat 1988). Algae are probably the most important erosive organisms, both in the intertidal and the supralittoral zones. Endolithic cyanophyta are boring organisms that actively contribute to the rock destruction. Fungi and lichens also are effective rock borers. Grazers consuming the microflora, such as the gastropods *Littorina* and *Patella*, can cause mechanical rasping of rock surfaces, which have been weakened by the penetration of endolithic algae. The chiton *Acanthopleura* has hard teeth that enables it to erode resistant limestones. Borers are responsible for excavations into the substratum. Penetrating habits of *Lithophaga*, *Lithotrya*, *Cliona* are frequently mentioned. *Lithophaga* acts through mechanical boring facilitated by acid secretion, which causes a softening of the rock. The sponges pertaining to the genus *Cliona*, which are able to bore microscopic to macroscopic excavations in limestones, play a particularly important role in the disintegration of rock substrates. Also worms, such as *Polydora*, may be active borers in calcareous substrates. Biological erosion is of great significance on the limestone coasts, which may justify the term of biokarst which has been proposed for the resulting forms (Spencer 1985).

In the supralittoral zone, physicochemical processes operate jointly with bioerosion. Spray action, implying wetting and subsequent drying, leads to salt crystallization and causes rock disintegration. Eolianites are especially prone to such kind of weathering.

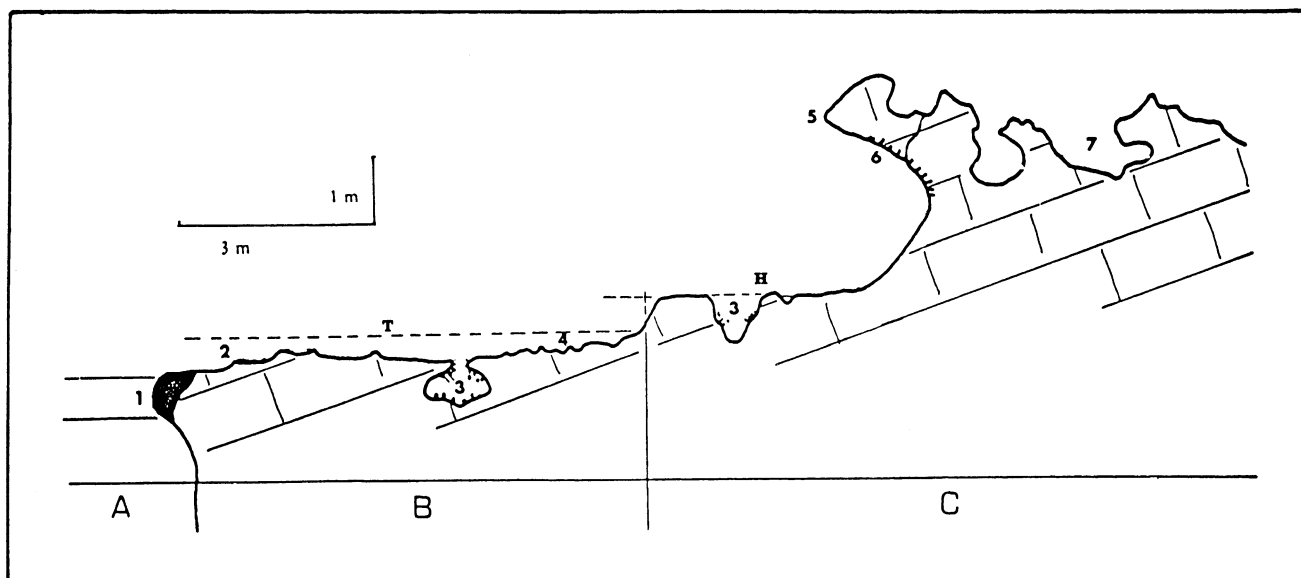
The rate of sea corrosion in calcareous rocks has been measured in a great number of sites all over the world. A figure of about 1 mm per year may be considered as an average rate.

Roland P. Paskoff: deceased.

Shore Platforms

Bare erosional platforms (see entry on “► Shore Platforms”) may be found on low carbonate-rich rocky coasts where waves are not supplied with clastic tools of allochthonous origin, which enhance their mechanical action. Figure 1 represents a typical profile of a corrosional shore platform from the Mediterranean which can be used as an illustration of a littoral karst in subtropical, low wave energy, and microtidal conditions. There is a general zoning of forms between low

water mark and the area reached only by spray. The main feature of the midlittoral (intertidal) zone is represented by a platform, several meters in width, which is called trottoir (*q.v.*), a French word for sidewalk (Fig. 2). In its upper part, the platform shows pools with overhanging edges and, lower down, wide shallow pools with flat bottoms, the so-called “vasques,” which are separated by low, narrow, continuous, sinuous rims made of residual rock or built by calcareous organisms. Seawards, before terminating abruptly in a small



Karst Coasts, Fig. 1 Schematic profile across a limestone shore platform in the Mediterranean, after R. Dalongeville (1977). A, infralittoral zone; B, midlittoral zone; C, supralittoral zone; H, elevated trottoir indicating a

Holocene relative sea level higher than the present one; T, trottoir; 1, vermetid ledge; 2, shallow pool of vasque type; 3, pool with overhanging sides; 4, alveoles; 5, overhanging cliff; 6, vermiculations; 7, lapiés and pools

Karst Coasts, Fig. 2 Typical trottoir developed into an upper Pleistocene eolianite, northern Israel coast. (Photo R.P. Paskoff.)



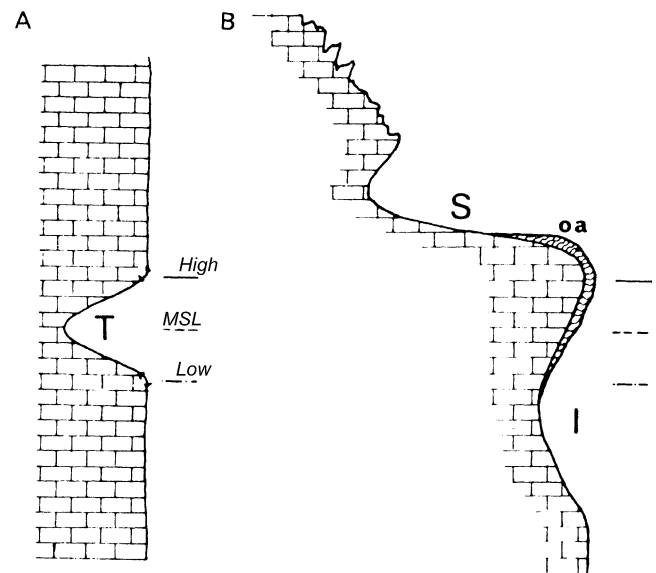
Karst Coasts, Fig. 3 Upper Pleistocene eolianite affected by lapiés and pools in the supralittoral zone, south of Tangier, Morocco. (Photo R.P. Paskoff.)



vertical infralittoral cliff, the platform is often characterized by an overhanging ledge made by vermetids (*Dendropoma*) and calcareous algae (*Neogoniolithon*). Sometimes, a fossil trottoir, a few tens of centimeters above the active one, characterized by crater-shaped pools, is found, pointing to a higher relative sea level during the Holocene. Further up, in the supralittoral zone where spray is acting, above an overhanging cliff pitted by alveoles and vermiculations, jagged and sharp lapiés are found (Fig. 3). They are separated by deep pools with overhanging rims. In fact, wave energy, tidal range, and mainly seawater temperature are the most important parameters which explain a great variety of shore platforms in carbonate rocks. In cool temperate regions, the trottoir is unknown, whereas in warm seas the presence of deep notches and protruding visors is noticeable (Guilcher 1953).

Notches

Coastal *notches* (*q.v.*) are indentations due to lateral cutting by sea corrosion, which is particularly active on calcareous-rich rocks in tropical waters (Fig. 4). Cliffs, low rocky shores, or wave-thrown large boulders on coral reefs may be affected by notches, which are good indicators of sea-level position, especially where the tidal range is low and the coastal environment sheltered. Site exposure is the most important factor and two main types of notches are to be distinguished (Pirazzoli 1986): (1) tidal notches, in relatively protected sites and cut in the intertidal zone, which are relatively narrow; (2) surf notches in exposed sites and cut above high tide level. In the case of the first type, when the undercutting is well developed, being 2–3 m deep, the notch roof often



Karst Coasts, Fig. 4 Sea corrosion notch profiles, after P.A. Pirazzoli (1986). A, sheltered environment; B, swell exposed environment; I, infralittoral notch; oa, organic accretion; S, surf notch; T, tidal notch

forms an overhanging rock ledge, called the visor. Deep tidal undercut on low stacks and isolated blocks may result in mushroom-shaped rocks. Surf notches have a distinctive morphology due to water turbulence and spray action. Organic accretion around high tide level by calcareous algae and vermetids protects the substrate calcareous rock and inhibits sea action, meanwhile erosion proceeds above, forming an asymmetric notch with generally a short roof and a developed floor which eventually may form a bench

protruding seaward. Simultaneously, bioerosion, is responsible for the development of another notch around low tide level. Consequently, double notches are not necessarily evidence of a relative sea-level change.

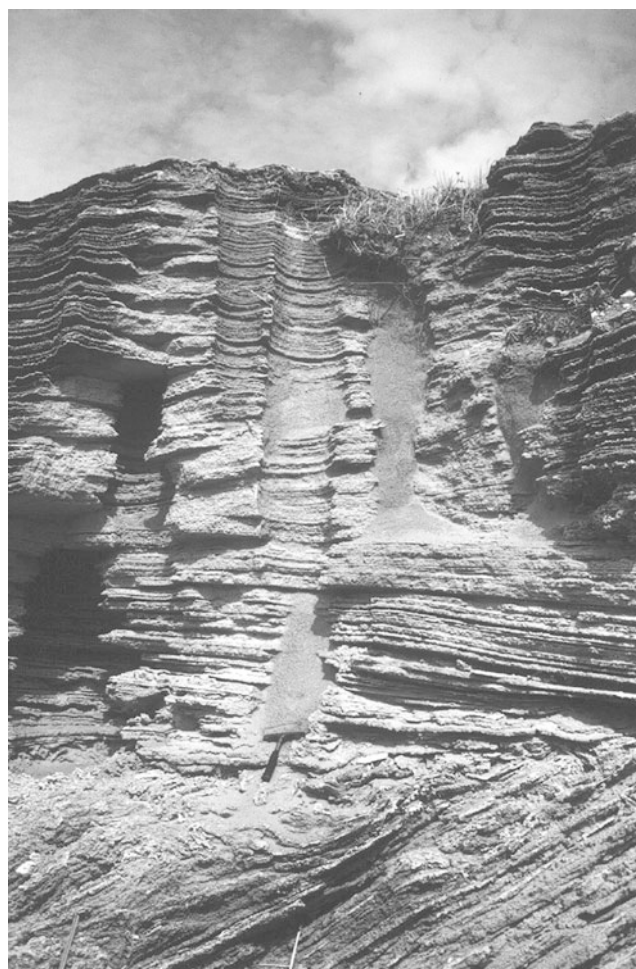
Solution Pipes in Eolianites

Calcareous eolianites or eolian calcarenites are subaerially cemented paleodunes (see entry on “► [Dune Ridges](#)”). They characterize many coastal regions which are semiarid at present and they may also be found in those which were semiarid at some stages during the Pleistocene, a period of important sea-level oscillations and climate changes. They have been described from the Mediterranean, the Canary Islands, South Africa, southern and western Australia, Bermuda. Diagenesis by continental waters of the carbonate-rich sand deposits, largely composed of fragments of marine organisms, took place in a vadose environment and generally includes dissolution of unstable aragonite and high-Mg calcite, and precipitation of relative low-Mg calcite.

Such eolianites may be affected by piping, which produces tubular underground conduits (Fig. 5). For instance, in northern Tunisia, conspicuous pipes, 20–40 cm in diameter and a few meters deep have been reported (Paskoff 1996). They are cylindrical in shape, taper vertically downward, and occur in aggregated clusters. They show a red coat which is a hardened calcitic crust and are filled by red silts and sands, sometimes strongly cemented. In the early literature, it was suggested that pipes represent the pseudomorphs of former tree stumps buried under advancing sand dunes. They were also simply explained as random solutional features corresponding to points where percolating water happened to converge and caused localized subsurface dissolution. A more appealing explanation was recently put forward for the Bermuda pipes by S.R. Hervitz (1993) who suggested that the cylindrical vertical conduits are the products of the stemflow of tree species capable of acidifying intercepted rain water and funneling large quantity of it down their trunks. The result is a subsurface dissolution forming pipes, which extend vertically downward through eolianites.

Exposed and Submerged Terrestrial Karst on Coasts

There are coasts where karst landforms can be followed from the land into the sea practically without any substantial modification. A conspicuous example of such a situation is given at Along Bay, near Haiphong, in northern Vietnam, where a typical tower karst, developed in a humid tropical environment, has been submerged and makes up an archipelago of islands and islets in a shallow sea (Fig. 6).

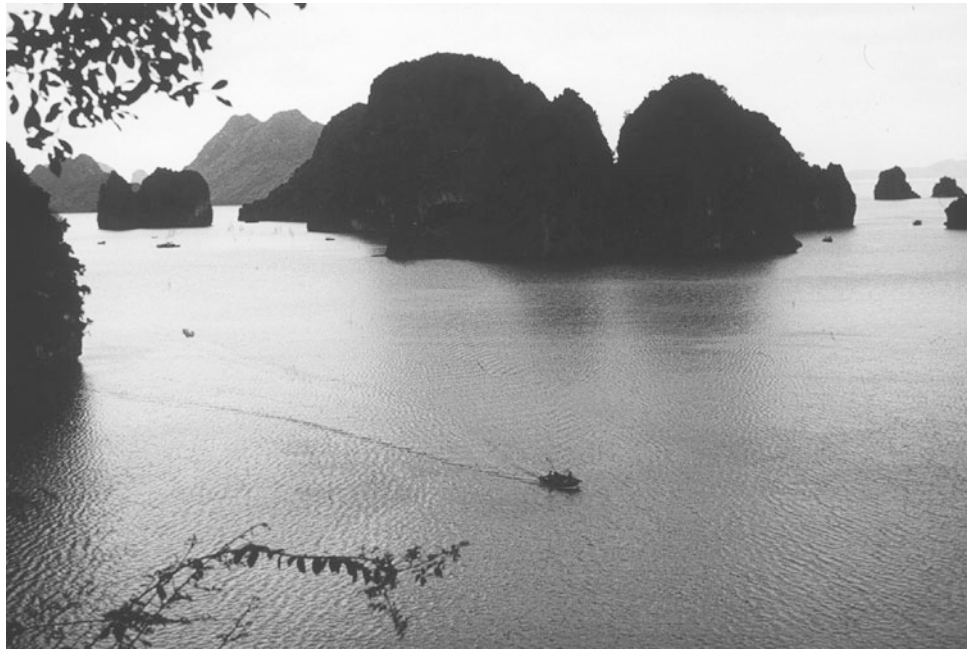


Karst Coasts, Fig. 5 Vertical section of a solution pipe on an active sea cliff developed into an upper Pleistocene eolianite in northern Tunisia. (Photo R.P. Paskoff)

Exposed Karstic Landforms

As a result of coastal erosion, continental karstic features may have been exposed and modified by marine processes. For instance, the nearly vertical chalk cliffs (see entry on “► [Chalk Coasts](#)”) of the Normandy coast, in France, show typical solution forms of terrestrial origin, which have been revealed by the shoreline retreat. There are examples of inherited karstic caves debouching in the face of the cliffs. Others have been captured by sea caves developed at the foot of the cliffs under mechanical wave attack. Coastal erosion has also exposed deep cylindrical hollows, which originally formed as solution pipes. The famous arch and pillars at Etretat owe their existence to marine action in a highly karstified area. The crest of the Normandy chalk cliffs shows a crenulated appearance, which is due to dry valleys, locally called *valleuses*, now hanging as a result of coastline recession. Near Cascais, in central Portugal, on a rock coast, bare and deeply developed lapies are exposed in the upper mid-littoral zone because decalcification red silts which once

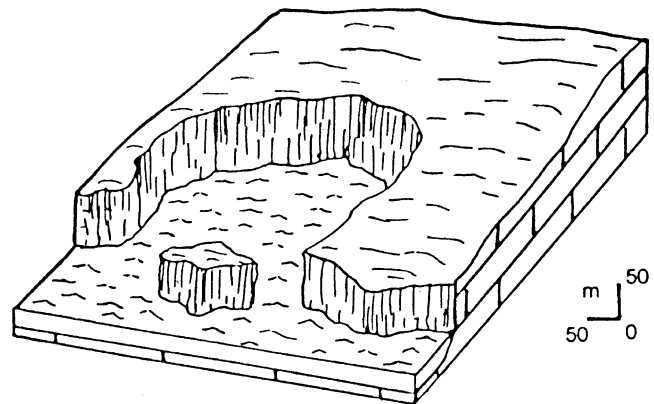
Karst Coasts, Fig. 6 Islands and islets corresponding to a submerged tropical karst, Along Bay, northern Vietnam. (Photo M. Paskoff.)



covered them have been stripped off by swash denudation. In Australia, west of Melbourne, the Port Campbell coast illustrates the effects of marine erosion processes on a highly karstified area of Miocene limestones. Sinkholes and caves have been cut by wave action (Baker 1943).

Drowned Karstic Landforms

The interaction between continental processes, sea-level variations, and wave erosion is well illustrated by subaerial or underground features which developed in limestone terrains, such as enclosed depressions or caves, and are now invaded by the sea. In this respect, Malta gives conspicuous examples of what could be called marine sinkholes, originally karstic features developed in coralline limestones (Paskoff and Sanlaville 1978). On Gozo Island, Dewra Bay is a semi-circular cove, measuring 340 m in diameter, which results from marine erosion breaching a doline wall whose eastern half alone has been preserved and whose bottom is now largely occupied by the sea. An islet, Fungus Rock, is the last remnant of the destroyed western wall (Fig. 7). Qwara, an immediately neighboring landform, is an identical circular sinkhole, 400 m in diameter and 70 m deep, bounded by vertical walls, which remained unbreached, but has been nevertheless partially inundated through a karstic gallery now connecting the open sea with the depression. Blue Grotto, in the southern part of the main island, corresponds to another kind of cove which is due to the sea eroding into a large cave system and causing subsequent roof collapse. In Asturias, Spain, different types of periodically or permanently flooded dolines, by salt water, have been described (Schülke 1968). Marine charts of the subsident Dalmatian coast,



Karst Coasts, Fig. 7 Dweyra Bay and Fungus Rock, Gozo Island, Malta: a semi-circular cove created through sea invasion of a doline. (After Paskoff and Sanlaville 1978)

in Croatia, clearly show broad enclosed basins of polje type, which are completely submerged (Baulig 1930). The extended Novigrad Bay is regarded as a polje invaded by the sea. The freshwater lake of Vratna, on the island of Cres, appears to occupy the floor of an uvala, a large depression resulting from the coalescence of sinkholes, which was deepened when sea level was lower than now and became a permanent lake as a result of the postglacial transgression. In the vicinity of Marseille, in France, giant lapiés have been identified at a depth of 40 m in the Veyron Bank. In the same area, the partly submerged Cosquer cave, which was discovered in 1991 and became notorious for its exceptionally nice paintings and engravings dating back to the upper paleolithic period, has only one entrance located at 37 m under

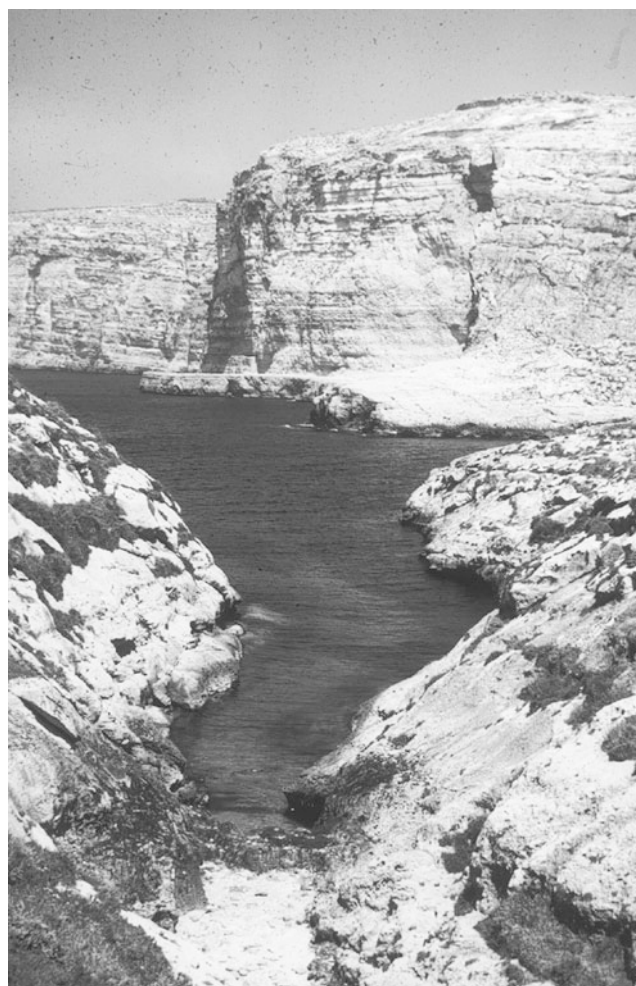
present sea level. The limestone littoral of Provence, as others calcareous coasts in the Mediterranean, shows many active submarine springs or resurgences, for instance the one at Port-Miou, near Cassis, east of Marseille, located at a depth of 12 m.

Calanques or Calas

Calanque from the French coast of Provence or *cala* from the Balearic islands in Spain are words which designate a narrow, short, and drowned valley with steep sides, developed in limestone terrains and continued inland by a dry course. Several types have been identified, as in Malta, for example (Paskoff and Sanlaville 1978). Some are of ria type. They were deepened during glacial periods of low sea level and pluvial climate, which facilitated the cutting of deep ravines by stream action, subsequently submerged in their lower portion by postglacial transgressions, as it was the case with the Holocene sea-level rise (Fig. 8). The amazing pattern of digitate calanques, which makes the site of the capital city, Valletta, one of the finest anchorages in the world, corresponds to a branching valley system whose deeply and extended drowning resulted from a marked subsidence in this area. Other calanques are linked with continental karst processes. They appear, at least partly, related to the formation of subterranean caves and conduits by freshwater solution of limestones along faults. Later on, wave erosion opened these cavities and the rushing of marine water caused the roofs to collapse. Such a case raises an important question about the extent to which karst features can develop below the sea without any change in the relative level of land and sea. If the response is positive, calanques may form without a marine transgression being necessary.

Karstified Coral Reefs

According to some authors, barrier reef and atoll morphology (see entry on “► Coral Reefs”) is fundamentally karst induced (Purdy 1974). This theory, sometimes called “the karstic saucer theory” (Guilcher 1988) and still in discussion, states that the shape of such reef forms derives from antecedent and horizontal coralline platforms, which were emerged and modified by subaerial solution processes. It is logical to assume that all reefs which are today at or near sea level were emergent during the glaciation periods. Being emerged, they became karstified. Rainfall and percolating water action is supposed to be more rapid toward the interior of the calcareous platforms than around their steep edges where runoff is rapid. The result is a saucer-shaped surface dissolution with a central depression and a peripheral raised rim interrupted by ravines through which a part of the water escaped, the other part percolating through the limestone. Subsequent submergence of karsteroded platforms started a revival of coral growth and the ramparts resulting from previous subaerial karstification became barrier reefs or atolls.



Karst Coasts, Fig. 8 Typical calanque in the island of Gozo, Malta. It corresponds to a partially drowned valley, with steep sides, cut into coralline limestones. (Photo R.P. Paskoff.)

The karstic saucer theory, which is not in contradiction with the Darwin's subsidence theory of coral reef formation and can be combined with it, is supported by geomorphic observations. Floors of many lagoons are characterized by numerous upstanding pinnacles or knolls which rise from various depths and have living coral on their surfaces. They are generally interpreted as remnants of karstification during low sea-level glacioeustatic phases and compared with the tower karst developed in rainy tropical environments. Lagoons may also show pits which are considered as submerged sinkholes or blue holes. At Mayotte, a Comoro island, the bottom of the lagoon shows several enclosed depressions with steep sides, 60–70 m deep, lying at approximately 20 m below the surrounding floor. These features are obviously the result of a subaerial karstification. Mataiva, an atoll in the Tuamotu archipelago, has a reticulated lagoon formed by a network of about 70 pools of varying sizes, with an average depth of some 10 m,

separated by shallow ridges. This strange honeycombed pattern whose exact conditions of formation remain uncertain is thought to derive from a complicated evolution during which tropical phases of karstification are supposed to have occurred when the structure was emerged. So, it appears that the sea-level lowerings related to the Pleistocene glaciations have in many cases determined a karstification of preexisting coral structures.

Cross-References

- Atolls
- Bioerosion
- Chalk Coasts
- Coral Reefs
- Eolianite
- Notches
- Rock Coast Processes
- Shore Platforms
- Trottoirs
- Weathering in the Coastal Zone

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Klint

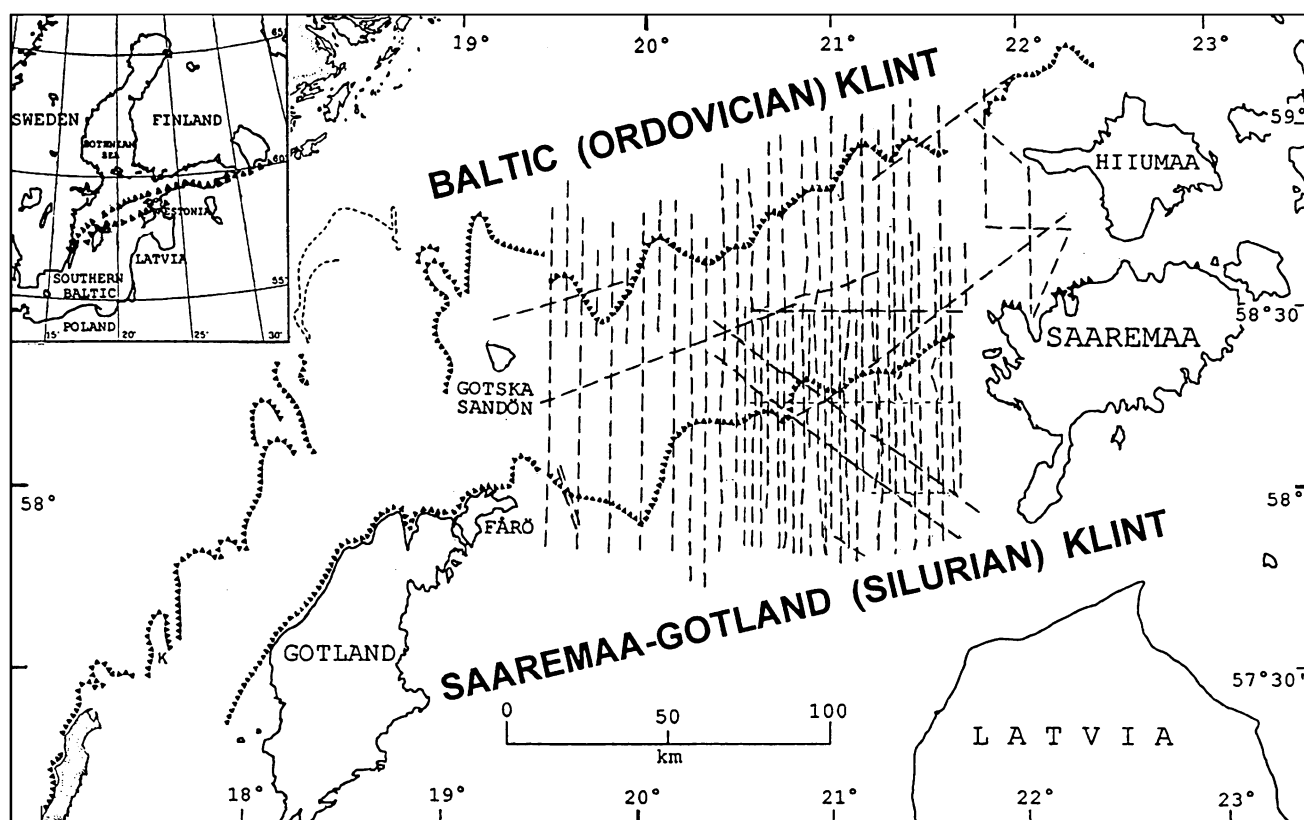
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The term klint, widely used in countries around the Baltic Sea, was originally a Danish and Swedish word synonymous with *klev*, signifying an escarpment in sedimentary rocks. Usually it comprises a line of marine abrasion or ancient (pre-Quaternary) fluvial erosion scarps. In Swedish, the word occurs synonymously with *griek*, signifying a type of hollow formed by karst weathering (Martinsson 1958), and also means mountaintop or bioherm (coral reef) hillock. Also the term *fjällglint* is used for rock formations in a Scandinavian mountain ridge (*fjällkedjan*). In German literature the word *glint* is preferred and in such form was adopted also in Estonia, the classical klint area (Tammekann 1940). The word in English is often recorded as *clint*, in Russian *glint* (more rarely *klint*), in Latvian *glints* (*klints* in Latvian means *rock*), and in Lithuanian *klintas* is used. The corresponding Estonian word is *paekallas*; for separate klint lobes and promontories the word *pank* is also used.

Distribution and Structure

The most well-known klint in the Baltic Sea area is the Ordovician. It consists of an almost continuous, but indented and lobated arc, from the western coast of the Island of Öland (Västra Landborgen) in Sweden over the Baltic Sea via the north coast of Estonia to Lake Ladoga in Russia (Fig. 1). This monumental escarpment, up to 56 m high (at Ontika in Estonia) and 1,200-km long, is called the Baltic Klint (Fig. 2). Its basal part consists of Cambrian rocks, dominated by sand- and siltstones and soft “Blue clays.” The hard crest layers of the klint, however, which primarily cause the steepness of the escarpment, consist of Ordovician limestones (Orviku 1940). In the westernmost area, on the Island of Öland, the klint crest is developed exclusively in Middle Ordovician limestone beds. A submarine Ordovician klint, as a morphological feature, has been identified on sea charts and with geophysical sounding methods (Fig. 1). The



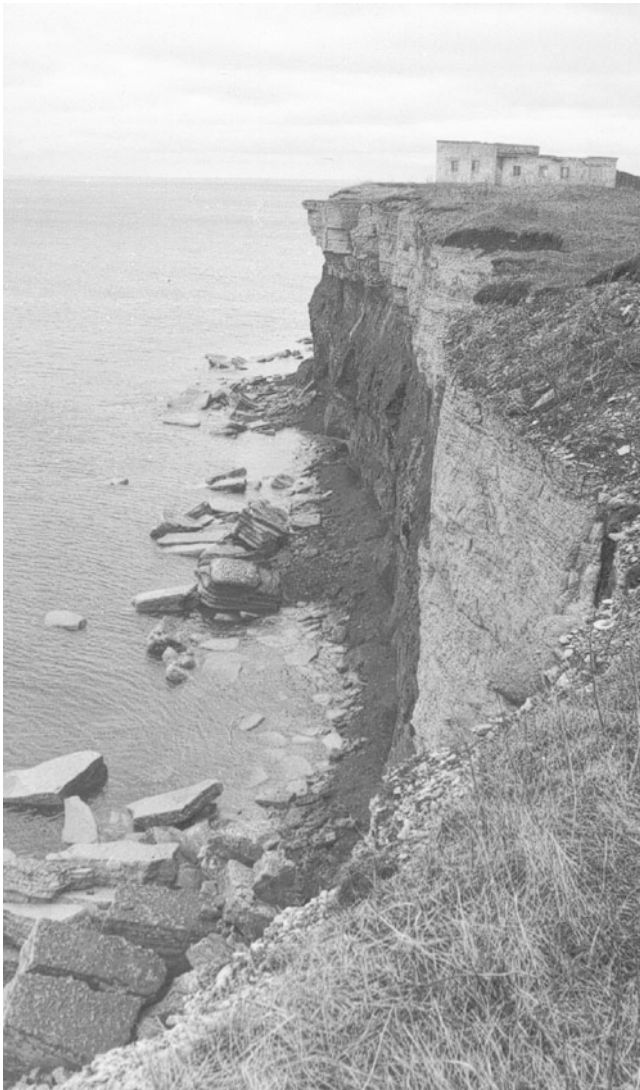
Klint, Fig. 1 The distribution of klint (black triangles) in the Baltic Sea area. Compiled by R. Vaher, based on the data published by A. Martinsson (1958) and I. Tuuling (1998). Dashed lines mark seismic profiles

Baltic Klint, called North-Estonian Klint in the Estonian mainland, is dissected by river valleys forming over 20 picturesque waterfalls (Aaloe and Miidel 1967). A submarine Vendian–Cambrian klint line in the scientific literature indicates not only the escarpment line, but also the border between the Fennoscandian Shield and the Russian Platform.

To the north of the Baltic (Ordovician) Klint on the seafloor, Vendian and Cambrian scarps and to the south step-by-step Silurian, Devonian, and Carboniferous klints crop out. The Saaremaa–Gotland (Silurian) Klint in Silurian strata forms an arc of more or less lobed scarps along the northwest coast of Gotland Island in Sweden, on the seabottom and on the north coast of Saaremaa Island in Estonia, following to Central Estonia. The westernmost part of this klint arc limits the shelf on which two Karlsö islands, Store Karlsö and Lilla Karlsö, are situated. The larger part of the klint crest on Gotland reaches between 40 and 50 m above sea level. On Saaremaa, the height of cliffs is up to 22 m at Mustjala and on the Estonian mainland rarely more than 10 m. Devonian and Carboniferous klints in the contemporary topography differ from each other in scarps of different height and length.

Formation

Some klints in the Baltic Sea area began to develop during the Late Silurian and pre-Middle Devonian continental period (Puura et al. 1999). Later the prolonged pre-Pleistocene erosional–denudational processes were of utmost importance, operating upon the main features of structural changes in the Precambrian crystalline basement and the sedimentary cover. The latter has a gentle southward inclination (6–18'). Due to the tectonic uplift, new areas were influenced by the lateral river flows forming questa-like topography with steep northern and gentle southern slopes. The influence of the old drainage systems upon the klint formation is evident. The Ordovician klint system developed throughout the entire length by the erosion of the soft, mainly Cambrian sandstone and clay strata, overlain by much more resistant limestones, which determined the retreat of the klint and caused the relative steepness of the feature. The bottommost part of the Saaremaa–Gotland Klint also consists of softer rocks in mainland Estonia, for example, from marls of the Jaani Regional Stage, overlain by the Jaagarahu limestones, a large part of which consists of reefs, which cause the dissected and lobated appearance of the klint. Zigzag contour lines in many places are dependant on tectonic joints.



Klint, Fig. 2 The Pakerort Klint on Pakri Peninsula, northwestern Estonia. (Photo by A. Miidel.)

During the ice ages the klints were influenced by glacial erosion and, after the retreat of the continental ice, the Ordovician and Silurian klints were strongly changed by wave action of the Baltic Sea. This influence was different, because different parts of the klints rose above the sea level at different times (Orviku and Orviku 1969). If the Ordovician

Klint at Ontika in northeastern Estonia was under the influence of waves during the Baltic Ice Lake more than 11,000 years ago, then the Island of Osmussaar in northwestern Estonia appeared above the sea level only some 2,000 years ago.

The term *klint* is used more and more not only in the Baltic Sea area, but also in North America and western Europe for the mentioning of steep escarpments with monoclinal bedding of sedimentary rocks. There is also a specific type of klint (glint)-line lakes in Norway and Scotland, which were formed between the ice and escarpments.

Cross-References

- [Changing Sea Levels](#)
- [Cliffed Coasts](#)
- [Cliffs, Erosion Rates](#)
- [Europe, Coastal Geomorphology](#)
- [Rock Coast Processes](#)

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Last Interglacial

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Definitions

Coastal sediment intervals that formed during the “*sensu stricto*” LIG highstand or “Last Interglacial Maximum” were first recognized in the mid-nineteenth century in the nearshore and brackish-lagoonal Eemian beds of the Netherlands. Shackleton (1969) identified the interglacial optimum with the MIS 5e isotope substage, while the four younger and cooler MIS 5 stades and interstades were assigned either to the *sensu lato* LIG or on paleoclimate considerations, to the earliest interval of the Wisconsinan glacial stage. In addition to the most frequent stratigraphic terms, Eemian in Europe and Sangamonian (earlier, “Sangamon”) in North America, a host of regional and local designations have been applied to LIG across the globe (Otvos 2015).

In terms of insolation variability, LIG does not resemble the interglacial history of the present Holocene interglacial as closely as the much longer and warmer MIS 11 interglacial between 424 and 374 ka (Candy et al. 2014). However, the worldwide abundance of LIG coastal and nearshore deposits render them critical benchmarks in identifying and calculating penultimate marine highstand levels, the distribution of coastal and nearshore depositional facies, and contributions by relative sea-level changes that have resulted in highly variable elevation values.

LIG highstand positions provide unique benchmarks for calculating sea-level changes. Attempts were made to establish local and regional relationships between vertical crustal movements, caused by both isostatic and tectonic processes. Complete reconstruction of a purely *eustatic* global mean sea-level record of the geological past is unattainable (*q.v.*, entry *Eustasy*

and Sea Level, present Encyclopedia). However, the definition of global ice-equivalent (“glacio-eustatic”) sea levels, clear of localized isostatic, tectonic, and other secondary influences, remains a critical objective (Murray-Wallace et al. 2014, 2016). By utilizing geostationary satellite systems, in routine use only since 1992, the sequential monitoring of mean global ocean level changes may become available in the future.

MIS 5e Climate Indicators: Coastal, Nearshore, and Offshore

Microfossils, fossil molluscs, and other macrofaunas in marine terrace and barrier deposits often provide strong evidence for warmer-than-present LIG conditions in various climate belts and, under favorable circumstances, some record of temperature decline at the onset of cooling stages (Hamilton and Brigham-Grette 1991; Bardají et al. 2009; Muhs et al. 2012b, 2014). In Antarctic Ice Dome C, the interglacial maximum was defined as occurring between 132 and 115 ka ago as a single peak and assumed by certain authors to be the warmest interval in the last 800 ka (Murray-Wallace et al. 2016). However, based on microfossil and IRD (ice-rafted debris) evidence, precipitation and temperature fluctuations and consequent sea-level changes within substage MIS 5e are known from several locations (e.g., Brewer et al. 2008; Bauch et al. 2011; Schwab et al. 2013; Dutton et al. 2015). These accounts provide detailed subdivision of highstand sequences characterized by warmer peaks and intervening cooler valleys. At northern sites, IRD-free interglacial conditions of peak warmth mark early LIG times between ~129 and 126 ka ago (Bauch et al. 2011); elsewhere between 122 and 119 ka (Müller and Sánchez-Gómez 2007; Bauch et al. 2011; Rovere et al. 2016). Indicative of near-current climate conditions, Key Largo Formation corals flourished in South Florida at least between 123 and 114 ka B (Muhs et al. 2011). An oxygen isotope curve by Lea et al. (2002), constructed by decoupling temperature effect by using Mg/Ca ratios, closely approximates various sea-level records.

Sea Level Markers

Predominantly three categories of relative sea-level markers have been used to establish the relative position of the paleo-sea level in relation to current sea level. (1) Abrasional bedrock markers include erosional marine benches, rock terraces and platforms, ledges, notches, and visors. (2) Unconsolidated inter- and supratidal shore deposits form terraces underlain by a variety of inshore and nearshore deposits, capped by strandplains and eolian dune fields. When identifiable by sedimentary structures, the interface between intertidal beach fore-shore and overlying foredune or backshore sand or gravel lithosomes may serve as local paleo-sea-level indicators. In regressive strandplain sequences, inter-ridge swale floors may constrain the maximum elevations of underlying intertidal deposits. If erosion did not remove their upper intervals, the elevations of estuarine or marine sediments exposed in the present land surface and the estimated water depth at deposition may offer such an opportunity (Murray-Wallace et al. 2016). Deposits of intertidal brackish or salt marshes and mangrove forests, particularly on microtidal shores provide high-precision benchmarks. (3) The optimal minimum depth range for reef-crest coral species that extended to very shallow water habitats (e.g., *Porites*, *Acropora palmata*) provides an approximate paleo-sea-level position (e.g., Muhs et al. 2011). Bedrock terraces and benches often are veneered by unconsolidated nearshore deposits and attached or detrital colonial or solitary corals. In seaward-tilted shore terraces, as on the Pacific coast of the Americas, the elevation of their inner

margin at the landward edge of a terrace is the best estimate of paleo-sea level. The seaward tilt angle of the terrace surface is referred to as the *shoreline angle*.

Terrace and Coastal Barrier Deposits: Stacked Versus Stair-Stepped Sequences

Tectonically apparently stable shores, as the Florida Keys and Cayman Islands, produce stacked terrace and sediment sequences, whereas combination of the two categories, occur found on Curaçao Island (Fig. 1) where there are constructional reefs but also wave-cut platforms. Stair-stepped terrace or barrier lithosomes and topography (Fig. 2) develop along tectonically relatively less active coasts (Barbados, Haiti, the US South Atlantic coast, California's Channel Islands; Muhs et al. 2011, 2012b). The Channel Islands include no barriers. Discrete barrier and terrace units with longer sediment records may result from slow uplift. Vertical accretion of individual barrier units that develop by this process increases barrier separation from laterally adjacent younger and older barriers that also will be fewer in numbers. On the other hand, when copious sediment supply is ensured by the recycling of old and new sediment sources, a stacked barrier lithosome and other stacked sediment sequences may contain several interglacial, glacial, and interstadial stratigraphic intervals (e.g., Brigham-Grette et al. 2011). Humid climate that fosters thicker vegetation cover, as well as gently inclined offshore bottom topography and

Last Interglacial, Fig. 1 MIS 5e marine terrace (foreground), directly underlain by MIS 7 terrace deposits. MIS 11 marine terrace: middle ground. High terrace: far background. Curaçao Island, Dutch West Indies (Muhs et al. 2012a) (Photo courtesy Dr. Daniel R. Muhs)



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Fig. 2 Composite MIS 5 (lowest) terrace and tectonically uplifted stairsteps of three or more older marine terraces (middle ground and right), Eel Point, Pacific shore of San Clemente Island, California (View to N. Photo courtesy of Dr. Daniel R. Muhs. Reference: Muhs et al. 2014)



wide dissipative beaches favor this style of eolian sandy shore barrier aggradation, with locally abundant marine-derived bioclastic carbonate fragments. Narrow inner continental shelves and relatively higher-energy wave climate environments with erosional episodes favor fewer discrete barriers (Bateman et al. 2010; Muhs et al. 2012a). Erosional terraces form in exposed bedrocks, overlain by poorly sorted, often fossiliferous sand and gravel (Muhs et al. 2012a).

LIG Age Range and Dating Methods: Climate Variations

Dating of LIG shorelines and their deposits can be direct, through numerical methods, or inferred using depositional facies and/or distinctive fossil assemblages (foraminifers, coccolithophores, corals, and mollusks) that are dated independently elsewhere. Uranium series has provided some of the most reliable numerical ages for marine terrace deposits, but the method can be applied only to fossil corals as this is the only marine organism that takes up uranium from sea water during growth (Edwards et al. 2003). Cosmogenic ^{10}Be dating has also been used in marine deposits (Martinod et al. 2016). In dating eolian and shallow marine sand deposits and marine deposits, OSL (optically stimulated luminescence) dating is routinely utilized in eolian deposits and has also been applied to shallow marine sand deposits (Lamothe 2016). The relative extent of amino acid racemization (AAR) for aspartic acid, glutamic acid, phenylalanine, and valine

reflects progressive diagenetic changes for the different amino acids and has been applied widely to fossil mollusks. The extent of AAR in fossil mollusks is most useful for ascertaining relative ages from D/L ratios (Wehmiller et al. 2013; Murray-Wallace et al. 2016). Lithological variations in finer- or coarser-grained sediment sources of the IRD (ice-rafted debris) render recognition and climate interpretation from drillcore sequences occasionally problematical.

In general, the LIG time interval is constrained between 132 and 113 ka ago and, at higher latitudes, between 128 and 118 ka ago. While sea level probably peaked between 129 and 126 ka, because of its time-transgressive nature and dependence on geographic factor, such as the degree of continentality in and near coastal regions, shifting climate zones (Šeirienė et al. 2014; Govin et al. 2015; Irval et al. 2016), and ocean current systems (e.g., Bardaji et al. 2009; Blanchon et al. 2009; Bauch et al. 2011), the time range of and subdivisions within LIG may differ considerably between locations (Otvos 2015). In southern Greenland and Iceland, the 126–122 ka peak warmth interval included significant amounts of IRD in North Atlantic deposits. IRD-free deposits of the interglacial peak, dated to 122–119 ka, followed (Bauch et al. 2011; Irval et al. 2016). Displaying low IRD counts, a two-step cooling trend at certain sites lasted until ~113 ka (Müller and Sánchez-Goni 2007; Govin et al. 2015; Irval et al. 2016).

High tectonic uplift rates left a detailed record of sea-level and climatic variations during the last interglacial. In southern Calabria, Italy, 11 minor oscillations were documented during

MIS 5e and four more during MIS 5c/5a. This chronology includes a higher amplitude sea level rise at 128 ka, 122 ka, and 116 ka ago (Dumas et al. 2005). The closing phase of the interglacial, ~122 to ~113 ka ago, tends to be less clearly defined than its start. In certain warmer regions, such as the Mediterranean of southern Europe, the LIG climate optimum extended into MIS 5d. In keeping with its diachronous character, LIG warmth in Southern Italy persisted until 109 ka. In the north, summers during the marginally glaciated Weichselian-Wisconsinan MIS 5d and 5b substages cooled only slightly, but the winters became more severe (Helmens et al. 2012; Šeirienė et al. 2014).

LIG Maximum Sea Levels: Tectonic, Isostatic, and Other Relative Sea-Level (RSL) Components of Global Sea-Level Changes

Tectonism Versus LIG Relative Sea Levels (RSL)

RSL values vary greatly in magnitude, generally being most intensive on collisional, including subduction coasts, with lower values in strike-slip (transcurrent) zones and spreading centers. Minimum uplift occurs at central plate mantle hotspots. In a large compilation of worldwide data, Pedoja et al. (2014) reemphasized that active plate margin tectonism, long known from New Guinea, leads to the most rapid vertical displacement. Seaward-tilted wave-cut abrasional terraces, platforms, and benches on the north Chilean coast occur at +25–60 m elevations (Martinod et al. 2016). Major uplift, possibly interspersed with faulting-related subsidence in Santa Catalina Island (Castillo et al. 2012; Schumann et al. 2012), characterized the western margin of North America (DeDiego-Forbisa et al. 2004; Muhs et al. 2012b, 2014). In comparison with the adjacent California Channel Islands, the uplift of certain mainland coastal areas was much more rapid (Fig. 13 in Muhs et al. 2015). Simms et al. (2016) attempted to separate isostatic aspects from tectonic components of the crustal uplift. In contrast, a Saudi coral terrace near transform-faulted Gulf of Aqaba has experienced relatively moderate (16–20 m) uplift since the concluding phase of the last interglacial, 120–108 ka (Manaa et al. 2016) (*q.v.*, Encyclopedia entries *Eustasy and Sea Level*; *Gulf Shorelines, Last Eustatic Cycle*).

Glacio-hydro-isostatic (GIA) Effect on LIG Sea Level

“Near-field” glacio-hydro-isostatic topographic effects involve areas near glaciated regions, while “intermediate-” and “far-field” changes influence sites that are more remote, respectively, farthest from (past or present) glacial centers (*q.v.*, Encyclopedia entries *Eustasy and Sea Level*; *Gulf Shorelines, Last Eustatic Cycle*). Isostatic rebound, following unloading of the disintegrated Laurentide Ice Sheet, accounted for the moderate uplift of the extensive shore-

parallel terrace complex between Virginia and NE Florida along the southeastern US coast (Winker and Howard 1977; Potter and Lambeck 2003). Part of the terrace sequence occurs in the uplifted Cape Fear Arch region.

The uplifted alluvial, in small part estuarine blanket of the late Pliocene Citronelle Formation in the NE Gulf of Mexico coastal plain, is marked by prominent erosional scarp topography and frequently deep fluvial incision. Following intensive pre-LIG coastal erosion, only the youngest Pleistocene coastal units survived on this coast (Otvos 2001, 2015, 2018). In northern Siberia, a major isostatic rebound during MIS 5e has raised deposits of the Karginsky boreal transgression to +140 m elevation. The transgression followed disintegration of the large Arctic Byrranga ice sheet of preceding glacial stage MIS 6 (Möller et al. 2015).

Karst Isostasy and Relative LIG Sea Level

In combination with variations in the degree of karstification, differential subsidence took place in Belize, Central America, along a series of continental margin fault blocks composed of LIG carbonates (Gischler et al. 2000). Karst terrain degradation and cavity collapse have been linked to isostatic crustal adjustment through mantle flow toward the karstified region (*q.v.* *Eustasy and Sea Level* chapter).

Passive Margin Coastal Terraces and Barriers

South Atlantic US Coast: Type Example of “Stair-Stepped” Pleistocene Siliciclastic Terrace Sequence

The terrace surfaces are underlain by a variety of nearshore, intertidal, backbarrier-lagoonal, and eolian dune deposits. Traditionally, the age of the Pamlico, lately of the Princess Anne Terrace in Virginia and the Carolinas, was considered the Last Interglacial (Doar and Kendall 2014). The coastal plain terrace stairsteps formed during an extended period of relatively slow uplift, moderate erosion and substantial seaward progradation. In the absence of a closed-system uranium content in coral samples and due to assumed wholesale erosion of LIG highstand terrace lithosomes, several crucial chronological issues remain unresolved (Wehmiller et al. 2004, 2013; Markewich et al. 2013).

Northern Gulf Coast: Uplifted Passive Margin Coast, Single LIG Barrier Complex, and Without Stair-Stepped Pleistocene Terrace Sequence

On the NE Gulf coastal plain, the elevated upland position of the alluvial, partially estuarine late Pliocene Citronelle Formation and absence of all but the Last Interglacial inshore and coastal-marine sediment sequence testify to an extensive Pleistocene uplift and concurrent wholesale, if episodic erosion of all Pleistocene brackish-inshore and coastal barrier deposits predates the LIG shoreline. Extensive, occasionally

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Fig. 3 Outcrop of LIG Gulfport Formation, subtidal–intertidal deposits, transitional upward from Biloxi Fm, overlain by cross-stratified LIG and Wisconsinan dune sand interval. Bottom (sea level) to -5.0 m, *Ophiomorpha*-laced shallow subtidal white sand; 5.0 – 7.2 m, dark humate-impregnated sand overlain by (7.2 – 16.2 m) undifferentiated cross-bedded LIG and Wisconsinan eolian sand to top of outcrop. Intracoastal Canal, NW Florida, $30^{\circ} 20' 43.1''$ N, $86^{\circ} 01' 38.8''$ W (Photo by Dr. J. Bryan 2013)



thick, mostly muddy, and locally sandy estuarine-lagoonal-to-full salinity open marine deposits of the Biloxi Formation mark the LIG transgression and highstand between SW Texas and Florida Peninsula beneath the Pleistocene coastal plain. The Ingleside chain is ~ 160 km long (Otvoš 1991, 2005, 2015). Thick shallow subtidal-to-supratidal dune sands overlie the Biloxi Formation in the ~ 200 -km-long, generally continuous Gulfport Formation barrier trend east of the Pearl River and in its Ingleside Barrier chain correlative along the central Texas coast (Otvoš and Howat 1996). Several OSL dates originated in the barrier lithosomes (Otvoš 2005; Simms et al. 2016). Cross-stratified LIG and Wisconsinan-age eolian sand beds overlie the interglacial intertidal-shallow subtidal deposits (Fig. 3).

Interglacial Relative Peak Sea Level on NE Gulf Coast

The interface between horizontally bedded, partially humate-impregnated subtidal sands that include an abundance of shallow subtidal *Ophiomorpha* burrows of the *Lepidophthalmus*/*Callinassa* decapod genus, along with rare molluscs, other trace fossils, and the overlying eolian dune sands, suggests a local interglacial relative sea level of at least $\sim +3$ m (Otvoš 2001; Figs. 6 and 10 in Oivanki and Otvoš 1994). Possibly also constraining the upper limit of LIG (Last Interglacial) relative sea level, the inter-dune swale floor level in the overlying Gulfport ridge plain occurs at $\sim +3.3$ m elevation. Assuming a 1 – 3 m depths of estuaries in which the brackish Biloxi facies was laid down (e.g., Fig. 7 in Otvoš 2001), these values conform with the approximate $+1$ m elevation of the estuarine Biloxi facies at its landward

limit beneath the coastal plain (Otvoš 2015, 2018). The depth is also compatible with coastal drillhole and outcrop data. Gulfport barrier intervals that include ghost shrimp burrows in Gulfport barrier lithosomes range from modern shallow subtidal levels in SE Mississippi and the Florida Panhandle to $\sim +5$ m elevation on the NW Florida coast (Bryan 2013 and Fig. 3). The elevation difference may reflect temporal and spatial variations in relative sea-level positions during barrier strandplain progradation. Spatial differences in post-LIG isostatic uplift rates may have also played a role.

In terms of age, appearance, fossil content, and passive continental margin setting, the Gulfport barrier lithosomes appear nearly identical to outcrops in the 620 -km-long, 7.7 -m-high Barrier III, along the similarly low-microtidal passive-margin coast of southeastern Brazil (Fig. 3 in Tomazelli and Dillenburg 2007). In contrast to the Gulf Coast, ghost shrimp burrow fossils, indicators of shallow subtidal depositional environments here extend to 7 m above sea level.

Search for a Global (“Eustatic and Quasi-ice Equivalent”) LIG Sea Level with Minimal Tectonic, Isostatic, and Other RSL Influences

In contrast with near- and intermediate-field locations, the most important sites for LIG sea-level studies are “far-field” locations at the greatest distances from formerly ice-loaded areas. There is a long history of tectonic stability. Dated sites with interpreted $+6$ – 10 m sea-level stands occupy a granitic

micro-continental site and Precambrian cratons. The sites are far removed from mantle hotspots in the Seychelles Islands, respectively, Brazil (Tomazelli and Dillenburg 2007) and South Africa (Carr et al. 2010). The +10 m RSL topographic elevation, associated with $+7.6 \pm 1.7$ -m relative sea level stand may have also been driven by localized *mantle flow* upwelling, linked to late Quaternary crustal processes. It has been reported from a stable, glacio-isostatically “far-field” setting, remote from former glaciation centers. The corals in the elevated Seychelles reefs formed after 125 ka (Dutton et al. 2015; Vyverberg et al. 2018). Habitat depth-based assumptions on 6–8 m paleo-sea levels were also made with regard to LIG in South Florida coral reefs that similarly lack a history of significant tectonic uplift. Glacio-hydro-isostatically, they occupy an intermediate-field position (Muhs et al. 2011). The unusually high reef elevations contrast sharply with the +2 to +5 m interglacial sea levels, at several sites worldwide. Numerous dates originated in well-preserved, *in situ* lagoonal shell fossils in a very ancient South Australian craton and, at other locations, deemed historically similarly stable (Murray-Wallace 2002; Murray-Wallace et al. 2016; Pan et al. 2018).

In facing this quandary and in departure from earlier modeling practices, Creveling et al. (2015) considered that “the peak eustatic value adopted for MIS 5e in most previous studies (i.e., +6 m) is likely incorrect. In addition, the local peak sea level during MIS 5e is characterized by significant departures from eustasy due to glacial isostatic adjustment in response to both successive glacial–interglacial cycles and excess polar ice-sheet melt relative to present day values. . . . Sea level during MIS 5e may depart from eustasy by 2–4 m, or more.” The authors challenged the assumption that a globally averaged sea-level value can be used to correct local sea-level record and emphasized that an observed (local) MIS 5e highstand marker should not be corrected by the peak global eustatic value. The highstand value reflects the combined effects of GIA at a given location. “This encompasses the full deformational, gravitational, and rotational perturbation in sea level driven by the redistribution of ice and ocean mass, and introduces significant geographic variability (i.e., departures from eustasy) into the correction.”

Summary

The MIS 5e (locally, MIS 5e-5d)-age Last Interglacial (LIG, ~132 to 109 ka B.P.) shore markers represent the most widespread and best preserved Pleistocene marine highstand indicators and related deposits globally. The duration, chronological start and end points of the interglacial are diachronous and depend on numerous variables, including geographic position, degree of continentality, exposure to warm and cold ocean currents, proximity to major ice sheets, and many other factors. A variety of biogenic, stratigraphic,

geomorphic, and isotope-based climate indicators suggest significantly warmer conditions during the LIG climate optimum. Shoreline elevations above present sea-level display significant variations even at short distances and range between different sites from a few meters to 140 m, possibly even higher. Variable relative sea-level components, including plate tectonic, tectonic, karst effect, and glacio-isostatic, and other influences masked the original eustatic (approximately ice-equivalent global sea level) signals. The search for LIG sea levels is driven to a large extent by the desire to establish the extent of the ice- and ocean-covered areas during that critical interval by factoring in local relative sea-level components, such as tectonic uplift or subsidence, vertical and horizontal isostatic movements, karst, and other effects. By comparing these with Milankovitch theory-based astronomical components of the LIG (see Shackleton 1969) and the Holocene interglacial, approximate and reasonable predictions for future global, regional, and critically important local sea-level changes may become feasible.

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transgression. Sometimes it is named the *Holocene transgression*, but this is a mistake as the Holocene Epoch began at about 11 ka BP.

Characteristics

The total volume of water in the world's oceans exhibits a nearly perfect negative correlation with global ice volume; when one increases, the other decreases. This is known as *glacial eustasy* (first proposed by Maclaren 1842). The balance between global ice volume and ocean water volume is controlled by climate. At the Last Glacial Maximum at about 24–22 ka BP (20,000 radiocarbon years BP), large quantities of water were withdrawn from the oceans and accumulated in the form of extensive continental ice caps. We may try to reconstruct past glacial volume changes by the following three means:

1. The recording of corresponding sea-level positions, which are affected by numerous other variables
2. The recording of corresponding oxygen isotope variations, which are affected by other factors, too, not least temperature
3. Volumetric estimates of corresponding ice caps, which is quite a rough method (presently stored ice in Antarctica, Greenland, and alpine glaciers are estimated in this way)

Though all three methods have their limitations and problems, there is a general agreement that the 23 ka glacial eustatic lowering was on the order of 120–130 m, as seen in sea level (e.g., Fairbanks 1989), in oxygen isotope values (e.g., Shackleton 1987), and in glacial volume (Flint 1969). According to Chappell et al. (1996), there is an excellent agreement between the coral record of the Huon Peninsula in New Guinea and deep-sea oxygen isotope records. Lambeck et al. (2014) gave a sea-level lowering of 134 m lasting from 29 to 21 ka BP.

The glacial eustatic rise in sea level as a function of the switch from ice age conditions at about 22 ka BP to interglacial climatic conditions is known as “*the postglacial transgression*.” It commenced about 22,000 cal. years BP and ended at around 6,000 cal. years. BP. This rise in sea level was neither smooth nor globally consistent (Mörner 1976; Carter et al. 1986; Rufin-Soler et al. 2014). The mean rate of sea-level rise was in the order of 10.0 mm/year (Mörner 2011).

Figure 1 gives the combined view of sea-level changes as established from the eustatic component as calculated from multiple sea-level records in northwest Europe (A), from sea-level data off West Africa (B), from coral reefs in Barbados (C), and from a compilations of multiple records (D).

Oscillations were induced both by glacial eustatic variations and by the interaction with others factors acting on sea

Late Quaternary Marine Transgression

Nils-Axel Mörner

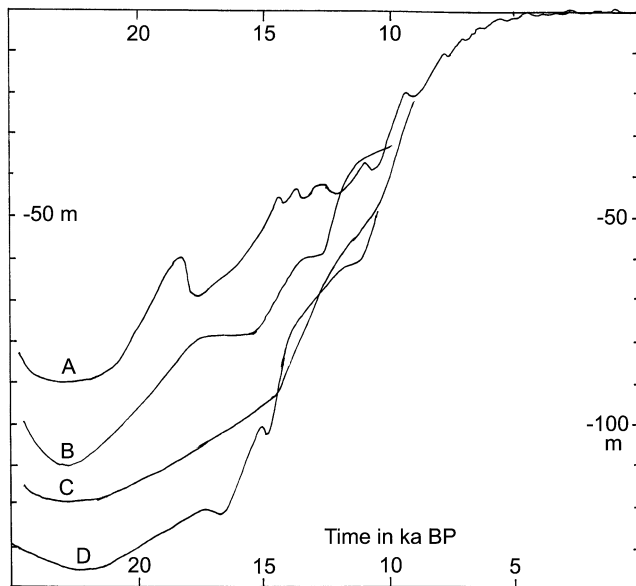
Paleogeophysics and Geodynamics, Stockholm, Sweden

Synonyms

Postglacial transgression

Definition

Late Quaternary Marine Transgression denotes the rise in level after the low sea level stand at the Last Glacial Maximum at about 22–24 ka BP. It is also known as the *postglacial*



Late Quaternary Marine Transgression, Fig. 1 The postglacial transgression illustrated by the sea-level changes recorded from (a) northwestern Europe (Mörner 1993), (b) West Africa (cf. Mörner 1993), (c) Barbados (Fairbanks 1989), and (d) a multiple-site record (Lambeck et al. 2014). Depth in meters and age in cal. year BP

level. The largest glacial eustatic oscillations were those associated with the high-amplitude climatic changes at around 16–11 ka BP ago, that is, the period including the classical climatic oscillations of the Bölling Interstadial, the Older Dryas Stadial, the Alleröd Interstadial, and the Younger Dryas Stadial (Mörner 1993).

Regional variations in amplitude and fine structures of the transgression were induced by additional variables (Mörner 2013). Those variables are deformation of the geoid relief (Mörner 1976), internal adjustment to glacial loading changes (e.g., Peltier 1998), changes in earth's rotation, and the ocean circulation system (Mörner 1995). Therefore, the actual sea-level rise after the Last Glacial Maximum differs significantly from place to place over the globe as illustrated by the atlas of Holocene sea-level curves (Pirazzoli and Pluet 1991).

Global sea level was dominated by the glacial eustatic rise in sea level up to about 6,000 cal. years BP (5,000 radiocarbon years BP). After that, it was dominated by the redistribution of ocean masses.

Cross-References

- Changing Sea Levels
- Coastal Changes, Gradual
- Coastal Changes, Rapid
- Coastline Changes
- Geodesy
- Holocene Epoch

- Sea-Level and Climate Change
- Sea-Level Fluctuations Over the Last Millennium

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Lifesaving and Beach Safety

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Lord, Lord! methought, what pain it was to drown!
What dreadful noise of waters in mine ears!
What ugly sights of death within mine eyes!
Methought I saw a thousand fearful wrecks,
Ten thousand men that fishes gnawed upon.

King Richard III
By William Shakespeare

Beaches are a major attraction for people throughout the world. In the United States, it is estimated that 85% of all tourism revenue stems from visits to coastal areas (Center for Marine

Conservation 2000). Like any natural area, however, hazards exist that can result in injury or death. Managing these hazards is key to ensuring that beaches can be safely enjoyed.

Drowning is the most serious problem related to beach and water use. Annual deaths by drowning, whether from floods, sinking ships, recreational swimming, home accidents, and other causes, easily outstrip deaths from war and terrorism. In 1997, for example, 4,051 Americans died by drowning (National Center for Health Statistics 1997). According to the Centers for Disease Control and Prevention, drowning is the second leading cause of injury-related death for American children aged 1–14 years (National Center for Health Statistics 1997). And death is not the only outcome of distress in the water. It has been found that for every child who drowns, 17 or more are treated at hospitals for complications related to near-drowning (Wintemute et al. 1998).

Beyond drowning, many other hazards exist in the aquatic environment. Spinal injury from shallow water diving occurs with unfortunate regularity, often caused by inadvertently diving into a submerged sandbar or bodysurfing over a wave and striking bottom. Scuba related illness has existed since the inception of scuba and, while the safety of equipment has improved, so has the volume of people scuba diving. Boating accidents, both recreational and commercial, are a source of injury and death. And of course, all of the common ailments related to any other physical activity occur in and around the water. All of these, and many more, have increased attention paid to accident prevention and, in particular, the provision of lifeguards at coastal beaches.

Ocean Beach in San Francisco provides a stark example. In 1998, three persons drowned on the first day of summer at this beach, which is beset by strong surf, consistent rip currents, and relatively cold water. No lifeguards were provided by the federal government, which owns the beach as part of a national park. Instead, strongly worded signs had been used in an effort to warn people away from use of the water. This was clearly not fully effective at preventing drowning.

As the summer of 1998 progressed, further drownings occurred, eventually totaling seven. There was heavy media attention and advocacy by groups like the International Life Saving Federation (ILS) and the United States Lifesaving Association (USLA) for the immediate provision of lifeguards.

Under public and political pressure, the National Park Service ultimately relented and provided preventive beach rescue services beginning at the start of the summer of 1999. During that summer, there were no drownings. Lifesaving services have continued since.

Lifesaving History

In historical terms lifesaving, as an organized response to persons in distress in the water, is young. The rescue of

shipwrecked sailors appears to have spurred some of the earliest organized efforts. China's Chinking Association for the Saving of Life was established in 1708 as the first of its kind in the world (Shanks et al. 1996). It eventually came to involve staffed lifesaving stations with specially designed and marked rescue vessels.

In the Netherlands, the *Maatschappij tot Redding van Drenkelingen* (Society to Rescue People from Drowning) was established in Amsterdam in 1767, primarily to address problems of drowning in the numerous open canals in Amsterdam. This society remains in existence today, now promoting a wide variety of drowning prevention initiatives. English lifesaving efforts began in 1774, though boating rescue operations were not initiated until 1824 (Shanks et al. 1996).

In 1787, the Massachusetts Humane Society began what was to become a lifesaving movement in the United States that evolved into the US Life-Saving Service (USLSS). The USLSS was eventually composed of an extensive national network of coastal lifesaving stations staffed by government paid lifesavers, and credited with saving over 170,000 lives. In 1915, this organization joined with the Revenue Cutter Service to become the US Coast Guard.

It was only in the late 1800s that swimming, then known as bathing, began to emerge as a widely popular form of recreation. When ocean resorts were built, in places like Atlantic City and Cape May, New Jersey, drowning quickly emerged as a problem. Various drowning prevention methods were implemented, including the use of lifelines in the water – fixed ropes to which bathers could cling.

When these approaches proved inadequate, police were assigned to lifesaving duties in Atlantic City. Eventually though, police resources became strained by this responsibility. Instead, a corps of lifeguards was employed in 1892. In Cape May, efforts began with rescue rings hung on bath-houses and the provision of dories on the beach that could be used for rescue. By 1865, hotels began hiring persons to staff the surfboats. Later, a municipal lifeguard operation was begun that continues to the present day.

Both the American Red Cross and the YMCA initiated efforts in the early 1900s to teach Americans to swim and to rescue each other when in distress. This grew into nationwide networks of swimming instruction and lifeguard training that exist today, with a focus on pools and inland beaches.

While the need for prevention and rescue services was evinced by drownings, the elemental lifesaving techniques were just that – rudimentary steps that one swimmer could use to rescue another – person to person and often without equipment. The rescue equipment, what there was of it, was adapted from other disciplines, such as the devices that had been used by the USLSS to rescue sailors from the sea.

Surfboats, similar to those used by the USLSS, were adapted for use by lifeguards to row to swimmers in trouble. They

remain in use in a few areas of the United States. The predominant method of rescue though, was by swimming to the victim.

One of the greatest difficulties for swimming lifesavers was the struggle sometimes required to overpower a panicked victim before the rescue could be completed. The line and reel (landline), was an early solution. A lifeguard would swim out to the victim while attached to the line, clutch the victim, and would be rapidly pulled back to shore by others.

This method had the advantage of quick retrieval, but there were some disadvantages too. The line produced drag, which could slow approach to the victim; it required two or more persons to operate; it was inadequate in cases of multiple rescues simultaneously occurring at different locations; and, it could become tangled. In Atlantic City, use of the line was discontinued after a lifeguard was strangled by the device. Nevertheless, it was widely used elsewhere for decades and is still in use in a few areas.

As an alternative, lifeguards fastened an eight-foot line and shoulder harness to a life ring. The lifeguard would swim out with the life ring, push it to the victim, and tow the victim to a dory or to shore. This avoided contact with the victim, but like the line and reel, the life ring created significant drag in the water.

Captain Henry Sheffield, an American with a variety of aquatic accomplishments to his credit, was touring Durban, South Africa in 1897 when he designed the first “rescue can,” also called the “rescue cylinder,” for a lifesaving club there (Brewster 1995). It was made of sheet metal and pointed on both ends, with the same over-the-shoulder harness and line as had been used on life rings. The advantage was that it moved much more smoothly through the water, providing little drag. A disadvantage was that the heavy metal and pointed ends could cause injury.

In Australia, the first volunteer lifesaving “club” was founded at Bondi in 1906. Prior to that, ordinances had proscribed swimming, but civil disobedience eventually resulted in making swimming permissible and prompting the need for lifesaving services. Surf Life Saving Australia, one of the largest volunteer organizations in the world today, grew out of the Australian tradition that began at Bondi, of voluntarily guarding the beach. Today, some Australian lifesavers are paid, but most are still volunteers.

In 1907, George Freeth was brought from Hawaii to Redondo Beach, California to help promote a seaside resort. Billed as, “the man who walks on water,” Freeth was the first person to surf on the American West Coast and is considered by many to be the first California beach lifeguard, as he made many rescues of persons in distress. According to the USLA though, it was not until the legendary Duke Paoa Kanhanamoku visited California in 1913 and introduced his redwood surfboard to Long Beach, California lifeguards, that the surfboard was adopted as a rescue tool. Later, the term “rescue board” would be coined (D’Arnall et al. 1981).

The initiation of lifeguard services has often resulted from drownings and, even today, drownings tend to spur the provision or augmentation of such services. In 1918, 13 people drowned in a single day in San Diego, California, spurring the creation of a lifeguard service that now counts some 240 lifeguards providing response to coastal emergencies 24 h a day, throughout the year.

In 1935, Santa Monica, California lifeguard, Pete Peterson, seeing a need for a device that could be wrapped around the victim for greater security in the surf, produced the first rescue tube as an inflated device. Though it was vulnerable to weather, it became quite popular among lifeguards, and even more so when, in 1964, it was made of foam rubber, hot dipped with a rubber coating. Known to many veteran lifeguards as the “Peterson Tube,” or just the “Peterson,” it is in wide use today.

The rescue tube is an excellent tool for surf rescue because the victim is less likely to become separated from the rescuer in breaking waves. It is a particularly valuable option for semiconscious or unconscious victims, though it is less useful for multiple victim rescues, since its design is intended for a single victim. Interestingly, it has now become a common tool at pools and waterparks. So a device developed by a single lifeguard for a particular environment has come to be used in all aquatic environments throughout the world.

Captain Sheffield’s sheet metal rescue “can” came to be constructed of aluminum in 1946, which lightened it substantially and allowed the ends to be rounded, but it was still heavy and itself presented a hazard. Then, Los Angeles County lifeguard Bob Burnside developed an improved rescue buoy, made of plastic, with handles on each side. It was on the beach in 1972 and greatly improved upon the ability of lifeguards to safely effect rescues, particularly in open water environs.

The “Burnside buoy” continues to be the rescue device of choice in situations where a highly buoyant and hydrodynamic float is needed to handle multiple victim rescues, since several victims can easily hold onto it at once. As a symbol of modern lifeguarding, it may have even come to eclipse the life ring.

The invention of the swim fin has changed lifesaving too. Lifeguards with fins are much faster in their swimming approach to victims and have the power to easily rescue several victims at a time, fighting the very currents that caused distress in the first place. Fins are particularly valuable in areas where the surf breaks gradually offshore and where distress may occur far from the beach. For some lifeguard agencies, swim fins are a required tool for swimming rescues and each lifeguard has a pair available at all times.

Rescue boards, a variation of surfboards and one of the original rescue devices, have been perfected to include handles for extra victims, specially designed decks for knee paddling, and lighter material. The lengths range from

3 to 4 m, with longer boards being more buoyant and faster, but heavier. Using a rescue board, a well-trained lifeguard can move quickly over the water and keep eight or more victims afloat. These devices can also be carried easily atop lifeguard emergency vehicles.

Statistics

The volume of beach use has expanded tremendously, as has the work of lifesavers. For 1998, the USLA reported the following from major reporting beach lifeguard agencies (Table 1).

Rip Currents

USLA statistics show that over 80% of rescues at surf beaches are due to rip currents. This phenomenon is caused by a variety of factors. First, wave action pushes water up the slope of the beach. Then, gravity pulls it back to sea level. As it seeks to return, the water takes the path of least resistance, which sometimes causes it to be concentrated in currents of water moving away from shore. These currents in the ocean are called rip currents.

Rip currents have three major components. The *feeder* is the main source of supply, composed of water that has been pushed up the beach by wave action. The *neck* is a relatively narrow river of water within the ocean moving back to sea. The *head* is typically a mushroom shaped area of water as the rip current disperses outside the surfline.

Wherever there is regular surf, there will be some form of rip currents. These currents vary in intensity according to wave energy, as well as bottom conditions. For example, a rocky bottom with channels can foment the formation of strong rip currents, as can reefs parallel to shore. They can also form due to channeling of water by undulations in the sand bottom, jetties, groins, and piers.

The USLA has identified several different types of rip currents (Brewster 1995), including:

- *Fixed rip currents*: These are found on sand beaches, and remain in the same place so long as underlying sand conditions remain the same.
- *Permanent rip currents*: These remain in the same area year round and are usually seen on beaches with rocky bottoms, near groins, or piers, where the underlying structure never changes and rip current intensity varies only with swell size and direction.
- *Flash rip currents*: These currents occur suddenly and unexpectedly typically due to sets of waves that are higher in size than other waves and bring unusually high volumes of water ashore quickly. They may form regardless of an obvious differentiation in underlying beach structure.

Lifesaving and Beach Safety, Table 1 Beach Lifeguard Statistics – 1998

Beach attendance	
Total	256,721,418
Rescues	
Total	63,088
<i>Primary cause</i>	
Rip current	27,030
Surf	3,141
Swiftwater	142
SCUBA	112
Cliff rescues	
Total	75
Boat rescues	
Total	2,618
Passengers	3,207
Vessel value	\$56,012,701
Boat assists	
Total	6,487
Passengers	15,865
Vessel value	\$87,693,000
Preventive actions	
Total	2,735,889
Medical aids	
Total	209,317
Major	9,529
Minor	199,788
Drownings	
Total	111
Unguarded area	104
Guarded area	7
Fatalities	
Total	43
Enforcement actions	
Total	618,111
Warnings	594,899
Boat/PWC	16,127
Citations	6,219
Arrests	866
Lost and found persons	
Total	23,958
Public safety lectures	
Total	60,979
Number of students	405,561

Source: United States Lifesaving Association

Note: The addition of the “Primary Causes” of “Rescues” will not add up to the “Total” rescues because some agencies do not specify the cause

- *Traveling rip currents*: These currents usually occur on sandy beaches and move along with the prevailing swell direction as longshore currents move water along the beach.

Worldwide, it would appear that the highest volume of rescues by lifeguards, by a significant margin, exists in the

southern California counties of Los Angeles, Orange, and San Diego. There, regular and strong wave action and resulting rip currents, combined with high, year-round beach attendance, ensure that many persons will need rescue. In 1998, the USLA reports that lifeguards in these three counties effected 43,882 rescues, which represented 66% of all rescues reported to USLA by American beach lifeguard agencies that year. In contrast, Surf Life Saving Australia reports that a total of 12,948 rescues were effected by Australian surf lifesavers in the 1998/99 season.

Drowning Prevention

Another important statistic is “preventive actions.” Preventive actions are typically warnings to swimmers and others to avoid areas of hazard that might result in distress or drowning. Not all agencies report these actions to USLA, but USLA statistics show that for every rescue, there are at least 43 preventive actions by lifeguards. Clearly, this is an essential action, without which the number of rescues and drownings might be much higher. As such, the value of lifeguards extends well beyond the reactive service of rescuing someone in need, to active prevention.

According to the USLA

USLA has calculated the chance that a person will drown while attending a beach protected by USLA affiliated lifeguards at 1 in 18 million (.0000055%). This is based on the last ten years of reports from USLA affiliated lifeguard agencies, comparing estimated beach attendance to the number of drownings in areas under lifeguard protection.

Historically, lifeguards have typically been placed on beaches and the level of lifeguard coverage increased only after drownings have occurred. This may be partially due to a view that swimming and related activities are considered discretionary, recreational activities, and to some, worthy of a lesser level of attention than more common or necessary activities. Nevertheless, the impact of death resulting from drowning, for whatever reason, is the same as that from other causes.

Another problem confronting lifesaving is the lack of standardized systems for rating the ambient hazards at swimming beaches that might dictate specific levels of lifeguard protection. The wide variety of factors that increase the likelihood of distress and drowning are quite complex. Such factors include attendance levels, weather, water temperature, surf, strength of rip currents, swimming skills of users, etc. Recent efforts by Professor Andrew Short on behalf of Surf Life Saving Australia have produced a system to consider these many factors and develop appropriate preventive strategies (Short 1997). It has been effectively applied in several areas. Further work and testing of this system is underway.

Beach Signs

Statistics indicate that the provision of lifeguards results in heightened safety and drowning prevention. In some areas,

however, passive warning systems, such as flags and signs, are the only source of drowning prevention measures. This is particularly true at areas with low attendance or relatively benign ocean conditions. However, as was demonstrated at Ocean Beach in San Francisco, where extremely strongly worded signs were initially employed in place of lifeguards, signs alone are of limited value in drowning prevention.

At present, there is no internationally recognized standard for beach signage. Since many beachgoers are tourists, this lessens the likelihood that preventive signs will be read or understood. Regardless of the reason, signs seem of very limited value. One study of beach signs found that 85–90% of beachgoers did not recall having seen the signs posted at beaches where they were recreating.

Modern Lifesaving

The major national lifesaving organizations of the world have organized themselves into a single worldwide confederation, known as the International Life Saving Federation. Through this organization, lifesavers exchange information, extend lifesaving aid to countries lacking preventive programs, and meet to vie in lifesaving competition.

Many changes have taken place in lifesaving over the years. Effective modern lifeguarding involves a carefully orchestrated system of drowning prevention in which lifeguards in elevated towers oversee designated areas of water and cross-check the edges of areas of responsibility of neighboring lifeguards. Some areas employ a so-called “Tower 0” system, which involves a permanent, elevated structure with an enclosed observation deck overseeing an entire beach area, sometimes more than a mile in length.

The lifeguard in Tower 0 acts as something of an overseeing traffic controller who coordinates activities of lifeguards in beach level towers, sending backup when they make rescues or need assistance for medical aid. This system depends on mobile backup services from emergency vehicles and boats.

The need for emergency vehicles initially arose as lifeguards were expected to patrol larger areas and were summoned away from their regularly assigned stations to emergencies elsewhere. Lifeguards in emergency vehicles can respond over long distances to deliver rescue personnel and equipment to remote areas or simply to better cover longer beaches. They are now an essential element of backup at major beaches that reduce the need for personnel and increase the rescue equipment that can be transported to assist at a rescue.

Vehicles can also provide emergency backup to other lifeguards stationed a significant distance away. The public address systems on lifeguard vehicles are invaluable for delivering preventive warnings to swimmers and communicating to lifeguards in the water.

Rowed dories continue to be used in a few areas of the world for rescue, but since the advent of motorized vessels with compact engines, motorboats have become the rescue vessel of choice in most areas. The 10 m Baywatch boats of Los Angeles County, with their inboard motors, were some of the first boats designed specifically for lifesaving. They allow lifeguards to rescue multiple victims, as well as to respond to offshore boating emergencies. The only limitation of these vessels is that they must remain outside the breaking surf, where victims are sometimes trapped.

When Australian lifesavers first modified commercially available inflatable vessels to operate effectively as rescue boats in the surf environment and dubbed them IRBs, they pioneered one of the most striking modern advancements in lifesaving. Unlike hard hull vessels, these 4 m boats with small outboards are able to navigate the largest and most powerful surf to rescue distressed swimmers and surfers. Even in the unfortunate event of a capsizing, these vessels are easily righted and their outboards rehabilitated for operation. They have proven their worth time and time again in the most inclement conditions.

It was not long ago that personal watercraft (PWCs; also known by the trade name Jet Ski) were introduced to the waterways. To some they are seen as a noisy irritant, or a water toy more suited to an amusement park; however, personal watercraft have transformed boating recreation, becoming one of the most popular types of recreational boat.

Initially, few saw the PWC as a viable rescue tool. Now however, thanks in particular to pioneering efforts by Hawaiian lifeguards and loaner programs provided by some manufacturers, PWCs are employed by many lifeguard agencies as rescue boats. These boats are extremely quick, powerful, usually unaffected by capsizing, and can be operated by a single rescuer; although they work best when two lifeguards work in concert using a towed rescue sled.

As the boats available to lifesavers have changed and adapted, so have the technologies that make boat rescues more effective. Global positioning systems, for example, can now be found aboard many of the larger and more advanced lifeguard rescue boats. And radio direction finders allow boat operators to hone in on an emergency radio signal from a boat in distress.

The most sophisticated rescue implement used by lifeguards today is aircraft, particularly helicopters. In some areas, lifeguards have made arrangements with a local police or rescue helicopter service to assist them in times of need; but in Rio de Janeiro, Brazil, Durban, South Africa, New Zealand, and Australia, helicopters have long been a basic tool of the lifeguard agencies themselves. They allow quick access to offshore or remote emergencies, unimpeded by traffic or surf conditions. They also allow for rapid evacuation of the injured to hospitals. When not used for emergencies, helicopters are excellent platforms for observation, patrols

of remote areas, searches, and dissemination of public information.

Communication has been a perennial challenge for lifesaving. Particularly at beaches, where lifeguards may cover broad expanses of shifting sand, remote areas, and the open sea, there is a need to quickly summon backup or advise others of an emergency in progress. Whistles, megaphones, and flags were some of the first tools used for this purpose, but their value declines rapidly as distance increases. For this reason, two-way radios have become the communication tool of choice.

First aid is another area of tremendous change. Bottled oxygen, suction devices, and one-way masks are all highly recommended pieces of first aid equipment, which many lifesavers have available. So are spinal and cervical immobilization devices. A recent innovation in this area is the floating backboard. Implements to protect against communicable disease have also become a must.

All of these improvements in equipment, technology, and rescue techniques, along with the exploding attendance levels at aquatic areas, now demand training levels among lifesavers that go well beyond the early days. Effective lifesaving has always called for superior resuscitation skills, but with advancements in techniques this has come to require tens of hours of training in both resuscitation and professional level cardio-pulmonary resuscitation. In some areas, lifeguards are trained to the level of paramedics, and many full-time lifeguards are emergency medical technicians.

General rescue skills have been advanced also. In America, lifeguards assigned to coastal beaches typically receive 80 to well over 100 h of basic training before they are given basic lifesaving duties. Full-time lifeguards, who work year-round in California, Florida, and Hawaii receive many additional hours of training. In the European Union, the minimum standard to become a lifeguard is now some 300 h. Other countries exceed even these levels.

In addition to the responsibility for water rescues, some lifeguards have been called upon to perform specialized emergency services in and around the aquatic environment. This has required special training and equipment appropriate to the task.

Expanding Lifeguard Services

From the rocky Irish coastline, to the soft sandstone of the California coast, people headed for the beach are sometimes stranded on the cliffs above. Some lifeguards are trained and equipped in high-angle rescue to pluck them from their plight. In San Diego, where over 25 of these rescues are performed each year, a special rescue vehicle with a crane aboard is used to lower rescuers and raise victims to safety.

All lifeguards have some degree of responsibility for controlling the swimmers they protect and regulating activities, but in some places that responsibility has advanced to regular law enforcement power. Lifeguards employed by the State of California and Volusia County, Florida are empowered as police officers, carrying firearms, and enforce even the most serious crimes that occur on their beaches.

Some of the other specialized services provided by lifeguards include marine firefighting and flood rescue. At least two lifeguard agencies, San Diego and Los Angeles County, have expanded upon the emergency callback system used by other lifeguard agencies to staff lifeguards 24 h a day to respond to the many calls that come in to their dispatch centers during nighttime hours.

Lifeguards are also participating in mutual aid networks within their communities. As many lifesaving groups align themselves with police and fire agencies, they are increasingly called upon to act in concert with other public safety providers when natural disasters strike.

Lifeguards around the world are taking their preventive responsibilities one step further through the development of youth programs. Called nipper or junior lifeguard programs, they train youngsters in ways to safely use the waters and encourage their later participation as lifesavers. Many thousands participate in these programs each year.

Lifesaving competition too, has become tremendously popular, involving some 600,000 people annually. Many of the early international exchanges of lifesaving information came through the solidarity brought about through competition and this tradition continues with events sponsored by the International Life Saving Federation and its member federations, as well as other groups. These competitions encourage not only information exchange and international goodwill, but also inspire lifesavers to maintain the high levels of fitness needed to effectively save lives in the water.

With the continually increasing responsibilities, technologies, and training, the role of the lifesaver has been transformed over the years. The lifesaver has acquired an internationally recognized and tremendously positive image of a well prepared, physically fit, and versatile person ready for any emergency that might develop in or near the water. This has benefited water safety, but it has also burnished the image of lifeguards, now more likely than ever to be seen as providing an integral layer of essential public safety protection, allowing safe use of a sometimes hazardous coastline.

Cross-References

- [Beach Use and Behaviors](#)
- [Coastal Currents](#)
- [Environmental Quality](#)
- [Rating Beaches](#)

- [Rip Currents](#)
- [Sandy Coasts](#)
- [Surf Zone Processes](#)
- [Surfing](#)
- [Water Quality](#)

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Littoral

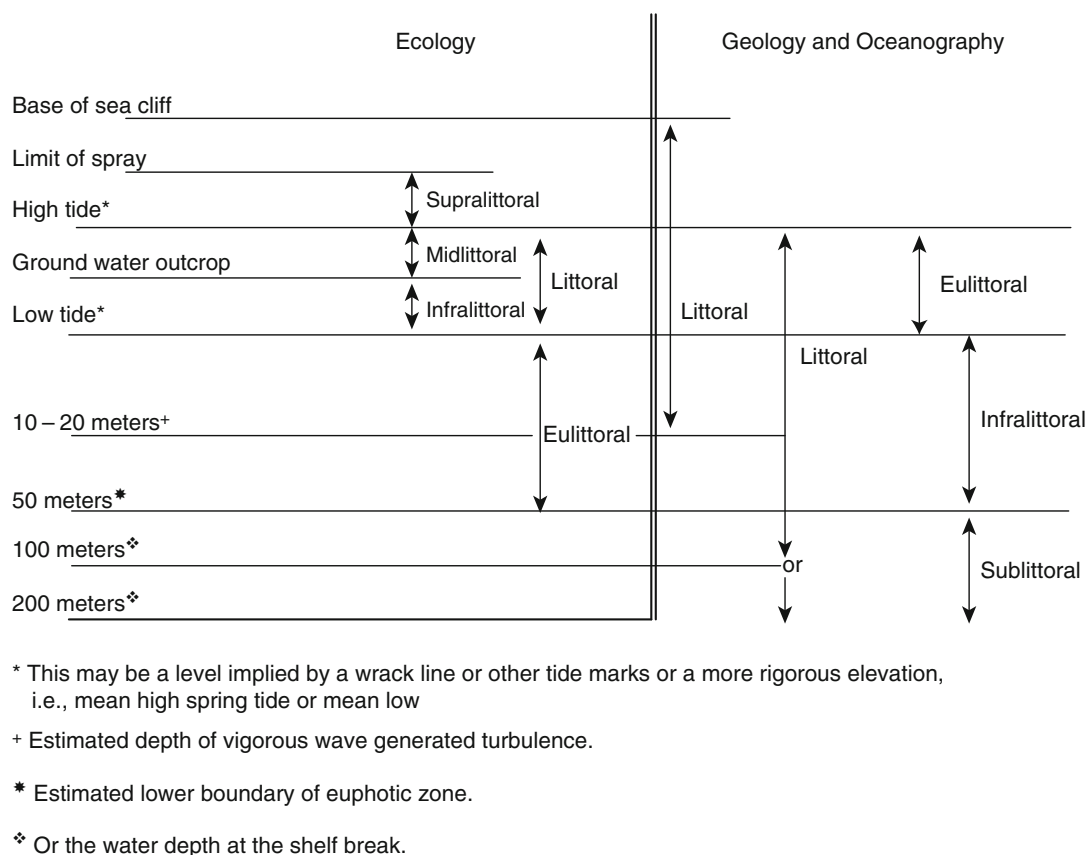
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Definition

In the vernacular, “littoral” refers to a shore or coastal region from the Latin *litus*, shore. In the technical usage, “littoral” and its associated derivative nomenclature are variously defined primarily depending on the disciplinary context. Even within a discipline, the terms tend to be used with some elasticity and in a semi-quantitative sense due to the quantitative imprecision of boundaries defined by “high tide” or “low tide,” “ordinary surf,” etc. or by primary and secondary biotic transitions (Fig. 1).

The first technical use of “littoral” in English was in a biological context (Forbes and Hanley 1853, as cited in Hedgpeth 1957) to designate the space between tide marks or the intertidal. “Supralittoral” and “sublittoral” were first employed in a similar context (Lorenz 1863, as cited in Hedgpeth 1957) with supralittoral being near the shore but above the level of high tide and sublittoral defined by water depth between two fathoms (3.65 m) and fourteen fathoms (25.60 m).



Littoral, Fig. 1 Vertical boundaries for various usage of "littoral" and its derivatives

Ecological Context

Marine ecologists have used the term to not only include the region between the tides but also the substrate wetted by ordinary surf above the level of high tides or "spray zone." In modern usage, the "supralittoral" or supratidal generally has referred to the spray zone immediately above either the highest high tide (Nybakken 1993) or the mean high spring tide (Eisma 1997) where the substrate is commonly, more-or-less, moistened by extreme tides, waves, and spray. This has also been called "extratidal." Ecologists divide the zone between the tide levels, into a "mid-littoral" from highest high tide to the level at which the water table on the beach outcrops at the beach surface; and an infratidal from the level of the water table outcrop to the elevation of the lowest low tide (Nybakken 1993). "Eulittoral" refers to that part of the littoral zone less than 50 m in depth (American Geological Institute 1960), 50 m, or between 40 and 60 m, being the lower limit at which "more abundant attached plants can grow" (Sverdrup et al. 1942 citing Ekman 1935). In practice, these terms are not so much defined by the physical

processes controlling the landward distribution of sea salt but by the flora and fauna characteristic of such an influence.

In limnology, "littoral" refers to the zone below the highest reach of typical waves in a calm season (i.e., where the ground is dampened by waves) and above the maximum depth of substrate supporting rooted plants with floating or emergent leaves or flowers (e.g., Pillsbury 1970; Jackson 1997). Strictly speaking, it is possible for the littoral zone of a lake to cover the entire submerged area.

Geological and Oceanographic Context

When applied to the marine environment by marine geologists, "littoral" generally refers to the zone between high tide and low tide. Classification of these zones was set forth by Johnson (1919) with reference to earlier investigations and to "legal authorities"; "shore" and "beach" both were synonymous for the intertidal and also referred to as "littoral." More strictly speaking, the "littoral" is that

region between mean high water spring tide and mean low water spring tide (Eisma 1997). Such a definition recognizes the complexity of the tidal signal and vicissitudes of a particular high tide level around some average position. “High tide” (or “low tide”) is an imprecise term because these levels vary from one tidal cycle to the next. Harmonic analysis provides dozens of tidal constituents, acting at different periods, which conspire to provide any instantaneous high tide (or low tide). In other geological context, the “littoral” zone has a lower boundary at the depth of vigorous wave-generated turbulence, generally 10–20 m or, in one case, one-third the wave length of storm-generated surface waves, and an upper limit at the seaward foot of sand dunes on the shore of the base of beach ridges or sea cliffs (Fairbridge 1968). In a common usage, where the lower boundary of the littoral zone is at the level of low tide, “sublittoral” is the region below the low tide level to a depth of 100 m (American Geological Institute 1960) or 200 m (Nybakken 1993). This has also been referred to as the “shelf zone.”

Oceanographers have tended to use “littoral” to designate the broad region from the high tide elevation down to the compensation point, generally between 200 and 400 m in the shelf. Alternatively, it is used as a synonym for “neritic” referring to the region below the low tide level to a depth of 200 m or to the shelf break (Kuenen 1950, p. 313; Jackson 1997). In this usage “eulittoral” can be synonymous with “intertidal.” “Infralittoral” is the shallow (or inner) sublittoral generally the euphotic zone or between low tide and the depth compatible with the occurrence of phanerograms or photophilous algae (Visser 1980, term 2839). Alternatively, the lower limit has been cited as about 50 m (Hedgpeth 1957). “Sublittoral” is the region between water depths of 50 and 200 m.

The terms, without the “intertidal,” “supratidal,” “subtidal” synonyms of course, may also be applied to the shores of tideless seas where regular water level variations due to, perhaps, storm surges, seiches, etc. are observed. In light of these various usages, these terms are best specifically defined in articles, and some care may need to be exercised when reading the literature.

The advent of laser altimetry and bathymetry (Light Detection and Ranging, LIDAR) add a new dimension littoral classification based on coastal geometry. This classification based on elevation provides high-resolution and can provide a set of morphological metrics, (e.g., distribution of elevation, shore width, slope, and roughness) at high resolution (Nijland et al. 2017; Sallenger et al. 2003). Yet another dimension to littoral zones is provided by legal, jurisdiction boundaries (e.g., Clark 1992). In the United States, these include lands in the public trust between the mean high water line and the mean sea level, State tidelands

from mean sea level to mean low water, the State submerged land to a distance of 3n miles from the mean high water shorelines and, beyond mean lower low water the 12-nautical mile Federal territorial sea, the 24-nautical mile Federal contiguous zone and the County’s Exclusive Economic Zone.

Cross-References

- [Beach Features](#)
- [Coastal Boundaries](#)
- [Hydrology of the Coastal Zone](#)
- [Tidal Environments](#)
- [Tides](#)

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Littoral Cells

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A littoral cell is a coastal compartment that contains a complete cycle of sedimentation including sources, transport paths, and sinks. The cell boundaries delineate the geographical area within which the budget of sediment is balanced, providing the framework for the quantitative analysis of coastal erosion and accretion. The sediment sources are commonly streams, sea cliff erosion, onshore migration of sand banks, and material of biological origin such as shells, coral fragments, and skeletons of small marine organisms. The usual transport path is along the coast by waves and currents (longshore transport, longshore drift, or littoral drift). Cross-shore (on/offshore) paths may include windblown sand, overwash, and ice-push. The sediment sinks are usually offshore losses at submarine canyons and shoals or onshore dune migration, rollover, and deposition in bays and estuaries (Fig. 1).

The boundary between cells is delineated by a distinct change in the longshore transport rate of sediment. For example, along mountainous coasts with submarine canyons, cell boundaries usually occur at rocky headlands that intercept transport paths. For these coasts, streams and cliff erosion are the sediment sources, the transport path is along the coast and driven by waves and currents, and the sediment sink is generally a submarine canyon adjacent to the rocky headland. In places, waves and currents change locally in response to complex shelf and nearshore bathymetry, giving rise to sub-cells within littoral cells (e.g., Figs. 2, 3, and 4).

The longshore dimension of a littoral cell may range from one to hundreds of kilometers, whereas the cross-shore dimensions are determined by the landward and seaward extent of the sediment sources and sinks. Littoral cells take a variety of forms depending on the type of coast. Cell forms are distinctive of the following coastal types: collision (mountainous, leading edge), trailing-edge, marginal sea, arctic, and coral reef. The first three types are determined by their position on the world's moving plates while the latter two are latitude dependent.

Background

The concept of a littoral cell followed from the observation that the southern California coast was naturally divided into

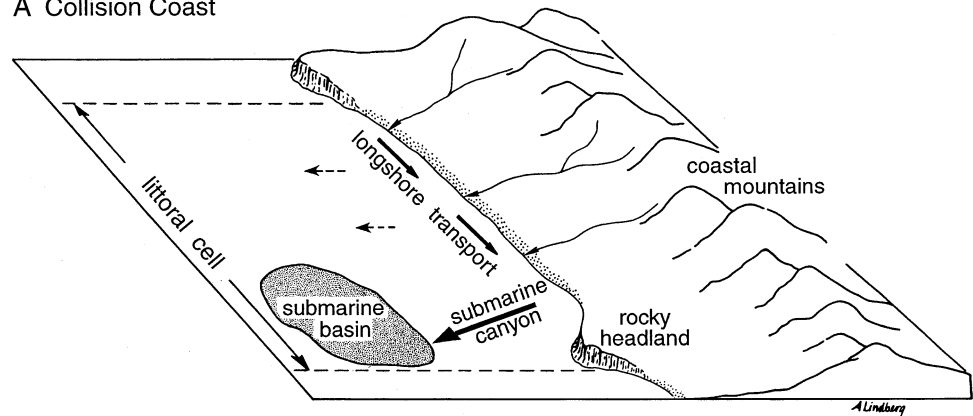
discrete sedimentation cells by the configuration of the coastal drainage basins, headlands, and shelf bathymetry. The principal sources of sediment were the rivers, that periodically supplied large quantities of sand to the coast. The sand is transported along the coast by wave action until the longshore drift of sand is intercepted by a submarine canyon that diverts and channels the flow of sand into offshore basins (Fig. 1a). It was found that littoral cells, because they contain a complete cycle of sedimentation, provided the necessary framework for balancing the budget of sediment. These concepts were first presented at the International Geological Congress, Copenhagen (Inman and Chamberlain 1960). The littoral cell now plays an important role in the US National Environmental Protection Act (1974) and the California Environmental Quality Act (1974), and it has become a necessary component of environmental impact studies. In the realm of public policy and jurisdictions, the littoral cell concept has led to joint-power legislation that enables municipalities within a littoral cell to act as a unit (Inman and Masters 1994).

The configuration of littoral cells depends on the magnitude and spatial relations among the sediment sources, transport paths, and sinks. These in turn have been shown to vary systematically with coastal type. Because the large-scale features of a coast are associated with its position relative to the margins of the earth's moving plates, plate tectonics provides a convenient basis for the first-order classification of coasts (Inman and Nordstrom 1971; Davis 1996). This classification leads to the definition of three tectonic types of coast: (1) collision coasts that occur on the leading edge of active plate margins where two plates are in collision or impinging on each other, for example, the west coasts of the Americas; (2) trailing-edge coasts that occur on the passive margin of continents and move with the plate, for example, the east coasts of the Americas; and (3) marginal sea coasts that develop along the shores of seas enclosed by continents and island arcs, for example, coasts bordering the Mediterranean Sea and the South and East China Seas.

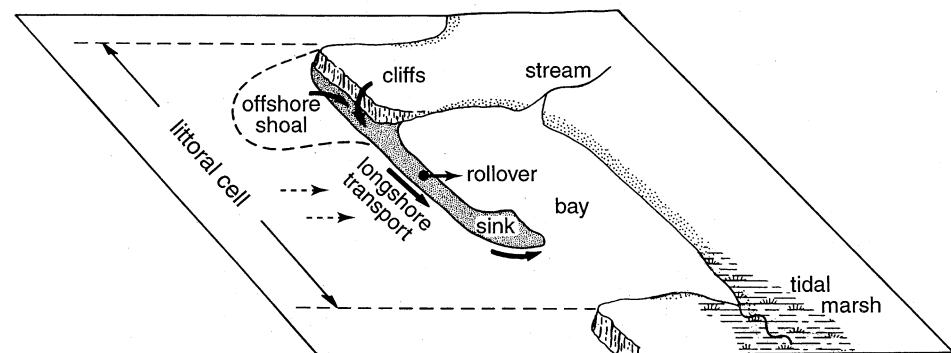
It is apparent that the morphologic counterparts of collision, trailing-edge, and marginal sea coasts become, respectively, narrow-shelf mountainous coasts, wide-shelf plains coasts, and wide-shelf hilly coasts. However, some marginal sea coasts such as those bordering the Red Sea, Gulf of California, Sea of Japan, and the Sea of Okhotsk are narrow-shelf hilly to mountainous coasts. A more complete coastal classification includes the latitudinal effects of climate and other coastal forming processes such as ice-push and scour and reef-building organisms. The examples of the latter two coastal types described here are (4) arctic form of cryogenic coasts; and (5) coral reef form of biogenic coasts. The kinds of source, transport path, and sink commonly associated with littoral cells along various types of coast are summarized in Table 1.

Littoral Cells, Fig. 1 Typical (a) collision and (b) trailing edge coasts and their littoral cells. Solid arrows show sediment transport paths; broken arrows indicate occasional onshore and offshore transport modes. (After Inman 1994)

A Collision Coast



B Trailing-Edge Coast



Collision Coasts

Collision coasts form at the active margins of the earth's moving plates and are best represented by the mountainous west coasts of the Americas. These coasts are erosional and characterized by narrow shelves and beaches backed by wave-cut sea cliffs. Along these coasts with their precipitous shelves and submarine canyons, as in California, the principal sources of sediment for each littoral cell are the rivers that periodically supplied large quantities of sandy material to the coast. The sand is transported along the coast by waves and currents primarily within the surf zone like a *river of sand*, until intercepted by a submarine canyon. The canyon diverts and channels the flow of sand into the adjacent submarine basins and depressions (Fig. 1a).

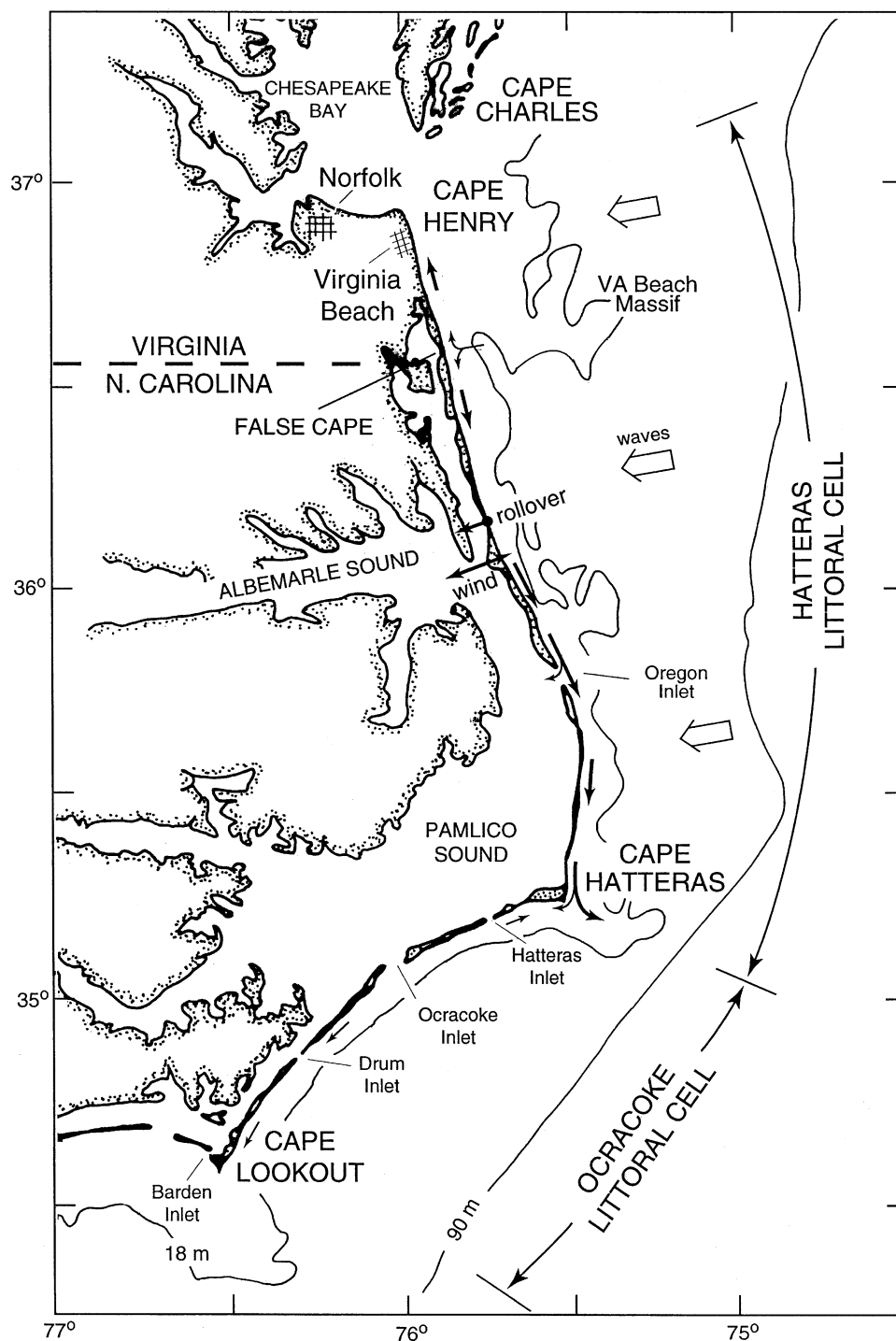
However, in southern California most coastal rivers have dams that trap and retain their sand supply. Studies show that in this area the yield of sediment from small streams and coastal bluffs has become a significant replacement for river sediment. Normal wave action contains sand against the coast and, when sediment sources are available, results in accretion of the shorezone. However, cluster storms associated with El Niño–Southern Oscillation events as occurred in 1982/83 produced beach disequilibrium by downwelling

currents that carried sand onto the shelf (Inman and Masters 1991). The downwelled sediment is lost to the shorezone when deposited on a steep shelf such as that off Oceanside, California, or it may be returned gradually from a more gently sloping shelf to the shorezone by wave action. The critical value of slope for onshore transport of sand by wave action varies with sand size, depth, and wave climate, but for depths of about 15–20 m it is approximately 1.5% (1.0 degree).

Trailing-Edge Coasts

Trailing-edge coasts occur along the passive plate margins of continents and include the coasts of India and the east coasts of the Americas. The mid-Atlantic coast of the United States, with its wide shelf bordered by coastal plains, is a typical trailing-edge coast where the littoral cells begin at headlands or inlets and terminate at embayments and capes (Figs. 1b and 2). This low-lying barrier island coast has large estuaries occupying drowned river valleys. River sand is trapped in the estuaries and does not usually reach the open coast. For these coasts, the sediment source is from beach erosion and shelf sediments deposited at a lower stand of the sea, whereas the sinks are sand deposits that tend to close and fill estuaries and

Littoral Cells, Fig. 2 Hatteras and Ocracoke Littoral Cells along the Outer Banks of North Carolina. (After Inman and Dolan 1989)



form shoals off headlands. Under the influence of a rise in relative sea level, the barriers are actively migrating landward by a rollover process in which the volume of beach face erosion is balanced by rates of overwash and fill from migrating inlets (e.g., Inman and Dolan 1989). For these coasts, the combination of longshore transport and rollover processes leads to a distinctively “braided” form for the *river of sand* that moves along the coast.

The Outer Banks of North Carolina, made up of the Hatteras and Ocracoke Littoral Cells, extend for 320 km and are the largest barrier island chain in the world (Fig. 2). The Outer Banks are barrier islands separating Pamlico, Albemarle, and Currituck Sounds from the Atlantic Ocean. These barriers are transgressing landward, with average rates of shoreline recession of 1.4 m/year between False Cape and Cape Hatteras. Oregon Inlet, the only opening in the nearly

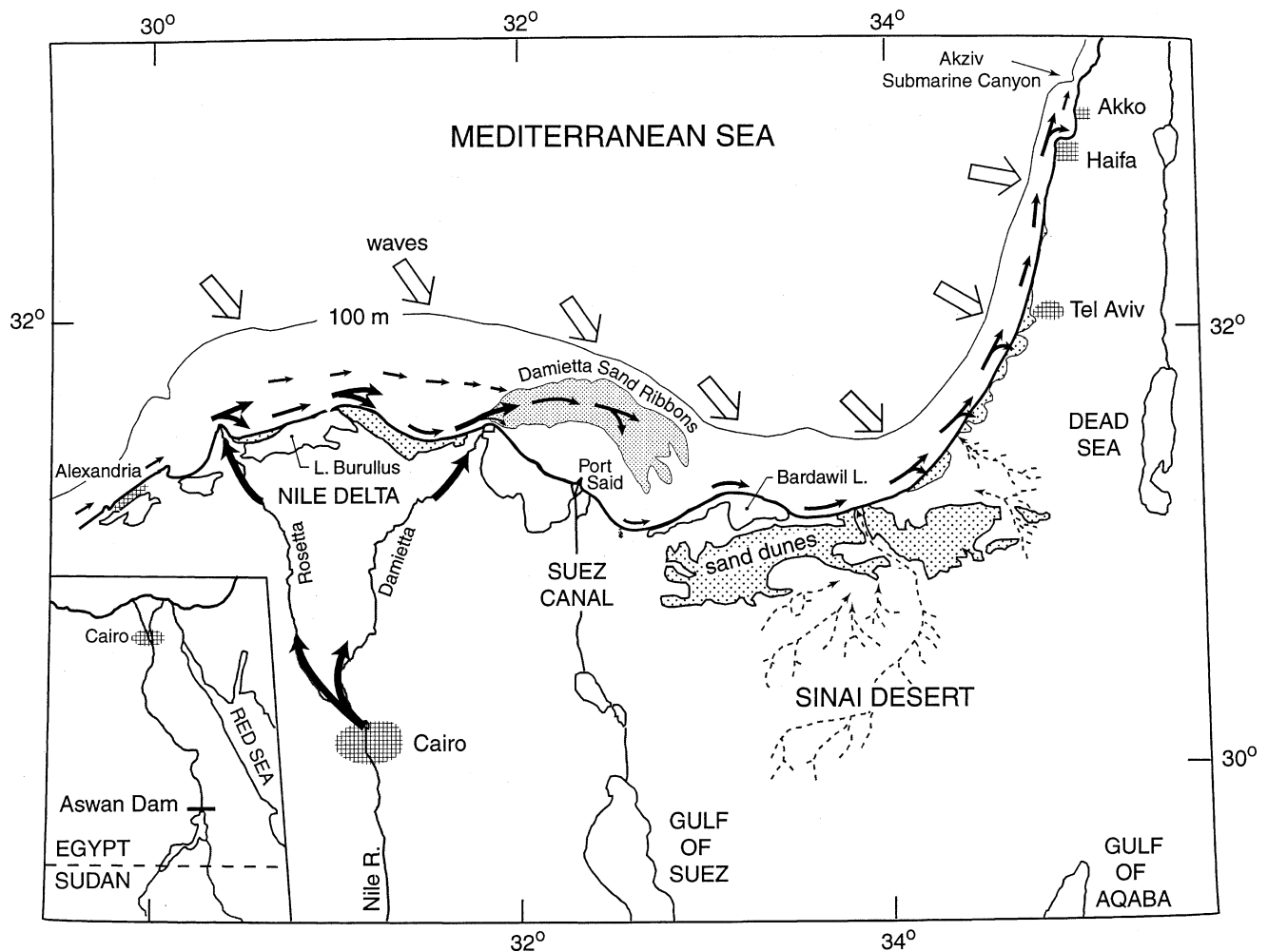
200 km between Cape Henry and Cape Hatteras, is migrating south at an average rate of 23 m/year and landward at a rate of 5 m/year. The net southerly longshore transport of sand in the vicinity of Oregon Inlet is between one-half million and one million cubic meters per year.

Averaged over the 160 km from False Cape to Cape Hatteras, sea-level rise accounts for 21% of the measured shoreline recession of 1.4 m/year. Analysis of the budget of sediment indicates that the remaining erosion of 1.1 m/year is apportioned among overwash processes (31%), longshore transport out of the cell (17%), windblown sand transport (14%), inlet deposits (8%), and removal by dredging at Oregon Inlet (9%). This analysis indicates that the barrier system moves as a whole so that the sediment balance is relative to the moving shoreline. Application of a continuity model to the budget suggests that, in places such as the linear shoals off False Cape, the barrier system is supplied with sand from the shelf (Inman and Dolan 1989).

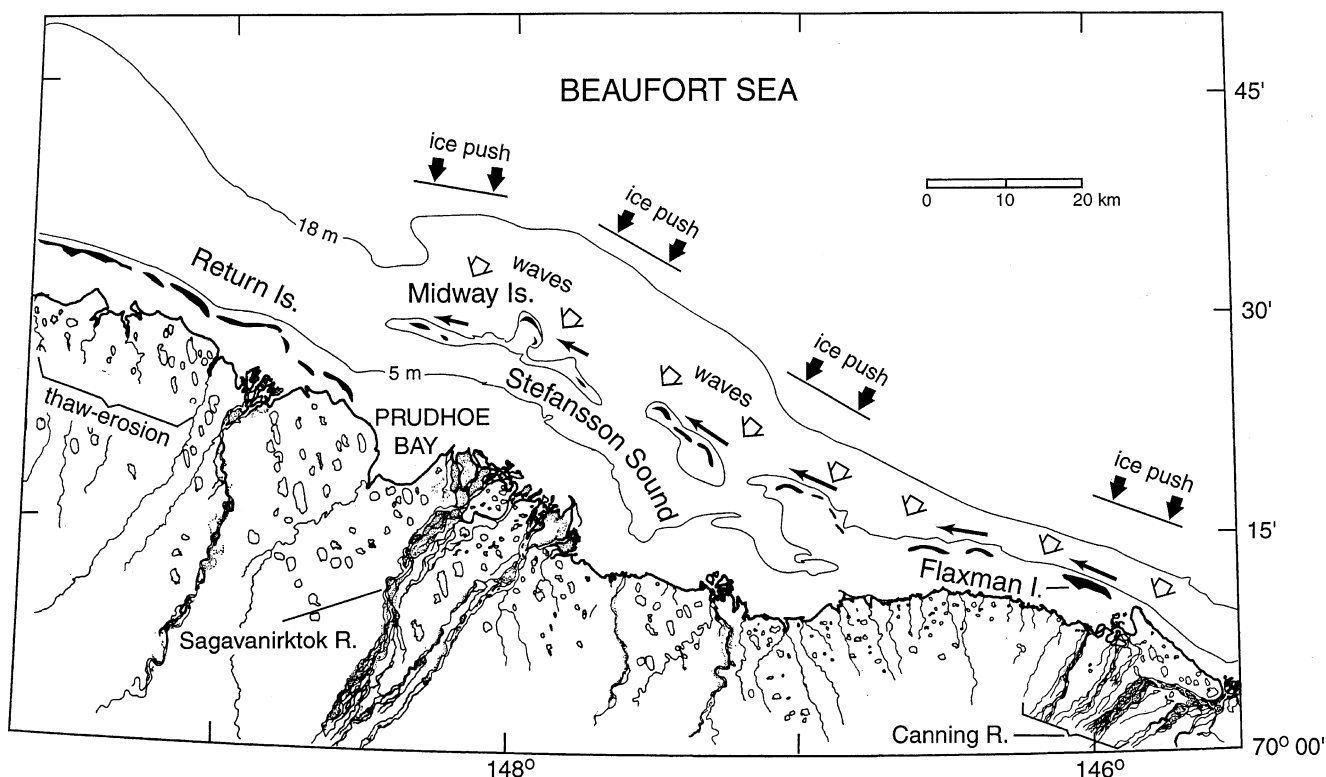
Marginal Sea Coasts

Marginal sea coasts front on smaller water bodies and are characterized by more limited fetch and reduced wave energy. Accordingly, river deltas are more prominent and are often important sources of sediment within the littoral cell. Elsewhere, barrier island rollover processes are similar to those for trailing-edge coasts. Examples of marginal sea coasts include the shores of the Gulf of Mexico with the prominent Mississippi River delta, the seas bordering southeast Asia and China with the Mekong, Huang (Yellow), and Luan river deltas, and the Mediterranean Sea coasts with the Ebro, Po, and Nile river deltas.

Although the Mediterranean area is associated with plate collision, the sea is marginal with restricted wave fetch and prominent river deltas. The Nile Littoral Cell extends 700 km from Alexandria on the Nile Delta to Akziv Submarine Canyon near Akko, Israel, one of the world's longest littoral cells (Fig. 3). Before construction of the High Aswan Dam,



Littoral Cells, Fig. 3 The Nile Littoral Cell extends along the southeastern Mediterranean coast from Alexandria, Egypt to Akziv Submarine Canyon off Akko, Israel. Sediment transport paths shown by solid arrows. (After Inman and Jenkins 1984)



Littoral Cells, Fig. 4 Flaxman Littoral Cell extending 100 km from the mouth of Canning River to the Midway Islands. The barrier chain of islands is enclosed by the 5 m depth counter. Major axis of oriented thaw lakes are normal to the direction of summer winds. (After Inman 1994)

Littoral Cells, Table 1 Typical source, transport path, and sink for littoral cells of various coastal types

Coastal features	Collision	Trailing-edge	Marginal sea		Arctic form of cryogenic	Coral reef form of biogenic
Morphology	Narrow-shelf mountainous	Wide-shelf plains	Narrow-shelf mountainous	Wide-shelf hilly	Wide-shelf ^a plains	Coral reef
Latitude/climate	Temperate and subtropical	Temperate and subtropical	Temperate and subtropical	Temperate and subtropical	Arctic	Tropical
Forcing ^b	Waves (1–10 kw/m)	Waves (1–5 kw/m)	Fetch-limited waves (1–2 kw/m) Tides ^c	Fetch-limited waves (1–2 kw/m) Tides ^c	Winter ice-push Summer waves	Waves (1–10 kw/m)
Littoral cell						
Sediment source	Rivers Cliffs Blufflands	Headlands Cliffs Shelves	Rivers Deltas	Rivers Deltas	Shelf Rivers Thaw-erosion	Reef material
Transport path	Longshore (river of sand)	Longshore and, rollover ^d (braided river of sand)	Longshore	Longshore and rollover ^d	Ice-push Rafting Longshore	Reef surge channels to beach, longshore to awa
Sink	Submarine canyons Embayments Dune migration	Estuaries Shoals Rollover Dune migration	Various including submarine canyons	Embayments Shoals Rollover Dune migration	Shoals Spit-extension	Awa channels to shelf

^aAll high latitude coasts appear to be trailing-edge coasts

^bAverage incident wave energy-flux per meter of coastline (Inman and Brush 1973)

^cTides may be important along any ocean coasts, but are sometimes amplified in marginal seas

^dRollover processes include overwash and dune migration

the Nile Delta shore was in a fluctuating equilibrium between sediment supplied by the river and the transport along the coast. Now the sediment source is erosion from the delta, particularly the Rosetta promontory, in excess of 10 million m^3/year . The material is carried eastward in part by wave action, but predominantly by currents of the east Mediterranean gyre that sweep across the shallow delta shelf with speeds up to 1 m/s. Divergence of the current downcoast from Rosetta and Burullus promontories forms accretionary blankets of sand that episodically impinge on the shoreline. The sand blankets move progressively downcoast at rates of 0.5–1 km/year in the form of accretion/erosion waves. Along the delta front, coastal currents augmented by waves transport over 10 million m^3/year , and the longshore sand transport by waves near the shore is about 1 million m^3/year (Inman and Jenkins 1984; Inman et al. 1992).

The Damietta promontory causes the coastal current from the east Mediterranean gyre to separate from the coast and form a large stationary eddy that extends offshore of the promontory, locally interrupting the sediment transport path. The jet of separated flow drives a migrating field of sand ribbons northeasterly across the delta (Fig. 3). The ribbons arc easterly than southeasterly towards the coast between Port Said and Bardawil Lagoon (Murray et al. 1981). The Damietta sand ribbons form the eastern edge of a subcell within the Nile Littoral Cell.

Off Bardawil Lagoon, the longshore sand transport is about 500,000 m^3/year and gradually decreases to the north with the northerly bend in coastline. This divergence in the littoral drift of sand results in the build up of extensive dune fields along the coasts of the delta, Sinai, and Israel. This sediment loss by wind blown sand constitutes a major “dry” sink for sand in the Nile Littoral Cell.

Arctic Coasts

Arctic coasts are those near and above the Arctic Circle ($66^{\circ}34'\text{N}$ Latitude) that border the Arctic Ocean and whose littoral cells have drainage basins in North America, Europe, and Asia. Tectonically, Arctic coasts are of the stable, trailing-edge type, with wide shelves backed by broad coastal plains built from fluvial and cryogenic processes. The coastal plains are permafrost with tundra and thaw lakes. A series of barrier island chains extends along the Beaufort Sea coast of Alaska (Fig. 4). For these coasts, cryogenic processes such as ice-push and permafrost thaw compete with river runoff, waves, and currents as important sources, transport paths, and sinks for sediment. Ice-push is a general term for the movement of sediment by the thrust of ice against it. Some common features include ice-push ridges and mounds, ice-gouge, ice pile-up, ride-up rubbing, and bulldozing.

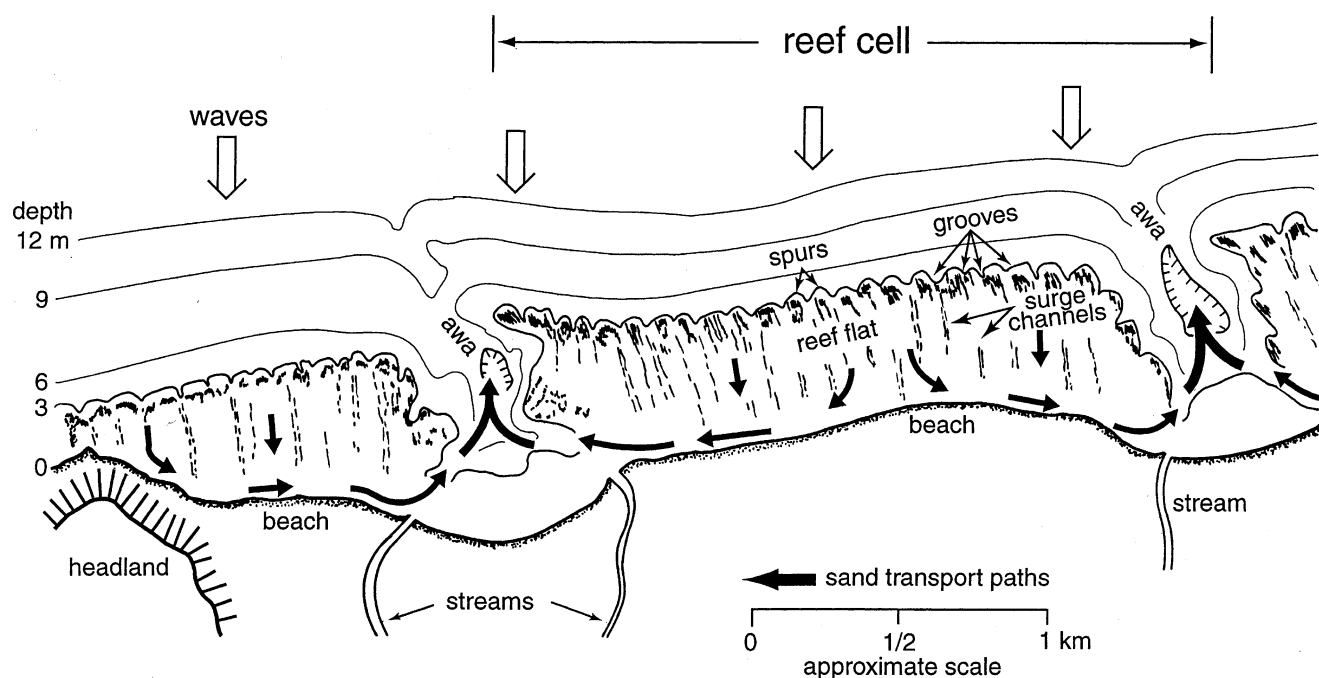
During the 9 months of winter, Arctic coasts are frozen solid and coastal processes are entirely cryogenic. Wind stress and ocean currents buckle and fracture the frozen pack ice into extensive, grounded, nearshore, pressure-ridge systems known as stamukhi zones. The stamukhi zone is a shear zone of ice grounded in 10–25 m depth that molds and moves shelf and barrier island sediment. The keels from the individual pressure ridges groove and rake the bottom, plowing sediment toward the outer barrier islands. Ice-gouge relief up to 2 m occurs across the shelf to depths of about 60 m (Barnes et al. 1984).

Winter is terminated by a very active transitional period of a few days to a few weeks during spring breakup when a combination of factors associated with ice movement, waves, and currents, and extensive fluvial runoff all work in concert along the coast. The grounded ridges in the stamukhi zone break up and move, producing ice-push features and vortex scour by currents flowing around the grounded ice, creating an irregular bottom known as ice-wallow topography. Closer to shore, vertical drainage of river floodwater and sediment through cracks in the shorefast ice form large strudel-scour craters in the bottom (Reimnitz and Kempema 1983).

Finally, a short summer period occurs in which the ice pack withdraws from the Beaufort Sea coast forming a 25–50 km wide coastal waterway. Although the summer season is short, storm waves generated in the band of ice-free water transport relatively large volumes of sand, extending barrier islands and eroding deltas and headlands. The summer processes are classical nearshore phenomena driven by waves and currents as shown by the beaches and barrier island chain beginning with Flaxman Island in the vicinity of Prudhoe Bay (Fig. 4). The sediment sources include river deltas, onshore ice-push of sediment, and thaw-erosion of the low-lying permafrost sea cliffs. Thaw-erosion rates of the shoreline are typically 5–10 m/year in arctic Russia and, over a 30-year period, averaged 7.5 m/year for a 23-km coastal segment of Alaska's Beaufort Sea coast midway between Point Barrow and Flaxman Barrier Islands (Reimnitz and Kempema 1987).

The Flaxman Barrier Island chain extends westward from the delta of the Canning River. It appears to be composed of sand and gravel from the river, supplemented by ice-push sediments from the shelf (Fig. 4). The prevailing easterly waves move sediment westward from one barrier island to the next. The channels between islands are maintained by setdown and setup currents associated with the Coriolis effect on the wind-driven coastal currents. The lagoons behind the barrier islands appear to have evolved in part from collapse and thaw-erosion of tundra lakes (Wiseman et al. 1973; Naidu et al. 1984).

However, even the summer period is punctuated by occasional “Arctic events,” including ice-push phenomena and unusually high and low water levels associated with storm surges and with Coriolis setup and setdown, a phenomenon



Littoral Cells, Fig. 5 Schematic diagram of littoral cells along a fringing reef coast. (After Inman 1994)

whose intensity increases with latitude. The active summer season ends with the beginning of fall freeze-up.

Coral Reef Coasts

Coral reef coasts are a subset of the broader category of biogenous coasts where the source of sediment and/or the sediment retaining mechanism is of biogenous origin as in coral reef, algal reef, oyster reef, and mangrove coasts. Coral reefs occur as fringing reef, barrier reef, and atolls, and they are common features in tropical waters of all oceans at latitudes within the 20 °C isotherm.

Although the concept of the littoral cell applies to all types of coral reef coast, the most characteristic are littoral cells along fringing reef coasts bordering high islands, where both terrigenous and biogenous processes become important. Reefs may be continuous along the coast or occur within embayments. In either case, the configuration of the fringing reef platforms themselves incorporates the nearshore circulation cell into a unique littoral cell (Fig. 5). The circulation of water and sediment is onshore over the reef and through the surge channels, along the beach toward the awas (return channels), and offshore out the awas. An awa is equivalent to a rip channel on the sandy beaches of other coasts (Inman et al. 1963).

Along coral reef coasts, the corals, foraminifera, and calcareous algae are the sources of sediment. The overall health of the reef community determines the supply of beach

material. Critical growth factors are light, ambient temperature, salinity, and nutrients. Turbidity and excessive nutrients are deleterious to the primary producers of carbonate sediments. On a healthy reef, grazing reef fishes bioerode the coral and calcareous algae and contribute sand to the transport pathway onto the beach.

The beach behind the fringing reef acts as a capacitor, storing sediment transported onshore by the reef-moderated wave climate. It buffers the shoreline from storm waves, and releases sediment to the awas. In turn, the awas direct runoff and turbidity away from the reef flats and out into deepwater. Where the reef is damaged by excessive terrigenous runoff, waste disposal, or overfishing, the beaches are imperiled.

Cross-References

- [Arctic, Coastal Geomorphology](#)
- [Barrier Island Landforms](#)
- [Climate Patterns in the Coastal Zone](#)
- [Coasts, Coastlines, Shores, and Shorelines](#)
- [Coral Reefs](#)
- [Deltas](#)
- [El Niño–Southern Oscillation \(ENSO\)](#)
- [Energy and Sediment Budgets of the Global Coastal Zone](#)
- [Holocene Coastal Geomorphology](#)
- [Littoral Drift Gradient](#)
- [Sediment Budget](#)
- [Tectonics and Neotectonics](#)

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Littoral Drift Gradient

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It is well known both empirically, and theoretically from sediment transport modeling that rates of net littoral drift or longshore sediment transport rates (m^3/year), vary progressively along a littoral drift coastline – see, for example, Fig. 9.9 in Komar (1998, p. 386; also pp. 434–435). Disparity in net littoral drift rates along the coastline may arise from several causes, but fundamentally important is variation of the angle of the breaking waves, which drive the alongshore sediment transport. This typically results from change in direction of the wave power resultant relative to the coastline alignment, or from irregularities in nearshore bathymetry influencing the wave shoaling and refraction processes, for example, wave focusing. Other causes include the loss of the littoral drift to a sedimentation “sink” such as a dredged tidal inlet, an offshore submarine canyon, or the action of headlands and cusped forelands partially blocking and accumulating the littoral drift. The general effect is that progressively alongshore in the net drift direction, some segments of coastline have increasing rates of net littoral drift, and others decreasing rates. This alongshore variation in net littoral drift rates constitutes a *littoral drift gradient*. Van de Graaff and Bjiker (1988) state: “In many cases a gradient in the longshore sediment transport is the main reason of the erosion problems of sandy coasts.” An alongshore increase in sediment transport rates in the net drift direction creates a *positive* littoral drift gradient; conversely a decrease in transport rates in the net drift direction results in a *negative* littoral drift gradient.

Littoral drift gradients have significant implications for the beach geomorphology. Consider a given point on a coastline with *positive littoral drift gradient* (increasing rates of net drift alongshore). Because more sediment is being transported away from the sector of beach at that given point than is being replenished from updrift, a net beach sediment budget deficit would be expected. Accordingly, the beach will likely exhibit long-term geomorphic manifestations of an erosive beach, such as a tendency for a dissipative beach state, faceted dune faces, frequent occurrences of lag deposits of surficial laminae of heavy mineral concentration, or cobble-pebble concentrations on the beach surface, slow beach recovery after an episodic erosive event, and a steepening of the beach-nearshore profile. Conversely, for a sector of beach exhibiting a *negative littoral drift gradient* (decreasing rates

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of net littoral drift alongshore), more material is being supplied from updrift than is being transported away in the net drift direction, and the beach will tend toward a positive sediment budget projecting an accretionary beach state with concomitant geomorphic features of well-developed beach berm, well-formed accreting frontal dune, and, in certain morphodynamic circumstances, a well-formed nearshore bar, and low gradient beach-nearshore profile. On a small scale, a littoral drift gradient is manifest as updrift accumulation against a jetty or groin with leeside down-drift erosion.

In situations of marked littoral drift gradient but little long-term beach morphological change, the effects of the littoral drift gradient may be compensated for by onshore or offshore (diabathic) sediment transport mechanisms. Thus on “lee coasts,” onshore sediment drift from the shelf and shoreface, induced by longer period swell waves, assisted by quasi-permanent offshore winds and concomitant bottom return upwelling currents, may mask the erosive effects on the beach induced by a positive (increasing alongshore) net littoral drift gradient. Conversely, for “exposed coasts” facing predominantly onshore winds, the wind forced waves, and downwelling current effects induce an erosive morphodynamic beach state, which may mask the potential beach sediment accretion associated with a negative (decreasing alongshore) net littoral drift gradient.

Cross-References

- [Beach Erosion](#)
- [Littoral Cells](#)
- [Longshore Sediment Transport](#)
- [Wave Focusing](#)

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Longshore Sediment Transport

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Longshore transport refers to the cumulative movement of beach and nearshore sand parallel to the shore by the

combined action of tides, wind, and waves and the shore-parallel currents produced by them. These forces usually result in an almost continuous movement of sand either in suspension or in bedload flows (see entry on “► [Cross-Shore Sediment Transport](#)”). This occurs in a complex, three-dimensional pattern, varying rapidly with time. At any moment, some sand in the area of interest may have an upcoast component while other sand is moving generally downcoast. The separation of the total transport into components parallel and perpendicular to the shore is artificial and is done as a convenience leading to a simpler understanding of a very complex environment. To be meaningful, the rate of longshore transport must be averaged over intervals of at least many wave periods and is typically predicted or measured over much longer times, ranging up to a year (see entries on “► [Gross Transport](#)” and “► [Net Transport](#)”). It is also summed algebraically, at least conceptually, in the direction perpendicular to the shoreline. The result is that the transport rate is defined either as the volume of sand passing a point on the shore in a unit time (Watts 1953), or the immersed weight of sand passing per unit time (Inman and Bagnold 1963). The latter method accounts for the average density of the sand, which can vary significantly between beaches comprised principally of carbonate rather than quartz sand.

The integral of the transport rate over time results in an estimate of the longshore transport in units of volume or weight, as appropriate. If this integration is done without reference to the direction of movement, upcoast or downcoast, the result is termed gross transport (see entry on “► [Gross Transport](#)”). On the other hand, if the sign of the direction is included, the result is termed net transport (see entry on “► [Net Transport](#)”). Net transport can vary from a tiny fraction of the gross transport to a value nearly equal to it, depending on the characteristics of the forcing functions at the site. Estimates of both gross and net values have engineering significance. Jettied harbor entrances, for example, may shoal at a rate related to the gross longshore transport passing them. If the jetties themselves are partial barriers to longshore transport, the net transport can yield guidance to the estimation of accumulation or erosion of sand on either side of the entrance, providing aid in the design of remediation measures.

Longshore transport is the result of a longshore current that conveys sand put in suspension or mobilized on the seabed by waves (see entries on “► [Waves](#)” and “► [Coastal Currents](#)”). The longshore current is usually dominated by the flows induced by waves approaching the shoreline at an angle, although this current can be enhanced or reduced by wind-driven or tidal currents (see entry on “► [Tides](#)”). Larger, more energetic waves mobilize more sand and produce stronger longshore currents so that the magnitude of longshore transport is directly related to the incident wave energy as well as to the angle of wave incidence.

One of the interesting features of longshore transport is that, for the most part, it is impossible to discern directly. The magnitude of the rapid shore-perpendicular motions of the individual sand particles are typically so much greater than the longshore velocity that they prevent the perception of the longshore motion. Even moving back from the particle dynamics and viewing the whole beach face does not provide any clues. If the longshore transport into a reach of shore equals the transport out; the result is no visible change in the configuration of the beach. Only when the transport rate changes along the shore, because of a change in the wave height or approach angle for example, or because of the construction of a barrier such as a groin, does the beach change in a manner than can be readily detected. Typically, these changes occur slowly such that images or measurements over intervals are required to sense them. The terms *littoral drift* or *littoral transport* are occasionally used interchangeably with *longshore transport*.

Prior to the twentieth century, it was generally assumed that tidal currents provided the longshore motivation for beach sand. Komar (1976, pp. 183–190) provides an interesting history of the development of the various explanations for the generation of longshore currents by the oblique approach of waves. The relationship between longshore transport and the properties of the incident waves that formed the basis for most of the research in the second half of the twentieth century was first suggested by Eaton (1951). The longshore transport volumetric rate was assumed to be proportional to the product of the longshore component of the wave energy flux, evaluated at the breaker zone and the sine of the angle that the breakers formed with the shoreline. This product was variously called the longshore component of either the wave power or the wave energy flux. Inman and Bagnold (1963) modified the theory to correct the volume to an immersed weight, which had the benefit of making the constant of proportionality nondimensional.

Local values of longshore transport are difficult to measure because it typically is a mixture of suspended and bedload transport (see entry on “► [Energy and Sediment Budgets of the Global Coastal Zone](#)”). Instruments have been developed to measure suspended sediment concentrations at a point, but

no satisfactory method of measuring bedload has yet been demonstrated. Largely based on inferences from the suspended sediment concentrations, it is generally believed that longshore transport is at a maximum in two zones. One is under the breaking waves and the other is in the swash zone on the beach face.

The simple analytical expressions for longshore transport, which attempt to predict the total transport across a shore-normal line, have been extended into two-dimensional numerical models. These two-dimensional models, in general, contain no additional physical insights into the sediment transport models, but they do allow the prediction of shore evolution caused by gradients in the longshore transport rate (Hanson and Kraus 1989).

Cross-References

- [Beach Processes](#)
- [Coastal Currents](#)
- [Energy and Sediment Budgets of the Global Coastal Zone](#)
- [Gross Transport](#)
- [Littoral Drift Gradient](#)
- [Net Transport](#)
- [Tides](#)
- [Waves](#)

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Machair

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Machair is a Gaelic word which applies to those areas of ancient sand dune systems mainly in the Hebrides of Scotland, but also in west Ireland, where long dune grasses have been superseded and the topography is essentially a low, plain surface (Ritchie 1979). Its main characteristics are summarized below,

1. A level, low-lying, surface at a mature stage of geomorphological evolution, which is part of a very old fully vegetated coastal sand-dune system and is normally marshy in winter.
2. A base of blown sand which has a significant percentage of shell-derived materials and a narrow range of grain sizes.
3. Lime-rich soils with a pH value normally greater than 7.0.
4. A sandy grassland-type vegetation with long dune grasses and other early dune species having been eliminated.
5. Evidence of a history of anthropic interference including heavy grazing (especially rabbits, sheep, and cattle), rotational cultivation and, in places, artificial drainage.
6. A moist, cool location with characteristically strong onshore winds.

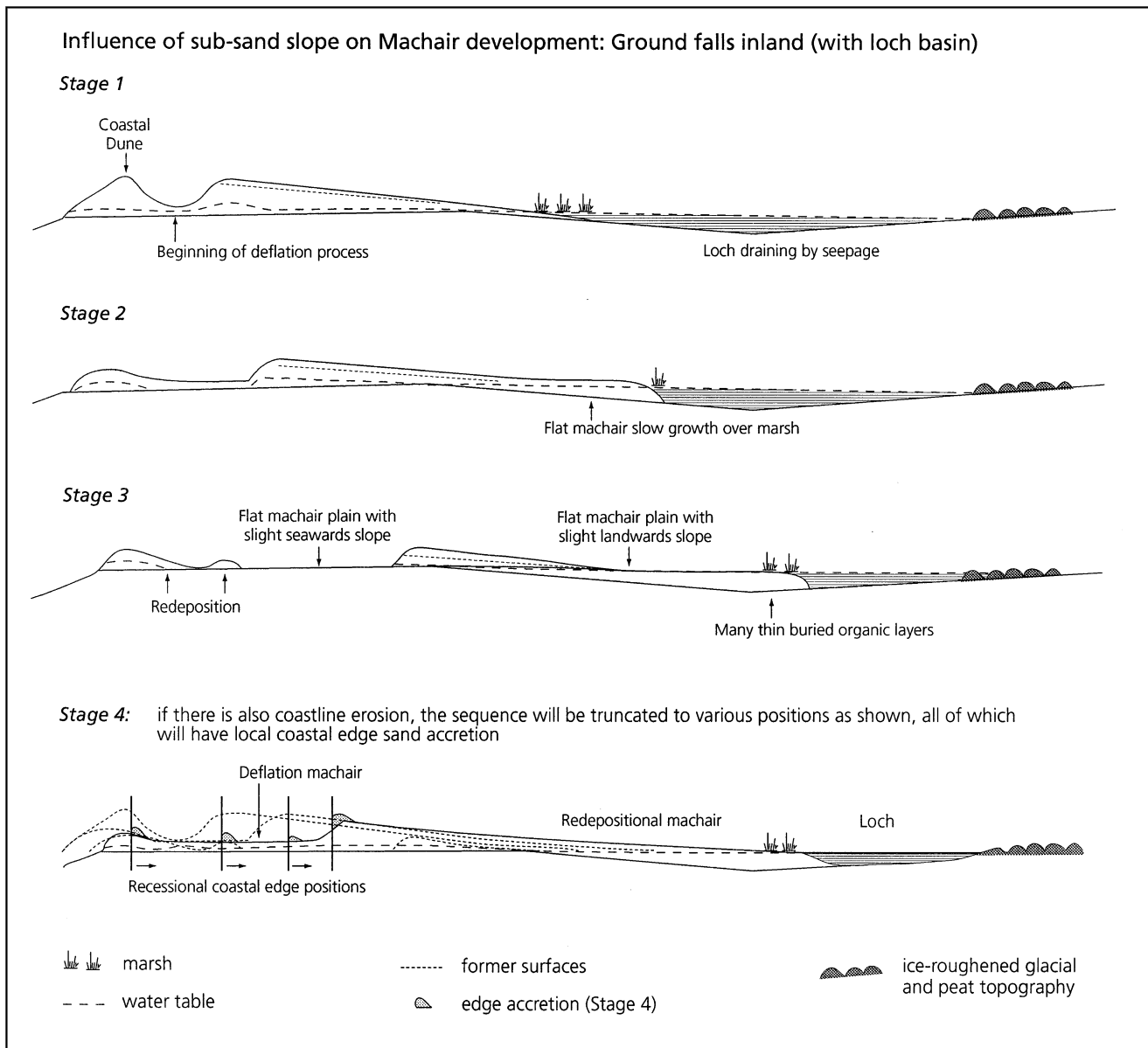
The origin of this dune system was more than 7,500 years ago, as a product of the rapid Flandrian marine transgression which began to level off at this time in this region. Powerful onshore Atlantic wave and wind energy transported preexisting glacial, fluvio-glacial, and shell-derived sediments across the extensive low gradient shallow continental shelf. Machair topography is influenced strongly by the underlying, preexisting landforms of the land extension of this shelf. As the sand-body transgressed landwards old

organic and sand-rich deposits became exposed on the fore-shore. More than 14 sites have been investigated for stratigraphy, pollen, and carbon 14 dating which provide evidence for the sequence of machair evolution (Ritchie 1985).

In the Outer Hebrides the total volume of sand in the beach, dune, and machair system has been more or less constant and, has therefore been subjected to extensive recycling. Nevertheless some ongoing supply from shells on the marine shelf is a possible addition to the original sediment bank. The most commonly used model for machair landforms is summarized in Fig. 1, which shows the distinction between the actively retreating coastal edge and the mature inland surfaces. Typically, machair will occupy more than 90% of the area of the total dune system. Most low machair plains are deflation surfaces which are close to the winter water table. Machair can also be produced at the landward margin of the system by redepositional encroachment into marsh or loch basins.

Submergence has produced strands and sand flats between and inland of some machair systems. The drying-out time at low tide is insufficient for these areas to provide for secondary sand dune development. The open Atlantic beaches, however, do exhibit normal beach and dune exchanges associated with a retreating system. Wind-erosion with net landwards redeposition is also found in some inland features such as sand hills and ridges, normally initiated by severe overgrazing (Angus and Elliott 1992). This erosion–redeposition process is the main mechanism for long-term machair evolution and landwards extension. Human interference including artificial drainage has also affected soils and landforms. There is an extensive literature of archaeological research from Bronze Age times, especially in the Western Isles, which has added greatly to the detailed evidence of machair evolution (Gilbertson et al. 1999).

The study of machair has been dominated by work done in the Outer Hebrides, notably in the Uist Islands where the machair is almost continuous along the entire Atlantic



Machair, Fig. 1 General model of machair evolution

coastline, and its typology has been dominated by the particular circumstances of these islands. In west Ireland, machair often occurs in bayhead locations (Carter 1990). In the Inner Hebrides, islands such as Tiree are almost entirely covered by dunes and machair, and the general model which depends on coastal submergence and finite sand supply may not be wholly applicable. Even on the Uist islands machair has developed on different timescales, as determined by location, offshore bathymetry, and coastal configuration (Ritchie and Whittington 1994).

Machair is not therefore a unique type of coastal sand dune landform. Its age, topography, calcareous soils, and general dependence on geomorphological recycling can be replicated elsewhere but the word “machair” has landscape and cultural

connotations which are associated with a distinctive history of land use and tenure which is now preserved only in the Gaelic fringes of northwestern Europe.

Cross-References

- [Archaeology](#)
- [Changing Sea Levels](#)
- [Dune Ridges](#)
- [Eolian Processes](#)
- [Human Impact on Coasts](#)
- [Sandy Coasts](#)
- [Sea-Level and Climate Change](#)

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Managed Retreat

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Definition

In the coastal zone, **managed retreat** is the application of coastal zone management and mitigation tools designed to move existing and planned development out of the path of both short- and long-term coastal hazards (e.g., hurricanes/typhoons, nor'easters, tsunamis, El Niños, king tides, erosion, flooding, storm surges, sea-level rise). This management strategy is based on a philosophy of avoiding or moving out of harm's way and is proactive in recognizing that the coastal zone dynamics should dictate the type of management employed.

Introduction

Hino et al. (2017) note two defining features of managed retreat: deliberate intervention to manage hazard risk and abandonment of land or relocation of assets. Identifying and mapping the impact zones of coastal hazards is the basis for establishing regulations that govern moving property and people away from migrating and/or storm-impacted shorelines. The mirror of this goal to protect property is that the

shoreline is allowed to advance landward unimpeded which, in turn, conserves coastal natural habitats (dunes, beaches, near-shore ecosystems). In contrast to management strategies aimed at holding the shoreline in place (beach nourishment, shore hardening), managed retreat minimizes financial costs from recurring hazard events.

Unfortunately the term “managed retreat” is often used synonymously with **managed realignment** (Beachapedia 2017; Wikipedia 2017). This latter term should be used in a more restrictive sense where shore-hardening structures are removed selectively to allow natural coastal environments to be maintained or reestablished locally. Ideally, this approach may allow parts of a coastline to erode, maintaining the sediment budget, but most often these plans retain some shore-hardening structures and allow buildings to remain in hazard zones. Esteves (2014) provides a good analysis of the managed realignment approach which has had wide application in Western Europe and the UK. However, an extensive list of managed realignment projects in Europe and the UK (Esteves 2014) shows that these projects are mainly in estuarine settings and involve realignment of defenses, breaching of defenses, or similar modifications to restore tidal flooding. The newly flooded areas are often reclaimed land, mostly in agricultural use. These realignments are not moving villages or developments back from the sea, or allowing unimpeded shoreline transgression. Esteves (2014) does give some genuine examples of managed retreat; all from the USA where buildings were moved landward.

Need for Managed Retreat

The need for managed retreat is as old as the sea-level rise, coastal storms, and erosion. Authors periodically rediscover Thomas Sheppard's 1912 book *The Lost Towns of the Yorkshire Coast* which documented the loss of 28 towns along the Yorkshire Coast, some dating back 2,000 years and some now located 3 miles (about 5 km) offshore (Pilkey et al. 2016). Across the North Sea, the shoreline history of loss is even greater, and these lost villages mirror the present: losses to shoreline erosion and flooding associated with big storms and in response to shore hardening. The responses of these North Sea villages was first to build seawalls and groins, then disorganized retreat or abandonment, but with a few exceptions of organized retreat by removal/relocation of buildings. America's shoreline history is much the same: forced retreat or abandonment as in the islands of Chesapeake Bay and at Edingsville Beach, South Carolina. However, in 1899 Diamond Head, North Carolina, became one of the earliest examples of managed retreat in America when the small community moved entirely off of a barrier island to a more protected island in the sound.

Perhaps in part because of an acceleration in coastal development since the 1950s, greater attention has been paid to coastal land loss. We now know that the modern sea-level rise has accelerated, and in this age of google, there is plenty of visual evidence of the impacts of coastal hazards, plus the known rapid increase in the costs of resulting property losses. And even though there is resistance to the term “retreat,” a disorganized retreat is taking place. Most coastal communities have some missing landmark buildings, now on sites offshore or under an artificial beach. Reviews of tax maps of older US coastal communities show rows of lots now underwater, under a nourished beach, or in a present inlet position.

As early as 1985, the Second Skidaway Conference on America’s Eroding Shoreline concluded:

...the American shoreline is retreating. We face economic and environmental realities that leave us two choices: (1) plan a strategic retreat now, or (2) undertake a vastly expensive program of armoring the coastline and, as required, retreating through a series of unpredictable disasters. (Howard et al. 1985).

That conclusion applies to developed shorelines globally, and the recommendation for strategic retreat is synonymous with managed retreat.

Since that Skidaway Conference, these predictions have proven accurate with the exception that beach nourishment replaced armoring, in part, as the preferred engineering method of stabilizing shorelines. Armoring continues to increase globally, and Dyckman et al. (2014) note in their survey of 30 US coastal states that over 80% still allow shore hardening. Even states that once considered progressive in banning/limiting shore-hardening structures are again allowing the wider use of these structures, although under veiled nomenclature such as temporary sandbag structures (a common response to shoreline erosion at the individual-property level), living shorelines, and terminal groins.

The most widely accepted alternative, beach nourishment, has proven to be costly (PSDS 2017) and has given the deceptive appearance of reduced risk actually encouraging development, and the cost is rising (Armstrong et al. 2016). In the Caribbean and along the US Atlantic and Gulf Coasts, the damage from hurricanes is rising (e.g., Hugo 1989; Andrew 1992; Opal 1995; Georges 1998; Dennis 2005; Katrina 2005; Ike 2008; Sandy 2012; and Matthew 2016). Their impact has induced random retreat at both the individual property and community levels and forced communities to reexamine their coastal zone management strategies. And regardless of the so-called climate change debate, sea level is rising for most of the world’s shorelines, and the rate of rise is accelerating. The beach nourishment approach changes neither where future storms (hurricanes/typhoons/nor’easters) will make their landfall nor the rate of the sea-level rise.

At the close of the twentieth century, a report by the Heinz Center for Science, Economics and the Environment

estimated that 10,000 coastal structures in the USA were within the estimated 10-year erosion zone. As of 1998, coastal counties in the USA, exclusive of the Great Lakes, had a total flood insurance coverage of \$466,874,000,000 (H. John Heinz III Center for Science, Economics and the Environment 2000). The Heinz report (2000) also noted that of the approximately 350,000 structures located within 500 ft of the nation’s 10,000 miles of coastline (including open ocean and Great Lakes), about 87,000 of these buildings, located on low-lying land and bluffs, are likely to erode into the ocean or Great Lakes over a 60-year period. In the coming decades, roughly 1,500 houses, and the land on which they are built, will be lost to erosion each year. During that same period, costs to coastal property owners will average \$530 million per year, in addition to the \$80 million per year spent by the National Flood Insurance Program (NFIP) for erosion-related damage (WHOI 2001). The post-2000 litany of hurricanes (e.g., Katrina 2005; Ike 2008; Sandy 2012; Matthew 2016) and winter storms have shown these projections for the USA are coming true, and in fact the Heinz report (2000) may have been too optimistic.

Historic Shift from Engineering to “Soft” Solutions, but Not Retreat

Trying to hold shorelines in place through engineering by armoring has a long history of failure. Yet it has been less than a century since states and communities began to adopt more stringent building codes to strengthen buildings in the coastal zone and begin to think in terms of coastal management. From the beginning, management has been segmented, both in locale and application, with each community or agency focused on a limited shoreline reach or single problem. On barrier islands, the focus was often on the shoreline rather than a holistic management approach for an entire island or chain of islands. By the 1960s, the US national experience dictated that something be done to control the losses incurred from hurricanes and great storms like the 1962 Ash Wednesday Storm. Impacts on habitat also were being recognized locally as salt marshes and shell fisheries were lost or closed. Results were twofold: the National Flood Insurance Act of 1968 (also the result of persistent property loss on riverine floodplains) and the Coastal Zone Management Act of 1972. Building requirements were upgraded, and many coastal states began to define critical environments and control development through permit processes. Communities and states adopted approaches such as zoning and setback requirements, but systematic retreat was not a common management component. By the early 1980s, Integrated Coastal Zone Management (ICZM) (Clark 1995) or simply Integrated Coastal Management (ICM) defined strategies in which a variety of management tools were being combined. But

even now as then, the times management methods receive serious political attention are mainly after major coastal catastrophes, most often tsunamis, hurricanes, and typhoons (e.g., Indian Ocean tsunami 2004, Hurricanes Katrina 2005 and Sandy 2012, Typhoon Haiyan 2013), rather than addressing management in advance.

Common regulatory methods, such as building codes, requirements for structures to be elevated, and zoning and land use planning controls on development density and type may lessen the impact of storms, but these methods do not remove development from the hazard zone. Raising your house may prevent the storm surge from flooding the living quarters while the waves and currents are eroding the land out from under the house (i.e., vulnerability to the hazard isn't necessarily decreased). Furthermore, these procedures do not recognize shoreline retreat as coastal adjustment takes place in response to sea-level rise, changes in sediment supply, variable wave regime, and other natural controls of shoreline equilibrium. High-density development along shores all over the world demonstrates that land use planning either hasn't worked or coastal management has come after the fact.

Continued beach loss with its associated losses of storm protection, recreational use, aesthetics, and beach economy has led to greater interest in "soft" solutions such as beach nourishment to combat erosion. Beach nourishment, however, is a modern equivalent of the engineering "fix" to hold the line, failing to recognize the natural dynamics of shoreline retreat and the accelerating rise in sea level. The nourishment approach has proven drawbacks including ongoing and increasing costs, diminishing sand supplies, the shorter half-life of nourished beaches, detrimental environmental effects as a result of dredging and sand placement (e.g., loss of shell beds, hard grounds, sea grass meadows, turtle and bird nesting habitats), and masking the risks from coastal hazards (Armstrong et al. 2016). Beach nourishment has not proven to be a panacea. As projects run short on available sand for continued re-nourishment, and/or when financial resources are exhausted or withdrawn (e.g., continued federal support for beach nourishment is tenuous at best), communities will again be faced with the seashore at their front steps.

In short, the best option for many coastal communities and properties is managed retreat. While "retreat" strikes a negative chord for some, elements of the strategic retreat option increasingly are being incorporated into coastal zone management. We are learning the lessons of the old North Sea villages and America's nineteenth-century barrier-island communities – managed or unmanaged, retreat will occur.

Methods of Managed Retreat

Managed retreat implies applying a management strategy from a menu of appropriate tools, often in some combination,

including stabilization techniques in some cases (e.g., certain urban shorelines). Mitigation of coastal hazards through managed retreat is built around one or more fundamental methods (Table 1) with respect to moving property and people out of high hazard zones: avoidance, abandonment, and relocation. These approaches utilize appropriate management tools, such as building restrictions, setbacks, land acquisition (buyouts), and disincentives to development. Management actions are usually planned/active (e.g., building relocated before any storms occur), but unplanned (passive) actions may be part of a retreat plan in anticipation of coming hazard events. For example, the "do nothing" approach is to let the building fall into the sea when the time comes (no protection allowed). Which buildings that will suffer this fate aren't known in advance, but a retreat plan can include such losses. For some urban shorelines, stabilization techniques might be included (e.g., managed realignment), but retreat remains the ultimate solution. Siders (2013) lists 20 coastal management tools to promote managed retreat, including incorporating plans that promote the three general approaches shown in Table 1.

Managed Retreat, Table 1 Primary methods of managed retreat including unplanned forced retreat as potentially part of management plans. These broader management methods should include other aspects such as ecosystem conservation/renewal, removal of shore-hardening structures, limits on artificial beach nourishment, and managed realignment

Management		Methods
		Avoidance
Active	Planned	Ban development/redevelopment in hazard zones (no-build areas: setbacks, rolling easements, zoning)
		Abandonment
Passive	Unplanned	Loss of development/redevelopment by ongoing attrition (erosion, flooding)
Active	Planned	Post-event condemnation/removal/reconstruction limits
Active	Planned	Emergency pre-event removal
		Relocation
Passive	Unplanned	Post-event relocation of salvageable buildings
Active	Planned	Pre-event relocation (short to intermediate term)
Active	Planned	Systematic relocation of community out of hazard zone (long-term) including acquisitions: planned pre- and post-event property buy-outs in hazard zones (no future construction), and rolling setbacks and easements: planned regulatory construction setback based on defined hazards and periodically revised, based on changing hazard zones (e.g., sea-level rise, storm surge levels, erosion rates, evacuation requirements)

Avoidance

Avoiding coastal hazards is the best way to eliminate losses of property and lives. The decision not to locate in a hazardous area may not seem like managed retreat, but defining areas where no development is allowed because of specific hazards, critical habitats, or limited sediment sources should be part of any managed retreat plan. Defined **No-Build Areas** are the common management tool for avoidance (Rabenold 2013; Siders 2013; NOAA 2012) and include setbacks, rolling easements, and zoning. However, these methods may be short term and result in only partial avoidance of hazards.

Setbacks (fixed and rolling), as the name implies, are a management tool to keep structures out of extreme-to-high hazard zones, or at least at a distance from the hazardous processes, such as storm surge and V-zone flooding (NFIP flood maps) and coastal erosion. A fixed setback simply requires that new construction be a fixed distance inland from a reference line (e.g., the back of the beach, the vegetation line, the crest of the dune line). The regulatory line is fixed, and not adjusted for changes such as those brought about by storms or net ongoing coastal retreat by erosion. A rolling setback is one in which the regulatory line shifts landward as the shoreline erodes (i.e., the bluff edge, back beach, or dune toe retreats). Coastal management (setback) lines can influence how existing development is maintained over time and how new development will be allowed to proceed. Although setbacks often are defined as creating zones in which no buildings or structures are allowed, in reality most setback regulations allow for variance applications, and in some jurisdictions, liberal granting of variances circumvents management intent.

How far back is a “safe” building setback? The answer is difficult to ascertain and will vary from place to place according to erosion rates and state and local regulations. No uniformity exists between coastal states’ setback regulations in terms of how they are defined or applied (for state-by-state summaries, see Rabenold 2013; NOAA 2012). In the USA, most states base setbacks on past average erosion rates which may no longer be valid given variables such as the loss of sediment supply, increased storminess, and sea-level rise. Establishment of setback distances are often determined as much by political and socioeconomic pressures as they are by defined hazard zones (Thomas 2015).

Although fixed setbacks put some distance between buildings and the shore, that distance does not remain constant. When shoreline retreat catches up to the buildings, the original setback distance is of no consequence. So, once again, the relocation or abandonment options must be considered. A rolling setback line is adjusted periodically, and no new construction is allowed within the no-build zone, but existing

buildings, that may be grandfathered in at the time the setback is established, ultimately are threatened.

Rolling easements are similar to the above in that they are adjusted as the defining regulatory line migrates landward. Easements include other aspects of land use to which owners agree (e.g., limits on future development, access, armoring, building replacement) (Titus 2011). **Zoning** regulations also can contribute to the avoidance approach by requiring safe siting away from coastal hazards and incorporating setbacks and easements.

No-build zones conserve the sacrificial eroding shoreline which is more than just a buffer zone. Erosion of the back-beach upland, bluff, or sea cliff is usually an indispensable source of sediment supply to the facing and down-drift beaches.

Disincentives may be used in an effort to encourage people not to build in high-hazard zones or in areas of critical habitat. Federal laws such as the 1982 Coastal Barrier Resources Act (COBRA) and the 1990 Coastal Barriers Improvement Act (CoBIA) designated areas in which development is allowed, but buildings there are not eligible for federal flood insurance or any post-storm federal assistance such as small business loans and funding to rebuild infrastructure. After Hurricane Fran in 1996, however, such assistance did go to the community of North Topsail Beach, NC, which is located in a COBRA unit.

Abandonment

Abandonment may be unplanned or part of a planned strategy of retreat. Historically, abandonment is often an unplanned, post-storm response to destruction of buildings and land loss such as bluff retreat so that reconstruction is impossible. Fallen houses in the water or on the beach are a common sight along open-ocean shorelines after hurricanes and nor’easters. Similar scenes are common in the Great Lakes during periods of high lake levels and along large embayments like the Chesapeake Bay. Destruction may be so complete that the property is abandoned. Ruins of buildings destroyed in Hurricane Gilbert, 1988, remained 12 years later along Mexico’s Yucatán Coast. In some cases, entire villages on barrier islands and along eroding bluffed shorelines have been abandoned. The village of Broadwater, Hog Island, Virginia, was abandoned in 1933 after losses to storms and shoreline erosion, although in part it was a short-term planned abandonment as some houses were relocated off the island. Earlier, Cobb Island, Virginia, and Edingsville Beach, South Carolina, met with similar fates. Abandoned towns on the West coast of the USA include Bayocean, Oregon, that went into the sea building by building from 1932 to 1960, and Cove Point, Washington. Like the England’s Yorkshire coast villages, these were unplanned abandonments.

Planned abandonment can be incorporated into managed retreat in several ways. Setbacks sometimes are regarded as “retreat” because they may disallow reconstruction in the no-build zone and also are viewed as denying property owners the right to construct shore-hardening structures, thereby forcing abandonment. Long-term planned abandonment can follow what is sometimes called the “do nothing” approach, in which buildings are regarded as having a fixed life span and when their time comes to fall into the sea, bay, or lake, no attempt is made to protect them. Buildings are razed either just before or after failing. Emergency programs can encourage buildings to be moved prior to a hazard event to avoid economic loss. Planned abandonment also can be achieved by prohibiting post-storm reconstruction or by requiring relocation landward of the revised post-storm setback control line. The original South Carolina Beachfront Management Act would only allow habitable structures damaged beyond repair (two-thirds or greater damaged) to be rebuilt landward of the no-construction zone. In part because of a poorly written law, post-Hurricane Hugo enforcement led to the famous court case of *Lucas v. South Carolina Coastal Council* in which the plaintiff prevailed, resulting in the law being rewritten and softened. Rebuilding after storms can be discouraged by other methods as well, such as denial of flood insurance and other subsidy programs. Removal of buildings that are still standing on the beach or in the water is often achieved through condemnation because of no accommodation for septic systems or sewer lines, or safe access.

Relocation

Relocation is the ultimate form of managed retreat. As noted above, **active relocation** is undertaken by moving a building back either before it is threatened, or, if threatened, before it is damaged. In contrast, **passive relocation** is achieved by rebuilding a destroyed structure in another area, away from the shore, and out of the coastal hazard zone.

Long-term relocation usually implies a broader strategy through community zoning or land use plans that identify a frontal zone of buildings likely to be affected by known erosion rates or predicted flood levels from storm surge and coastal flooding. These buildings are then scheduled for relocation over an extensive period of time (assuming they will not be lost in coming storms). In effect, this is an engineered retreat, and on a barrier island, the plan may include creation of new land on the sound side of the island for relocating the structures. However, artificial island migration is achieved with this method. Because barrier islands are usually backed by sensitive marshes and wetlands, this approach is questionable and raises complicated issues of property rights and changing ownership. This kind of strategy is more easily achieved when moving communities off of riverine floodplains and non-barrier coasts. As noted above, even where

setbacks are used, a retreating shoreline will catch up with the property, and relocation will again become an option.

Relocation is often the best economic option even though the up-front cost may be high. One can find examples along almost every shoreline where armoring is used in which the cost of seawalls, groins, and breakwaters or nourishment, over the lifetime of the property, exceeds the value of the property and greatly exceeds the cost of moving the structure.

Hino et al. (2017) give international examples where governments have supported the relocation strategy. In the USA there was FEMA support for moving buildings as early as the 1987 Upton-Jones Amendment to the National Flood Insurance Program (NFIP) that allowed owners of threatened buildings to use up to 40% of the federally insured value of their homes for building relocation purposes. This law recognized relocation as a more economical, more permanent, and more realistic way of dealing with long-term erosion and flood problems. The NFIP would pay a relatively small amount to assist in relocating or razing a threatened house, rather than paying a larger amount to help rebuild it only to see the rebuilt house destroyed in a subsequent storm and then paying to rebuild again. The program ended in 1995, but has been replaced by similar government programs. Some states encourage relocation with similar programs, or require houses to be moveable through the building permit process, thus allowing future relocation.

For large structures the relocation alternative often is regarded as technically impossible, but the International Association of Structural Movers says that moving large structures is technologically feasible, although expensive. The relocation of lighthouses demonstrates that large structures can be moved, and such relocation is not a new mitigation strategy. Lighthouses have been moved to new sites in North America since the nineteenth century. Recent examples include Cape Cod (Highland) and Nauset Lighthouses 1996, Massachusetts; Cape Hatteras Lighthouse 1999, North Carolina; and Gay Head Lighthouse 2015, Massachusetts (Beachapedia 2017). Numerous buildings, services, and infrastructure in national, state, and local parks have been relocated (Beachapedia 2017). Moving houses and communities off of riverine floodplains is not uncommon (e.g., English, Indiana; Rhineland, Missouri; Glasgow, Virginia). To date, most relocations of coastal communities have been partial at best and restricted to small communities. Such relocation by community decision is rare, but an exception was the relocation of El Choncho on Colombia’s Pacific coast (Correa and Gozalez 2000).

At the individual property level, **deep property lots** are an important element in planning for future relocation. In effect, lot depth determines possible future on-site relocation. Deep lots allow homeowners to relocate houses threatened by

erosion to another location on their own property. While relatively deep lots are found in some coastal communities, new developments are often designed to maximize the number of dwelling units, resulting in small lots. Despite this trend, some communities now require deep lots (ocean side to sound side on barrier islands) in order to provide for relocation (e.g., Nags Head, North Carolina).

Acquisition has become an important component of managed retreat plans. Land in the public trust through federal, state, and local ownership usually provides benefits in terms of conservation, providing public access to the shore, contributing to recreational and tourism needs, preserving aesthetics, and protecting habitat. In the USA, most coastal states have land acquisition programs, and governments can purchase land through negotiated agreements where owners voluntarily sell land, or, less common, by eminent domain (condemnation proceedings). Other strategies used by the various governing bodies include tax incentives, donations of conservation easements, trading of land, and transference of development rights (easements). Condemnation usually results in a much higher cost for the land. Most land acquisition programs are hampered by a lack of funding, but two states with fairly successful programs are Florida and California. Post-storm buyouts also are becoming more common as in Texas after Hurricane Ike, 2008, and New York after Hurricane Sandy, 2012 (Pilkey et al. 2016).

Examples of actual planned relocation of communities are hard to come by, and the few that can be cited are small villages such as El Choncho, Colombia (Correa and Gozalez 2000), or the study examples of Hino et al. (2017) including Vunidogoloa, Fiji, and the Carteret Islands in the Pacific (Iverson 2013). As of 2016 there were several noteworthy plans for relocation of communities; however, some of these projects have languished for years. The aftermath of Katrina led to the planned move of Isle de Jean Charles, Louisiana, but no resettlement has taken place. Plans for moving several coastal villages in Alaska have been in the works for years (Iverson 2013). These small communities face a triple threat due to climate change: less sea ice, rising sea level, and thawing permafrost – all leading to accelerated erosion.

Piecemeal building-by-building relocation is more common. Nags Head, North Carolina, provides an early example of a community that incorporated retreat aspects into its management plans. The town adopted building standards more restrictive than required by either FEMA or the North Carolina Coastal Area Management Act (CAMA). Incentives encouraged development to be located as far back from the ocean as possible, including strict setbacks. Because small, single-family structures are easier to move, the town has limited the development of oceanfront hotels and condominiums. Deep lots running perpendicular to the shore provide considerable room for relocation. Prior to rebuilding

after a storm, the town may require adjoining lots in common ownership to be combined into a single lot. New construction of wood frame, multistory, multi-family buildings is not permitted. Strict limits are set on the amount of impervious surfaces within the oceanfront zoning districts. Post-storm measures include building moratoriums, policies on reconstruction, and land acquisition. Relocating property is an old concept in Nags Head as the Outlaw House, named for the family who owned it, has been moved back 600 ft from the retreating shoreline in five separate moves over 100 years.

Impediments to managed retreat remain greater than the incentives to move people and property out of high-hazard zones. Nearly every study of managed retreat in the coastal zone recognizes a long list of obstacles to the adoption of this management strategy (Dyckman et al. 2014; Gibbs 2016; Hino et al. 2017). Most of these impediments are political and socioeconomic, rather than management constraints (e.g., deep lot depths or alternative building sites, moveable structures). Some of the more common obstacles include the following: entrenched property interests and escalating land values with property viewed as growth assets; local governments desire for continued growth (expanding the tax base); local governments with dominant representation by persons with vested interests in development; government programs intended to suppress development in hazard zones that actually encourage the opposite (subsidized flood insurance, support for beach nourishment, post-disaster support for restoring infrastructure); intentional vague or contradictory management legislation (lack of political will); lack of appropriate authority for managers and financial support for management programs; and the general trend for “taking” challenges vs. property rights to lose in the courts.

Conclusion: Managed Retreat: Reality or Myth?

Some of the tools of managed retreat are questionable as to whether or not they are meeting the ideal retreat concept. Elevating structures, stronger building codes, living shorelines, and most managed realignments are examples of management tools sometimes listed in managed retreat plans, but these approaches do not move buildings and infrastructure out of harm's way. No-build zones are often short-term buffer zones at best and are often backed by buildings that ultimately are unprotected as these buffer zones narrow. Abandonment tends to be unplanned retreat, driven by storm destruction of buildings, or erosional retreat of the shoreline leaving buildings stranded on the beach or in the surf. Condemnation may result in removal, but the retreat is haphazard. Only planned relocation is an ideal retreat.

Hino et al. (2017) concluded that about 1.3 million people have relocated in the previous 30 years through managed

retreat, based on 27 global examples studied. However, their examples are drawn from all forms of geologic hazards (e.g., riverine flooding, earthquakes, landslides), and the resulting relocation of populations was generally not planned managed retreat, but rather disorganized and unplanned responses to major impact events. Much of the relocation after the 2011 Tōhoku (Sendai) Japan earthquake and tsunami was the result of the nuclear plant meltdowns. Examples of planned pre-event managed retreat of coastal communities are few. Even piecemeal retreat is random, depending on which buildings or infrastructures are destroyed in hazard events.

As long as beach replenishment is subsidized by the government, and economically feasible with ample sand supplies, communities will opt to hold shorelines in place for the revenue they generate and artificial high property values. But declining sand supplies, escalating project costs, and the end of government subsidies will make beach nourishment a less acceptable management tool.

In summary, to hold the line against the sea-level rise for all of the world's developed shorelines is unrealistic. Managed retreat provides the best set of tools for mitigating coastal hazards and reducing property losses. Avoidance remains the best solution for undeveloped and lightly developed areas, while various forms of abandonment and relocation are the long-term solution for even urbanized shores. Coastal land acquisition is one method of meeting both of these objectives; however, ever more funding will be needed for future acquisitions to be successful. Setback lines are a temporary solution, even when redefined periodically as rolling setbacks. In order for managed relocation to work, integrated land use planning and zoning efforts will have to take a broader, holistic approach. For example, barrier-island management policies must consider the entire island, moving from a focus on site-specific and linear (island front) regulation to a whole-island perspective and from shore-hardening/hold-the-line programs to approaches which concentrate on preservation, augmentation, and repair of the natural systems. The time has come for serious consideration to adopt managed retreat alternatives.

Cross-References

- Beach Erosion
- Beach Nourishment
- Coastal Boundaries
- Coastal Management Practices
- Global Vulnerability Analysis
- Sea-Level and Climate Change
- Setbacks
- Small Islands

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Mangroves, Ecology

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Introduction

Mangroves have always been considered as marginal ecosystems for at least three main reasons. First, the global mangrove area does not exceed 180,000 km² representing less than 2% of the world's tropical forest resources. Second, their discontinuous distribution, at the land and sea interface of tropical and subtropical coastlines, is primarily characterized by tidal regimes, which is a unique forest habitat. Third, the frequent wide fluctuations of environmental factors (dissolved oxygen, salinity, organic, and inorganic suspended matter) have induced in mangrove flora, a complex range of adaptations, lacking in other woody species, unable to compete or to survive in these highly variable and adverse environmental conditions (low oxygen content in soils, sulfate toxicity, high NaCl in water and soils, exposure to hurricanes and surges, muddy soils, instability, etc.). Yet, these ecosystems are highly productive with an average primary productivity often higher than that of neighboring continental forest types.

Many species of invertebrates and vertebrates of commercial value use mangrove habitat for food and shelter during their life cycle. Most mangrove ecosystems around the world have been depleted during the twentieth century. Until the 1980s they have been extensively converted to other uses. For the last 20 years, many mangrove areas have come under full or partial protection, and restoration programs are being implemented in almost every one of the 70 countries possessing mangroves.

Most of the mechanisms and processes regulating mangrove ecosystems; primary productivity, food webs, nutrient fluxes, physiological adaptations of plants and animals, etc. are still poorly known, and this fragmentary knowledge is mainly restricted to species of commercial value.

Present Distribution of Mangroves

Six geographical zones have been recognized (Chapman 1976; Snedaker 1982; Rao 1987; Saenger and Bellan 1995; Duke et al. 1998).

With rare exceptions, mangroves are restricted to coastal areas where mean monthly air temperatures, in winter, are higher than 20 °C and where ground frost is unknown (Fig. 1). The tallest (up to 35 m tall) and more dense mangroves are found in bioclimatic conditions with high annual rainfall (>2000 mm) and a short dry season (<3 dry months). They can survive in arid areas (Persian Gulf, Mauritania, Red Sea), in the form of low or dwarf, monospecific stands (Dodd et al. 1999). The largest contiguous surface area of mangroves, covering more than 6,000 sq. km, is located in the upper Bay of Bengal, on the delta of the Ganges.

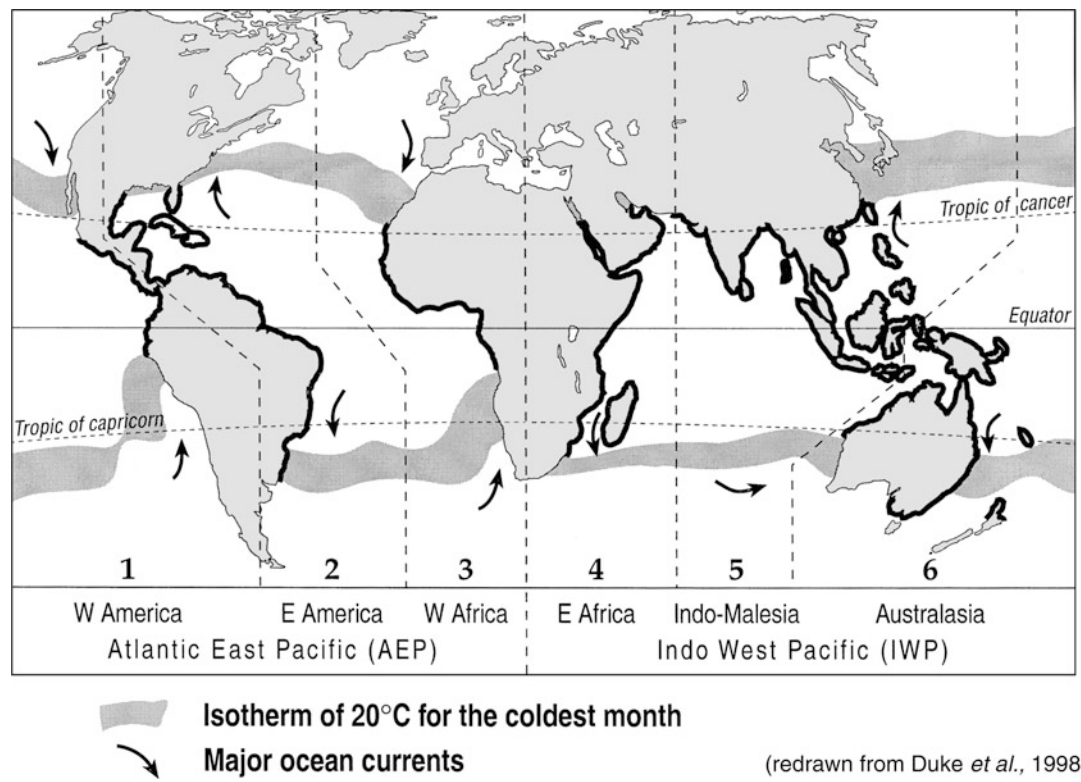
Recent estimates (Spalding et al. 1997) indicate that the total mangrove area is about 180,000 sq. km, most of it being located in South and Southeast Asia (Table 1). A few nations dominate these area statistics. For example, of the approximately 70 countries with this ecosystem, Indonesia, Australia, Brazil, Nigeria have about 43% of the world's mangroves. Indonesia alone has 23% and 12 countries have two-thirds (Table 1). Political and management decisions relating to mangrove stands of these countries will have significant effects on the global status in the future (Hamilton and Snedaker 1984). It is assumed that at least 30% of these ecosystems are degraded or very degraded.

Productivity Goods and Services of Mangrove Forests

Our general knowledge of mangrove structural properties, above ground biomass and litter production (Saenger and Snedaker 1993) is rather well-advanced as extensive work has been carried out, especially in Malaysia (Sassekumar and Loi 1983), Australia and New Zealand (Duke 1988; Woodroffe et al. 1988), USA (Lugo et al. 1980; Twilley 1982; Lahmann 1988), Brazil (Adaime 1985), French Guyana (Fromard et al. 1998). The following general conclusions can be drawn:

- in equatorial and sub-equatorial areas, mangrove height is about 12–20 m, rarely exceeding 25 m, and the litterfall varies from 12 to 16 t ha⁻¹ year⁻¹ (dry material).
- near the tropics, their average height is about 8–12 m and the known mean annual litterfall is of the order of 8 t ha⁻¹.
- along warm temperate coastlines, where mangroves are exceptionally found (New Zealand), their height does not exceed 4 m and the litterfall declines to 4 t ha⁻¹ year⁻¹.

High productivity and short retention times of organic matter in most mangrove ecosystems make them extremely important, not only for fisheries (Snedaker and Lugo 1973). In the last 25 years, it has been shown that mangroves



Mangroves, Ecology, Fig. 1 Main biogeographic mangrove area. (Redrawn from Duke *et al.* 1998)

throughout the world serve a multitude of functions for the people inhabiting them (ISME 1993), but more and more foresters and shrimp farmers are looking at the profits that can be derived from converting the mangroves into fish and shrimps ponds. As clearly shown by recent world satellite surveys (Spalding *et al.* 1997; Blasco and Aizpuru 2001) conversion of mangroves to aquaculture, to agriculture, and to urban development are the main causes of mangrove destruction.

Large-scale conversion to agriculture is conspicuous in the Gangetic delta (India and Bangladesh) where the population density exceeds 800 inhabitants km⁻², in Myanmar (Irrawady delta) and in most coastal areas of West Africa, from Senegal to Nigeria. Paddy fields, sugar cane, orchards, and oil palm are the main products. Concerning the conversion to aquaculture, fish ponds (*Chanos chanos*) have replaced most mangroves in the Philippines, whereas shrimp farming in Thailand, Indonesia (*Penaeus monodon*, *Penaeus merguensis*) and Equador (*Penaeus vannamei*) has converted large areas of mangroves in recent years. In 1977, it was estimated that 1.2 million a of mangroves in the Indo-pacific region has been converted to aquaculture ponds (Saenger *et al.* 1983). In many countries (Vietnam, Malaysia, Thailand, etc.) highly pyritic soils have lead to high acidity, causing major difficulties for the operators of aquaculture ponds with spectacular declines in yields (less than 400 kg ha⁻¹ year⁻¹

Mangroves, Ecology, Table 1 Estimates of mangroves areas. (After Spalding *et al.* 1997)

Region	Global (km ²)
South and Southeast Asia	75,173 (41.5%)
Australasia	18,789 (10.4%)
The Americas	49,096 (27.1%)
West Africa	27,995 (15.5%)
East Africa and the Middle East	10,024 (5.5%)
Total area	181,077
Major countries	(km ²)
Indonesia	42,500
Australia	11,500
Brazil	13,800
Nigeria	10,500
Malaysia	6,400
India	6,700
Bangladesh	6,300
Cuba	5,500
Mexico	5,300
Papua New Guinea	4,100
Colombia	3,600
Guinea	2,900
Total	119,100

instead of 1,400 initially). And the development of acid sulfate soils in drained mangrove areas of West Africa has had catastrophic consequences with failure of rice crops.

M

Mangroves Versus Chronic Coastal Environmental Problems

Many investigators have described the building role of *Rhizophora*. The actual efficiency of sediment retention is very unequal from one place to another. The coastal afforestation program which is in progress in Bangladesh (Saenger and Siddiqi 1993), with more than 200,000 ha, demonstrates that plantations with *Sonneratia apetala* contribute both to the acceleration of land accretion and to the stabilization of the exposed soft sediments.

In the last few years, new findings have been published concerning the role of mangroves as biochemical barriers to pollutants. Although metal mass balances in mangroves are still at an exploratory stage, with net import and net export occurring in different oceanographic cycles, the final balance, after a long period of monitoring, seems to be a net import of trace metals, as a result of complex physical and chemical coastal processes (Clark et al. 1997; Lacerda 1998).

Recent studies in coastal Brazil have shown that large heavy-metal concentrations (Hg and Zn) are retained in the rhizosphere sediments under very refractory chemical forms, making their uptake extremely difficult. They are preferably accumulated in perennial tissues, such as below ground biomass and trunks, whereas their concentrations in leaves remain extremely low. Although this is a new field of research there are strong and convergent indications that mangrove soils and plants minimize pollution by heavy metals in tropical coastlines (Lacerda et al. 2000).

Concerning trace metals concentration and distribution in mangrove animals (Zn, Cd, Cu, Cr, and Pb), few studies have been carried out on mangrove oysters (*Crassostrea*) and mussels (*Mytella guyanensis*). No general conclusion can be drawn so far because of the scarcity of reliable data.

Flora, Zonation, and Role of Salinity

The earliest mangrove fossil materials (pollen, fruits, hypocotyls) have been recorded from Brazil, Europe, Asia and Australia, etc. during the lower Cenozoic, about 55 million years before present (Eocene).

Throughout the world's tropics, growing preferably on muddy soils of deltaic coasts, in lagoons and estuarine shorelines, about 70 species of trees and shrubs, including putative hybrids, are considered as exclusive to the mangrove habitat (29 genera and 20 families) and about 24 are important but not exclusive.

The commonest species belong to three genera: *Rhizophora*, with nine taxa bearing conspicuous stilt roots, *Avicennia*, with eight taxa having dense, slender aerial roots known as pneumatophores and *Sonneratia* (nine taxa) with stout, conical pneumatophores (Table 2).

One of the most conspicuous and enigmatical features in mangrove ecosystems is the zonation of species. Such a

common spatial patterning seems to be the result of complex processes involving dispersal strategies, plant succession mechanisms, physiological attributes of species, interspecific interactions, and seasonal fluctuations of physico-chemical parameters (Snedaker 1982; Smith 1992).

The importance of salinity has been stressed by most mangrove specialists (Saenger et al. 1983; Tomlinson 1986; Hutchings and Saenger 1987; Duke et al. 1998). Each species of tree and shrub has its own tolerance to salinity concentrations. This partly explains the upstream and downstream distribution of each species. *Avicennia integra*, *Heritiera fomes*, *Sonneratia lanceolata* occur upstream, or in freshwater-dominated habitats, where tidal penetration is limited and the average annual water salinity is lower than 20 ppt. The majority of remaining mangrove species (Table 2) are found in the intermediate part of the estuaries (all members of the Rhizophoraceae, *Avicennia alba*, *Avicennia officinalis*, *Excoecaria*, *Lumnitzera*, *Pelliciera*, *Xylocarpus*, etc.), indicating that their optimum salt requirements oscillate between 20 and 30 ppt. A few woody species are most generally found at the mouth of estuaries, in downstream and sea-front sectors, thriving year round in highly saline environmental conditions (30–45 ppt); this is the case for *Avicennia marina*, *Rhizophora mangle*, *Bruguiera gymnorhiza*, *Laguncularia racemosa*, and *Sonneratia alba*.

Avicennia marina has the largest, almost continuous, biogeographical distribution of any mangrove species, extending from east Africa to the Red Sea and Pakistan, to the Indian Ocean and the Southern Pacific. Some of its ecotypes are found in the Arabian Gulf, occupying one of the driest mangrove habitats in the world, in which salt concentrations may reach levels that are beyond the physiological limit for other species (Dodd et al. 1999). Under laboratory conditions, *A. marina* can survive a very wide range of salinities (Ball 1996, 1998), from distilled water to salt concentrations as high as 175 ppt. The exact mechanisms of the physiological adaptation to saline environments, involving osmoregulation, ion compartmentation and selective ion uptake by roots, salt excretion, are still unclear.

Several attempts have been made to correlate species richness with important environmental factors, especially in Southeast Asia and Australia (Duke 1992; Ball and Pidsley 1998). Although temperature, rainfall, tidal and soil peculiarities, water salinity, etc. are known to be determining factors, the interplay between such factors and species richness is not properly understood.

Mangrove Fauna

Most mangrove fauna can be subdivided in three main groups: resident species (occurring primarily in mangroves), seasonal migrants, and occasional mangrove species. In contrast to the list of mangrove plants given in Table 2, associated

Mangroves, Ecology, Table 2 Trees and shrubs of the world's mangroves. (Based on Saenger et al. 1983; Duke et al. 1998)

Flora		Form	Biogeographic regions (see Fig. 1)					
Genus	Species							
<i>Acanthus</i>	<i>ebracteatus</i> Vahl.	S					5	6
	<i>ilicifolius</i> L.	S					5	6
<i>Acrostichum</i>	<i>aureum</i> L.	F	1	2	3	4	5	6
	<i>danaeifolium</i> Langsd.	F	1	2				
	<i>speciosum</i> Willd.	F					5	6
<i>Aegialitis</i>	<i>annulata</i> R. Br.	S					5	6
	<i>rotundifolia</i> Roxb.	S					5	
<i>Aegiceras</i>	<i>corniculatum</i> (L.) Bl.	S					5	6
<i>Aglaia</i>	<i>cucullata</i> Roxb.	T					5	
<i>Avicennia</i>	<i>alba</i> Blume	T					5	6
	<i>bicolor</i> Standl.	T	1					
	<i>germinans</i> L.	T	1	2	3			
	<i>integra</i>	T						6
	<i>marina</i> Forsk. Vierh	S/T	1			4	5	6
	<i>officinalis</i> L.	T					5	6
	<i>rumphiana</i> Hall. f.	T					5	6
	<i>schaueriana</i> Stapf.	T		2				
<i>Bruguiera</i>	<i>cylindrica</i> (L.) Blume	T					5	6
	<i>exaristata</i> Ding Hou	S/T						6
	<i>gymnorhiza</i> (L.) Lam.	T				4	5	6
	<i>hainesii</i> C.G. Rogers	T					5	6
	<i>parviflora</i> Roxb.	T					5	6
	<i>sexangula</i> (Lour.) Poiret	T					5	6
<i>Campostemon</i>	<i>philippensis</i> Becc.	T					5	
	<i>schultzei</i> Mast.	T						6
<i>Ceriops</i>	<i>australis</i>	S/T						6
	<i>decandra</i> Griff.	S/T					5	6
	<i>tagal</i> C.B. Rob.	S/T				4	5	6
<i>Conocarpus</i>	<i>erectus</i> L.	S/T	1	2	3			
<i>Cynometra</i>	<i>iripa</i> Kostel	S					5	6
<i>Diospyros</i>	<i>littoralis</i>	T						6
<i>Dolichandrone</i>	<i>spathacea</i> K.	T					5	6
<i>Excoecaria</i>	<i>agallocha</i> L.	T					5	6
	<i>indica</i> Muell. Arg.	T					5	
<i>Heritiera</i>	<i>fomes</i> Buch. Ham.	T					5	
	<i>globosa</i> Kost.	T					5	
	<i>littoralis</i> Aiton	T				4	5	6
<i>Kandelia</i>	<i>candel</i> L. Druce	S/T					5	
<i>Laguncularia</i>	<i>racemosa</i> Gaertn.f.	S/T	1	2	3			
<i>Lumnitzera</i>	<i>littorea</i> Voigt	S/T					5	6
	<i>racemosa</i> Willd.	S/T				4	5	6
	× <i>rosea</i>	S						6
<i>Mora</i>	<i>oleifera</i> Ducke	T	1					
<i>Nypa</i>	<i>fruticans</i> Van Wurmb.	P		2	3		5	6
<i>Osbornia</i>	<i>octodonia</i> F. Muell.	S/T					5	6
<i>Pelliciera</i>	<i>rhizophorae</i> Pl. Triana	T	1	2				
<i>Pemphis</i>	<i>acidula</i> Forster	S				4	5	6
<i>Phoenix</i>	<i>paludosa</i> Roxb.	P						6
<i>Rhizophora</i>	<i>apiculata</i> Blume	T					5	6
	<i>mangle</i> L.	S/T	1	2	3			
	<i>mucronata</i> Lam.	T				4	5	6
	<i>racemosa</i> G. Mey	T	1	2	3			
	<i>samoensis</i> (Horchr.) S.	T						6

(continued)

Mangroves, Ecology, Table 2 (continued)

Flora		Form	Biogeographic regions (see Fig. 1)					
Genus	Species							
	<i>stylosa</i> Griff.	S/T					5	6
	× <i>harrisonii</i>	T	1	2	3			
	× <i>lamarckii</i>	T					5	6
	× <i>selata</i>	T						6
<i>Scyphiphora</i>	<i>hydrophyllacea</i> Gaertn.	S					5	6
<i>Sonneratia</i>	<i>alba</i> J. Smith	T				4	5	6
	<i>apetala</i> Buch.-Ham.	T				5		
	<i>caseolaris</i> (L.) Engl.	T				5	6	
	<i>griffithii</i> Kurz	T				5		
	<i>lanceolata</i> Bl.	T				5	6	
	<i>ovata</i> Backer	T				5	6	
	× <i>gulgai</i>	T				5	6	
	× <i>urama</i>	T					6	
	<i>alba</i> × <i>ovata</i>	T				5		
<i>Tabebuia</i>	<i>Palustris</i>	S	1					
<i>Xylocarpus</i>	<i>granatum</i> Koenig	T			4	5	6	
	<i>mekongensis</i>	T				5	6	

P, Palm; S, Shrub; F, Fern; T, Tree

biota recorded from mangrove often occur in other habitats, adjacent to mangrove.

As a general rule, many species of terrestrial mammals are found in mangrove ecosystems but none seem to be strictly confined to mangroves. Even the estuarine crocodile (*Crocodylus porosus*) one of the largest species, and the endangered Bengal tiger (*Panthera tigris tigris*), common in the mangroves of the Ganges, are not restricted to these ecosystems. In the same way, fruit bats and nectar-feeding bats which often roost in mangroves, the crab-eating Macaque (*Macaca fascicularis*) and the Proboscis monkey (*Nasalis larvatus*) are more abundant in mangroves than in other habitats.

The use of mangrove ecosystems as nursery grounds for larval and juvenile fish was first demonstrated by Odum and Heald (1972) in Florida. The group of fish restricted to mangroves is mainly that of mudskippers, belonging to the gobiid subfamily Oxycneminae, with two genera (*Periophthalmus* and *Periophthalmodon*) and 5–10 species. The main peculiarity of mudskippers which are common practically throughout tropical mangroves, is their ability to withstand exposure to air especially during low tide. According to Hutchings and Saenger (1987) the main importance of mangroves to fish, is the availability of food, especially prawns, for juveniles, and protection, as piscivores are often under-represented in mangroves. Fish fauna is now rather well-known for most mangrove regions (Tholot 1996).

Most mangrove birds are found in a variety of other habitats but mangroves provide secure roosting sites at high tide and the food for many species. The most spectacular population is probably that of scarlet ibis (*Eudocimus ruber*) in Trinidad's Caroni mangrove, gathering more than 20,000 birds.

Mangroves are known to serve as nursery grounds for young crabs and juvenile prawns and as a source of seed for aquaculture. The mud crab (*Scylla serrata*) is probably the most popular Crustacea in coastal Southeast Asian countries. Among molluscs, several oysters (*Crassostrea commercialis*, *Crassostrea lugubris*, *Crassostrea iredalei*, *Crassostrea malabonensis*) and the green mussels (*Perna vividis*) are cultured commercially, whereas the cockle (*Anadara granosa*) has a great commercial value, particularly in Malaysia and Thailand.

Air-breathing Arthropods (Myriapods, Arachnids, and Insects), are probably extremely numerous in all mangroves (Murphy 1985) in terms of specific biodiversity, but little work has been done so far on the insect fauna. They usually remain in air-filled cavities during high tide; a conspicuous "cave fauna" has been described in the mud-lobster mounds (*Thalassina*). Diptera are richly represented in these ecosystems where highly distinctive and unique genera have been identified; some families are also characteristic of mangroves (Tethinidae, Canaceidae, etc.). Likewise, among Lepidoptera, some butterflies and moths are found only in mangroves, the most diversified being perhaps among Nymphulinae. Finally, some wasp and bee taxa (Hymenoptera) are endemic to mangroves where they may play a noteworthy commercial role (high quantities of natural honey are collected from the mangroves, in Bangladesh).

The complex interactions between invertebrates and plants are very poorly known (Robertson et al. 1990), unless they are conspicuous. Many propagules (*Rhizophora*, *Bruguiera*, *Avicennia*) are destroyed by borer beetles and a tortricoid moth causes mass mortality among *Sonneratia* seedlings.

Mangrove Microorganisms

Microbiota are known to play essential roles in mangrove food chain and biogeochemical cycles (carbon, sulfur, nitrogen, iron, phosphorus, etc.), under very peculiar anaerobic and aerobic conditions. Fungi and bacteria may become associated, in mangrove sediments, providing intensive, rapid, and almost continuous recycling of organic matter, during floods, under anoxic conditions, as well as during exposure to air at low tides. Given the high temperature of water and sediments (25–35 °C), the primary production is rapidly attacked and decayed.

The role of fungi in the processes of degradation of lignocellulose materials, is still obscure; very little is known, as few species have been tested for their ability to degrade the organic matter (Jones and Hyde 1988). Even the taxonomic inventory is only now beginning. About 100 species are known today (73 Ascomycotina, 25 Deuteromycotina, and only 2 Basidiomycotina).

From a biogeographical point of view it appears that most mangrove fungi are distributed across the tropical world (Jones 1996). This could be due either to their very efficient dispersal mechanisms or to an early evolution, before the separation of landmasses.

According to Boto (1988) much further research is needed to properly understand, the exact role of microbiota, especially of bacteria, in mangroves.

Conclusion

These are a few of the many areas that briefly illustrate the extreme complexity of mangrove ecosystems and the numerous issues to be solved. There are a number of poorly understood processes involved in the productivity and biogeography of mangrove species. Some preliminary studies (Duke 1995; Maguire and Saenger 2000), are indicating infraspecific variation having a genetic basis. The exact genetic variation within mangrove taxa is unknown.

But there is a danger that most of the natural processes and interactions will never be properly understood, because the mangroves of the world, except in Australia, are being transformed, fragmented, degraded, or converted at a very rapid pace, as a consequence of the population growth in the tropical coastal world (Field 1996).

Cross-References

- Coral Reef Coasts
- Desert Coasts
- Mangroves, Remote Sensing
- Vegetated Coasts

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Mangroves, Geomorphology

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Definition

Mangroves are halophytic shrubs and trees that grow in the upper part of the intertidal zone on the shores of estuaries and lagoons, and on coasts sheltered from strong wave action, as in inlets or embayments or in the lee of headlands, islands, or reefs. Mangroves can influence coastal geomorphology by

trapping sediment to form depositional intertidal terraces that prograde the coastline.

Mangrove Ecosystems

Mangroves show their greatest extent and diversity on tropical coasts, where they occupy a similar niche to temperate salt marshes (Saenger 2002). They grow sparsely on rocky shores and coral reefs, where their roots penetrate fractures in the rock, and on sandy substrates, but are more luxuriant, forming dense scrub and woodland communities, on muddy substrates and shoals exposed at low tide, particularly where there is a supply of muddy sediment. Where wave energy is low, they spread forward to the mid-tide line, but as wave action increases along a coastline, the mangrove fringe thins out and disappears. On the other hand, mangroves colonize areas that have become more sheltered as the result of the longshore growth of sand bars, spits, or barriers.

The width of a mangrove fringe generally increases with tide range, and on macrotidal coasts can attain several kilometers, as on the tide-dominated shores of gulfs and estuaries in northern Australia, where wide mangrove areas are backed by sparse salt marshes and saline flats flooded during exceptionally high tides and summer rains. Within the humid tropics, mangroves grow to forests with trees 30–40 m high, as on the west coasts of Malaysia and Thailand, in Indonesia, Madagascar, and Ecuador. On drier and cooler coasts within, and outside, the tropics (extending to about 30°N and up to 39°S), mangroves generally form extensive scrub communities (Chapman 1976).

Where mangroves are spreading seaward, there is an abundance of seedlings and young shrubs on the adjacent mudflats and a smooth canopy rising landward as the trees increase in age and size. A receding mangrove coast is indicated by a microcliff and the truncation of the mangrove canopy, so that trunks are exposed with trunks and roots being undercut, or where the mangroves have died, and any seedlings fail to survive.

Mangroves are physically adapted to survive in a marine tidal environment. Some species (e.g., *Avicennia marina*, *Sonneratia alba*) have root systems with networks of pneumatophores, snorkel-like breathing tubes that project vertically from the muddy substrate (Fig. 1), and allow the plant to respire when the tide falls. With the aid of these structures, mangroves can grow in areas submerged by the sea at each high tide, but the need for several hours subaerial exposure between each submergence sets a seaward limit, usually close to mid-tide level. Pneumatophores occupy a roughly circular zone around the stem of each plant and can be very numerous and closely spaced, with densities of up to 300/m². Other mangroves, such as *Rhizophora*, *Bruguiera*, *Ceriops*, and *Lumnitzera* species, have subaerial prop or stilt roots that branch downward to the mud and support the stems.

Mangroves can be damaged by storm surges or tsunamis. There was extensive damage to mangrove areas in Sri Lanka during the Indian Ocean tsunami of 2004. Restoration of mangroves on coasts where they have been damaged or eroded requires the application of ecological engineering (Lewis 2004; Jin-Eong 2004) to restore the habitat for mangrove plantations.

Formation of Mangrove Terraces

Sediment carried into mangroves by a rising tide is retained by the filtering network of stems, pneumatophores, or prop roots as the tide ebbs. This leads to the gradual building up of a depositional terrace between high spring and high neap tides, with a transverse slope (usually $<1:50$), descending seaward to the outer edge of the mangroves and continuing across the lower intertidal zone, which is either unvegetated or has patches of seagrass. Terrace formation is also aided by the presence of a subsurface root network, which binds the accreting sediment. The depositional terrace under mangroves is similar to that formed beneath salt marshes (q.v.).

Mangroves initially colonize substrates that are stable or slowly accreting, but once established, they promote further accretion because they diminish current flow and wave action (Augustinus 1995). A mangrove fringe also shelters near-shore waters from winds blowing off the land, and thus reduces seaward losses of mud from the shore, whereas shoreward drifting by waves produced by onshore winds and rising tides continues.

On some coasts, the pioneer mangroves are *Avicennia* spp., with pneumatophores that trap muddy sediment and raise the

intertidal surface until it is colonized by *Rhizophora* and other mangrove species (Fig. 2). On other coasts, *Rhizophora* species occupy the seaward margin. To the rear, sedimentation proceeds more slowly, but eventually, the terrace is built up to high spring tide level, where the mangroves may be backed by forest, salt marsh, or unvegetated dry mudflats. Most mangrove substrates are dominated by muddy sediment, but there is also peat accumulation where the mangrove community generates large quantities of organic matter from decaying leaves, stems, and roots, and from the various organisms that inhabit mangrove areas. In due course, these can form peat deposits, raising the substrate level; on the coast of Florida, *Rhizophora* is growing on vertically accreting peat deposits.

As the terrace attains high spring tide level, it is submerged only by infrequent exceptionally high tides, occasional storm surges, or river flooding. Sedimentation is thus very slow, but accretion of peat and drift litter may raise the substrate to levels where it can be colonized by freshwater and land vegetation, a process that is aided by emergence on coasts where sea level is falling relative to the land.

Sections through mangrove terraces (exposed in the banks of tidal creeks or in microcliffs at the seaward edge) generally show stratified deposits, with layers of fine sand or organic material within the mud. These variations are related to wave conditions, storm waves washing fine sand into the mangroves, and mud accretion continuing as the tides rise and fall in calm weather. Although often characterized as tide-dominated morphology, most mangrove terraces are influenced by wave action as the tide rises and falls. When the depositional terrace is submerged, occasional storm surges may wash sand, gravel, and shelly material up into the mangroves to be deposited as cheniers (q.v.).

Mangroves, Geomorphology,

Fig. 1 Pneumatophores projecting from a muddy intertidal surface around *Avicennia marina* shrubs in Westernport Bay, southeastern Australia. (Copyright-Geostudies)



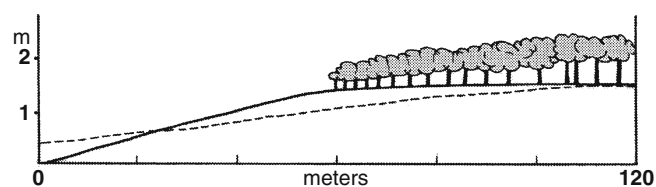
Mangroves, Geomorphology, Fig. 2 Mangroves (*Avicennia marina*) advancing on to intertidal mudflats in Cairns Bay, north-eastern Australia. The darker zones to the rear are *Rhizophora* and other mangrove species. (Copyright-Geostudies)



Former cheniers may be indicated by lenticular deposits of coarser sediment within mud in mangrove terrace stratigraphy.

Rates of accretion in mangroves can be measured by laying down marker layers of colored sand, coal dust, or similar material on the surface, and returning to put down borings and measure the thickness of sediment added subsequently. There are often difficulties where burrowing crabs churn up the substrate. Changes can also be measured on implanted stakes. In Westernport Bay, Australia, such measurements showed sustained mud accretion of up to 4.5 cm/yr in the *Avicennia* fringe, with slower deposition at higher levels and continuing vertical fluctuations on adjacent mudflats. In the absence of mangrove vegetation, the substrate remains a mobile intertidal slope. The pattern of mud accretion is strongly correlated with the density of the pneumatophore networks, which trap and retain muddy sediment, and low accretion mounds form above the general level of the muddy shore within pneumatophore networks around isolated *Avicennia* trees. When pneumatophore networks were simulated artificially by planting a network of wooden stakes, a similar accretion mound formed within the staked area, but when the stakes were removed, the accreted sediment was quickly dispersed. Mangroves with pneumatophores thus trap sediment as effectively as salt marsh plants (Bird 1986). Mangroves with prop roots, such as *Rhizophora* spp., may be less effective in trapping mud, but depositional terraces have been formed on coasts fringed by such mangroves.

The widening of a mangrove terrace depends on a continuing sediment supply, as on deltaic and estuarine shores close to the mouths of rivers or within embayments where intertidal muddy areas are extensive. Changes within river catchments, such as deforestation, increase soil erosion and augment the rate of sediment yield to the coast, thereby



Mangroves, Geomorphology, Fig. 3 A mangrove terrace formed by vertical accretion of sediment as mangroves spread seaward. If the mangroves then die, or are removed, the intertidal profile is degraded (dashed line). If they revive, the mangrove terrace is restored. (Copyright-Geostudies)

accelerating accretion in intertidal areas and promoting the spread of mangroves. This has occurred in the Segara Anakan lagoon in southern Java, where a greatly increased sediment yield resulted in siltation and the rapid advance of the mangrove fringe (Bird and Ongkosongko 1980).

If mangroves are killed or cleared, their substrate is soon lowered by erosion. In Westernport Bay, Australia, mangroves were extensively cleared in the mid-nineteenth century, and the depositional terraces that had formed beneath them were degraded to a steeper transverse slope and dissected, so that the roots of former mangroves were laid bare. Subsequent recolonization of these areas by mangroves was followed by mud accretion, which has rebuilt the depositional terrace (Fig. 3) (Bird 1986).

Nevertheless, the idea that mangroves are land-builders has not been universally accepted (see discussions by Vaughan 1910; Davis 1940; Carlton 1974). Some have suggested that mangroves merely occupy intertidal areas that become ecologically suitable as the surface is raised by accretion, independently of any effects of vegetation, implying that the depositional terraces would have developed even if mangroves had not been present (Watson 1928;

Scholl 1968). Others have deduced an interaction between colonizing mangroves and intertidal deposition (Thom 1967). It is possible that *Avicennia* and other mangroves with pneumatophores promote accretion and coastline progradation as they spread forward on to the intertidal zone, whereas *Rhizophora* and other mangroves without pneumatophores are less effective in trapping sediment.

Tidal Creeks in Mangrove Areas

As mangrove terraces build up in the form of a sedimentary wedge, there are alternations of tidal submergence as water invades the mangroves and emergence as it drains off. The ebb and flow of the tide forms and maintains a system of tidal creeks, the dimensions of which are related to the volume of water entering and leaving as the tide rises and falls. Typically dendritic and intricately meandering, these are channels within which the tide rises until the water floods the marsh surface; then they become drainage channels into which some of the ebbing water flows from the mangroves. They are thus like minor estuaries, particularly, where they receive freshwater from hinterland runoff. Where cheniers are present, the mangrove creek network may be reticulate, as on the Niger delta (Allen 1965), while straight parallel creeks are more often found where the tide range is large, the transverse gradient small, or the rate of seaward spread of mangroves rapid, as on Hinchinbrook Island, north-eastern Australia.

In the early stages of the formation of a mangrove terrace, tidal creeks are relatively wide and shallow in cross section (like the broad channels on intertidal mudflats), but as the mangrove terrace expands, they become narrower and deeper, and their banks higher and steeper. Oversteepening results in local slumping, when blocks of compact mud, often with clumps of mangrove vegetation, collapse into the creek, especially where the banks have been burrowed by crabs.

If the mangrove vegetation dies, or is cleared away, the tidal creeks become wider and shallower, and if revegetation then occurs, they again narrow and deepen (Fig. 4). There are also changes as tidal creeks meander or migrate laterally, so that mangrove trees are undermined on one bank while

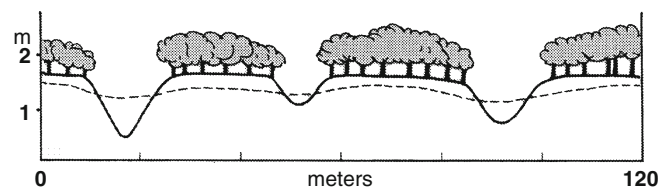
mangrove seedlings colonize sediment deposited on the other. Channel banks may oscillate in response to variations in the volume of tidal ebb and flow; they advance during phases of local accretion when the channel diminishes, and recede during episodes of tidal scour, when the channel is enlarged.

Seaward Margins of Mangroves

Mangrove terraces are being eroded on coasts that are now receiving little or no sediment. Low cliffs have been cut into their seaward margins, particularly on deltaic coasts where the sediment supply has been reduced because of dam construction or the natural or artificial diversion of a river outlet. The seaward edge of mangrove terraces is often undercut by a muddy microcliff up to a meter high (Fig. 5). Microcliff recession may be accompanied by continuing vertical accretion of muddy sediment in the mangroves, building up the terrace even though seaward advance has come to an end. In some places, the cliffing results from lateral movement of a tidal channel, undercutting the outer edge of the mangroves, but generally, it is due to larger waves reaching the mangroves as the result of deepening of the lower intertidal zone, either because of progressive entrapment of nearshore sediment drifting into the upper vegetated area or because of continuing submergence of the coast in response to a rising sea level (Guilcher 1979).

It may be that, as on the sides of developing tidal creeks, seaward margins become oversteepened and cliffed, particularly during occasional storm wave episodes. Cliffling of this kind is repaired if there is an abundant supply of sediment to restore the profile, permitting mangroves to spread again, but if there is a sediment deficit, a mangrove cliff will persist and recede until the mangrove terrace has been completely removed.

If global warming proceeds, mangroves are likely to spread to suitable habitats beyond their present poleward limits. A rising sea level will however, impede the seaward advance of mangroves, and increase erosion on their seaward margins, except where there is a compensating increase in sediment supply. It has been estimated that a sea-level rise of



Mangroves, Geomorphology, Fig. 4 Transverse section across a mangrove terrace with intervening tidal creeks. If the mangroves die, or are removed, the creeks become wider and shallower (dashed line). If they revive, their cross-profile is restored. (Copyright-Geostudies)

Mangroves, Geomorphology,

Fig. 5 Erosion of the seaward edge of a mangrove terrace, exposing root structures. (Copyright-Geostudies)



more than 1.2 mm/yr will lead to widespread destruction of mangroves and erosion of their substrates (Ellison and Stoddart 1991).

Summary

Mangroves are halophytic shrubs and trees that grow in the upper part of the intertidal zone on the shores of estuaries and lagoons, and on coasts sheltered from strong wave action.

They are extensive on tropical coasts but extend locally on to temperate coasts. They can influence coastal geomorphology by trapping sediment (including peat produced by mangroves) to form depositional intertidal terraces that prograde the coastline. This process is assisted where mangroves have aerial roots (pneumatophores) or buttress roots. Rates of accretion in mangroves can be measured by laying down marker layers of colored sand, coal dust, or similar material on the surface, and returning to put down borings and measure the thickness of sediment added subsequently. Where mangroves are spreading seaward, there is an abundance of seedlings and young shrubs on the adjacent mudflats and a smooth canopy rising landward as the trees increase in age and size. A receding mangrove coast is indicated by a microcliff and the truncation of the mangrove canopy, so that trunks are exposed with trunks and roots being undercut, or where the mangroves have died, and any seedlings fail to survive.

Cross-References

- [Cheniers](#)
- [Mangroves, Ecology](#)

- [Mangroves, Remote Sensing](#)
- [Muddy Coasts](#)
- [Peat](#)
- [Salt Marsh](#)
- [Tides](#)
- [Vegetated Coasts](#)

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Mangroves, Remote Sensing

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Introduction

Most research activities and published documents (at least 95%) on the biology and ecology of mangroves concern the factors influencing the productivity, the biodiversity, and the biogeographical distribution. Naturally, the biodiversity itself and the adaptive mechanisms to salinity and waterlogging constitute major fields of research. In comparison, published documents on the use of remote sensing for mangrove studies are in a minority.

In recent years, “Remote Sensing” has become a term applied to all kinds of information acquired by satellites, although in its broad sense, it refers to the gathering and analysis of images acquired by sensors and cameras located at some distance from the target of study including aircrafts, balloons, etc. (Haines-Young 1994). The main advantages of satellite over aerial photographs have often been advocated, especially by foresters, who rightly insisted that large-scale surveys are feasible through computer processing. In addition, monitoring events such as erosion, degradation processes, human impacts, flooding, phenology, etc. can be carried out as each satellite makes regular overhead passes on a fixed orbit.

A brief review of the most commonly used remote sensing technology for mangrove studies is given here. The capabilities and constraints of spaceborne instruments are outlined. This entry focuses only on conspicuous achievements from a few square kilometers to the whole tropical coastal world.

Sensors Characteristics

The application of remote sensing tools to mangrove forests began tentatively in the early 1970s with some optimistic

attempts to use Landsat MSS data, with a ground resolution of about 60–80 m (“*ground or spatial resolution*” is “*the minimum distance between two objects that a sensor can record distinctly*”- Simonett 1983), in huge mangrove areas, especially in the Gangetic delta which has the largest contiguous mangroves in the world. The main hope was to increase the frequency of forest observations and inventories, especially in remote fast changing areas, where access and field-work are particularly difficult. At best, the repeated frequency was 16 days and 4 spectral bands were available in the visible (VIS) and near-infrared (NIR) wavelengths. Since 1982, Landsat satellites have an instrument (Thematic Mapper) with a ground resolution of 30 m and 7 spectral bands including middle-infrared (1.57–1.78 μm) and a thermal-infrared (2.10–2.23 μm). After the launching in 1986 of the first high-resolution satellite, SPOT 1, operating at 20 m in a multispectral mode and at 10 m in a panchromatic mode, and an improved repeat cycle, the possibility of acquiring accurate data, for almost all mangroves of the world, has been conspicuously increased. This is how the first World Mangrove Atlas was achieved and published (Spalding et al. 1997), followed by the first worldwide mangrove inventory carried out by the European Community (Aizpuru et al. 2000).

These instruments, including radar, generate a signal only in each of a small number of very broad bands (Table 1). In the case of radar, SAR (Synthetic Aperture Radar) data, the spatial resolution is rather good (6–25 m) and these products are widely available since the end of the 1990s from a number of recent spacecrafts (ERS-1, ERS-2, JERS, Radarsat ...). Mangroves have already been studied at various research levels with radar, which have the ability to penetrate clouds, permitting very frequent repetitive observations. Most of these instruments are primarily dedicated to physical oceanography and polar observation. Their use for environmental applications is yet very limited. When SAR data are used for mangrove studies, they are often processed in combination with optical data as one of the major issues is the suppression of the random noise associated with SAR data (the speckle), which induces strong limitations for the delineation of coastal units and land cover types (Mougin et al. 1993; Kushwaha et al. 2000; Phinn et al. 2000; Proisy et al. 2000).

More complex hyperspectral scanning systems are now providing new types of data sets. They measure the intensity of the radiations received from coastal ecosystems in each of a large number and very narrow bands (50–300 bands, about 2 or 3 nm each). MODIS (MODerate resolution Imaging Spectro-radiometer) is already in space on EOS (Earth Orbiting System) satellite launched in 1999. Several comparable instruments are airborne: VIFIS (Variable Interference Filter Imaging Spectrometer) with 64 spectral bands, AVIRIS (Airborne Visible InfraRed Imaging Spectrometer), CASI (Compact Airborne Spectrographic Imager), etc. Since 1981, tens of thousands of photographs, including color

Mangroves, Remote Sensing, Table 1 A summary of the main satellites used for vegetation mapping

Satellites, sensors, launch year, and distance to earth	Band	Spectral bands	Spatial resolution	Overhead passes and swath	Suitable mapping scale	Discrimination
AVHRR	1	0.58–0.68 μm			Global and regional	
NOAA	2	0.725–1.1 μm	HRPT, LAC: 1 km	Daily		Forest/nonforest
(Since 1978)	3	3.55–3.95 μm	GAC: 4 km			
(860 km)	4	10.5–11.3 μm	GVI: 15 km	2,700 km	1/1,000,000	Biological rhythms
	5	11.5–12.5 μm			–1/10,000,000	
LANDSAT MSS	4	0.45–0.6 μm		16 days	National	
(1972)	5	0.56–0.7 μm	56 $\times$ 79 m			Physiognomy
(915 and 705 km)	6	0.67–0.8 μm		180 km	1/200,000–1/1,000,000	
	7	0.78–1.1 μm				
LANDSAT TM	1	0.45–0.52 μm				
(1982)	2	0.53–0.61 μm				
(705 km)	3	0.62–0.69 μm	30 m	16 days	Local	Phenology
	4	0.78–0.91 μm				
	5	1.57–1.78 μm		180 km		Physiognomy
	7	2.10–2.35 μm				
	6	10.4–12.6 μm				
SPOT 1, 2, 3	1	0.500–0.590 μm		26 days		Dominant floristic
HRV	2	0.615–0.680 μm	20 m		1/50,000–1/200,000	groups
(1986)	3	0.790–0.890 μm		60 km		
(830 km)	P	0.510–0.730 μm	10 m			
ERS-1(1991)	SAR	5.3 GHz			Local	Physiography
and ERS-2 (1995)			30 m	100 km	1/50,000	Vegetation cover?
(785 km)	C	5–7 cm			1/100,000	Soil water content
Radarsat (1995)	SAR	5.3 GHz	Multi-resolutions	24 days	1/50,000	Same +
(798 km)	C	5–7 cm	25–100 m	100–500 km	1/250,000	crops
JERS-1 (1992)	SAR	1.2 GHz	20 m	44 days	1/50,000	Same
(568 km)	L	24 cm		75 km		
ENVISAT (2000)	SAR	5.3 GHz	multi-resolutions	35 days	1/50,000	Same + coastal zones topography
(800 km)	C	5–7 cm	25 m	100–400 km	–	
			150–1 km		1/1,000,000	
	B0	Blue 0.44–0.47 μm				
SPOT4	B2	Red 0.61–0.68 μm	1 km	daily	All scales	
VEGETATION	B3	PIR 0.79–0.89 μm				Forest types
	MIR	MIR 1.58–1.73 μm		2,250 km	since	Main crops
and						Phenology
					1/25,000	Physiognomy
SPOT4	B1	Green 0.50–0.59 μm				Dominant floristic
HRVIR	B2	Red 0.61–0.68 μm	20 m		–	groups
	B3	PIR 0.79–0.89 μm		60 km		Fire monitoring,
(1998)	MIR	MIR 1.58–1.73 μm			1/5,000,000	etc.
(830 km)	P	0.59–0.75 μm	10 m			

infrared, have been taken from the space shuttle orbit and stereoscopic coverages are available. To date, very high-ground resolution data (about 1 m) have not been made available to scientists mainly for security reasons. They are progressively appearing on the market (IKONOS data).

All these sophisticated instruments are generating an incredible amount of digital georeferenced data, covering all the ecosystems of the coastal world in a repetitive manner. Have we been able to develop the necessary technology to adapt the quality and the Timescale of observation to the magnitude and rapidity of human-induced degradations in mangrove areas?

Basic Principles

Each satellite has its own technical properties designed for specific missions. Whatever satellites are used, the physical principle remains almost constant, based on the fact that different mangrove subtypes show different reflectance patterns. Spaceborne sensors measure the solar radiation that are reflected or radiated by the “targets.” The “reflectance” of a given ecosystem (a mangrove with *Nypa fruticans* or a salt marsh with *Chenopodiaceae*) is the ratio between the reflected solar energy and the incident solar energy. In order to minimize the distortions induced on each signal by the atmosphere and to take into account the sensors-object-sun geometry, several corrections have to be applied to raw recorded data before they are processed (i.e., atmospheric and geometric corrections). Ultimately, it is quite obvious from Table 1 that spatial and temporal resolutions determine the type of information that can be derived from satellites (Holben and Fraser 1983; Graetz 1990; Blasco and Aizpuru 2001). It is implicit that permanent global monitoring of mangroves, at high spatial resolutions (Landsat TM, SPOT HRV and HRVIR, IRS1C-LISS-III,

ERS, etc.), is practically not feasible and that local mangrove monitoring at low spatial resolution (NOAA-AVHRR, SPOT-VEGETATION) is also impossible (Justice et al. 1985; Townshend and Justice 1986; Malingreau et al. 1989).

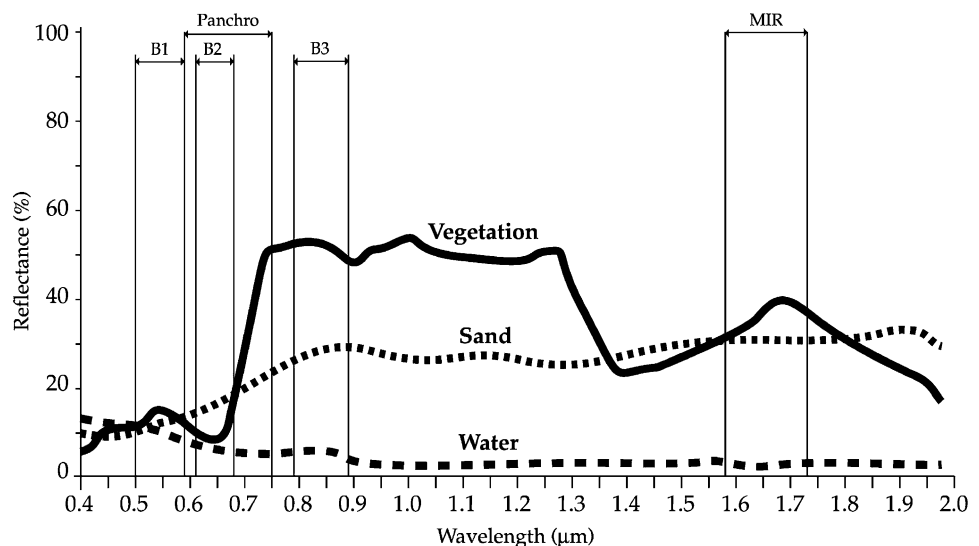
The different coastal ground units reflect differently in the VIS parts of the spectrum (dense continental green trees, wet clayey soils, sandy beaches, mangroves . . .) as well as across all the wavelengths from the ultraviolet to the thermal-infrared and microwaves (Fig. 1). Some signals are not reflected but are radiated by ground units especially in the thermal-infrared. The different patterns of reflectance from different ecosystems are often termed “spectral signatures.”

Usually the “signal mangrove” is distinct from others coastal ecosystems for two main reasons:

1. It is necessarily confined to the nearshore tropical intertidal zone.
2. It is the result of two main signatures often recognised in the VIS and NIR domains:
 - The evergreen character of mangrove trees leads to strong reflectance in the NIR channel (0.7–1.1 μm), generally coded in red on a color composite.
3. The permanently wet soils which have a noteworthy reflectance in the red channel (0.6–0.7 μm), generally coded in green on a color composite.

This is why dense mangroves always appear dominantly in a reddish violet color, turning to bluish violet in open mangrove stands. A given tall dense mangrove stand with *Rhizophora* sp. (Malaysia) may have the same “signature” as a low thicket with *Avicennia marina* (New Zealand). This is one of the major limitations of these technologies, especially in forestry in general, where one of the main goals is to discriminate the species of trees and to estimate the structure of the forest and the size of its components.

Mangroves, Remote Sensing,
Fig. 1 SPOT4-HRVIR: Spectral signatures from representative land cover types in mangrove areas



In order to transfer experimental results to field situations, several “vegetation indices” have been proposed. The most commonly used is the Normalized Difference Vegetation Index (NDVI). It relies upon differential reflectances of the mangrove canopy (which is always green) in the red wavelength (here the response is mostly determined by the absorption band by the chlorophyll) and in the NIR wavelength, where the response is the result of scattering determined by the cuticles of leaves and the density of the cover.

$$\text{NDVI} = \frac{\text{NIR} - \text{VIS}(\text{red})}{\text{NIR} + \text{VIS}(\text{red})}$$

The thermal domain (4–12 μm or bigger wavelength range) has not been exploited so far in mangrove deforestation processes, although a local sharp increase in surface brightness temperature could most probably be related to the replacement of dense mangrove stands by the so called saltish “blanks” which are barren soils, almost denuded of any vegetation. One of the limitations of the thermal domain is due to the fact that the relationship between the size of barren mangrove area and the temperature of the concerned pixels is difficult to establish. Thermal-infrared radiation allows rapid measurement of variations of water surface temperature. On the other hand, most features to be analyzed and mapped in coastal areas are narrow and linear. They are not always observed from space and may be lost at inadequate resolutions.

To date, the most common resolutions of satellites used for coastal surveys range between 10 and 30 m (SPOT, IRS, Landsat TM). When these resolutions are available, they lead to rather accurate maps at 1/50,000–1/100,000 scales. When these resolutions are not available, a number of statistical problems arise linked to difficult comparisons of maps drawn at different scales.

For each object and each pixel, the ecologist has at his disposal its reflectance, its longitude and latitude. This is why remote sensing data are usually exploited with the help of Geographic Information Systems (GIS). These databases and software combine time-independent parameters (geology, physiography, average climatic parameters, etc.) and a number of parameters whose values vary with time (phenology of trees, stand density, radiometry, etc.).

A lot of frameworks have been proposed to guide the existing classification systems; they combine various concepts from remote sensing, plant biogeography, landscape ecology, etc. (Phinn et al. 2000).

Processing Methods

In terms of forestry and vegetation classification, mangroves belong to the broad class of “closed evergreen tropical forest.” Mangrove subtypes are identified according to the species

composition, the structure of the stand and the ecological status (salinity of water and soils, climatic aridity, etc.), but neither the floristic composition nor structural properties nor environmental parameters are perceptible from space, which identifies only the spectral and textural parameters of the top canopy. This explains the low statistical separability of dense mangrove subclasses in the VIS and NIR wavelengths of high-resolution satellite data. This relates to the canopy architecture and the textured nature of reflectance patterns in which variance within forest classes can be greater than between forest classes (Hill 1999; Kay et al. 1991). Then, the easiest way to increase the spectral separability is to consider that the criteria of density of the cover is the starting point of any mangrove classification system.

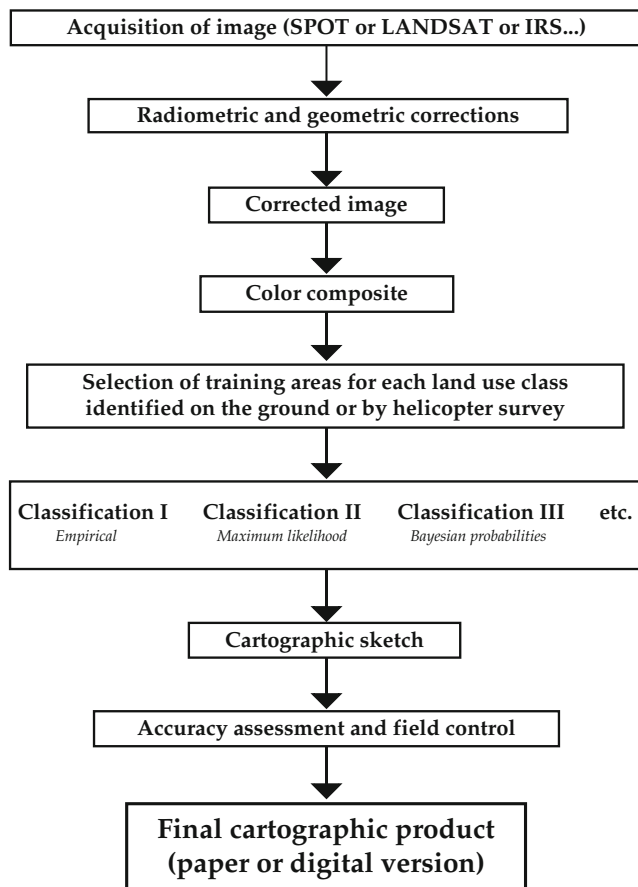
Many approaches have been suggested for satellite image processing. None of them is yet fully convincing. They logically fall into four main groups:

1. *The visual interpretation* is probably the commonest but least explicit. In this case, a digital image has been converted into a photographic product which is studied by eye, with reference to field data. The delineations obtained with this empirical method are often interpreter-dependent.
2. *In the unsupervised classification*, the data are digitally processed with automatic enhancement techniques. One of major limitations of these approaches is caused by the fact that there are sometimes no significant difference between the spectral properties of mangroves and other neighboring ecosystems.
3. *Supervised classifications* are the most frequently described in the scientific literature. Field data, assisted or not with a Global Positioning Systems (GPS), are used for the selection of training samples to direct this type of classification, using several algorithms such as the maximum likelihood, the minimum distance, etc. (Figs. 2 and 3). Practically, the scientists use existing maps as a reference in order to obtain a preliminary image segmentation, separating mangrove areas from other plant communities.
4. Finally, *temporal series*, principal component analysis, bands ratios, vegetation indices which convert multispectral information into a single index, etc. constitute another group of methods rather often employed.

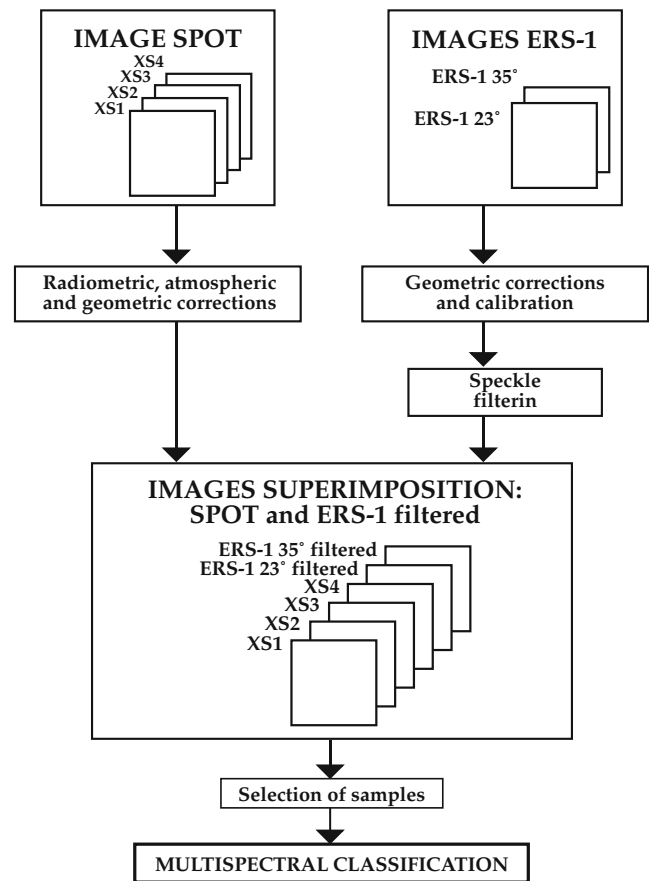
According to the aims of the study and the concerned area, each method has its own advantages and insufficiencies. To evaluate each method, a rigorous quantification of its accuracy would be necessary. However, such a test is very rarely carried out (Green et al. 1998).

Mangrove Features Discriminated from Space

The level of discrimination achieved with remote sensing products varies from one sensor to another and according to



Mangroves, Remote Sensing, Fig. 2 Classical approach for mangroves mapping using remote sensing data (supervised classification)



Mangroves, Remote Sensing, Fig. 3 Synergy between SPOT and ERS-1, general methodology

the processing method. As a general rule, when the amount of field data increases, the quality of the image processing improves. Almost all authors using aerial photos distinguish floristic classes labeled by genera and species. This level of discrimination which separates *Rhizophora apiculata* from *A. marina* stands is impossible at present from space. With spaceborne sensors, the easiest and primary objective is to separate mangroves from non mangrove vegetation (Biña et al. 1980). Some authors confess their inability to discriminate satisfactorily these two classes with SPOT XS (Green et al. 1998).

Several attempts have been carried out to distinguish stand density classes (Aschbacher et al. 1995) and height classes (Gao 1999). All these individual attempts are to be considered as preliminary research findings. From a practical point of view, it appears that seven most useful physiognomic classes can be currently detected and mapped, with or without GIS assistance, from space products:

Dense natural mangroves. This is the most important class, often located in protected areas. These stands are multi-specific and the ground coverage exceeds 80%.

Degraded mangroves where the ground coverage by trees and shrubs is about 50–80%. The spectral signal integrates the response of chlorophyllous elements and water-soaked soils.

Fragmented mangroves (or mosaics) where the remaining mangrove trees have a ground coverage about 25–50%. In such a case, the signal is primarily determined by the moist soils although the response of the remaining green vegetation is noticeable.

Leafless mangroves. As mangroves are almost always ever-green trees, a strong absorption in the NIR band (0.70–0.95 μm) has to be considered as totally abnormal, being induced either by a mass mortality of mangrove trees (Gambia, Côte d'Ivoire, etc.) or by an unexplained disease (cryptogamic, virus, insects, etc.).

Mangrove deforestation areas or clearfelled mangroves. Forest exploitation and clearfelling are dominant causes for the destruction of mangrove ecosystems (Saenger et al. 1983). Any opening in a mangrove canopy can be detected from space because corresponding pixels have been replaced either by water at high tide or by crusts of sodium chloride deposits during the dry season, at low tide.

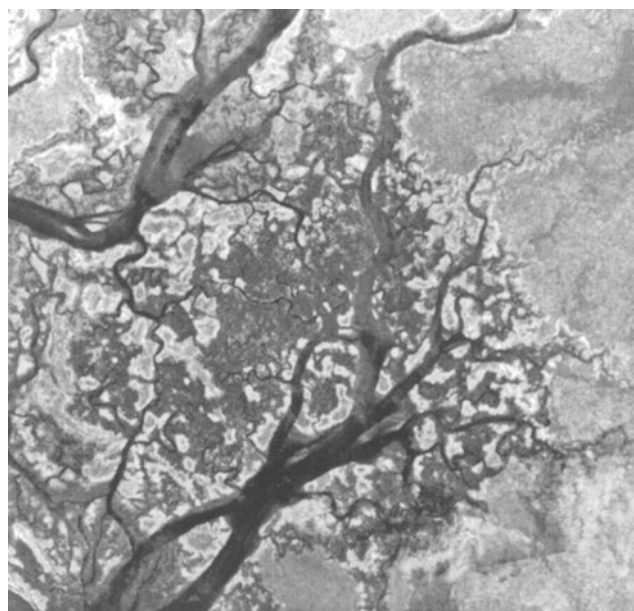
Mangrove converted to other uses. The most conspicuous human impacts on mangrove ecosystems is their conversion to shrimp ponds (Thailand, Ecuador, Viet Nam, Indonesia, etc.) or to agriculture (mainly paddy fields in Asia and West Africa). The signals of irrigated crops, mainly paddy fields and sugarcanes, being very distinct from that of mangroves (strong absorption in the NIR band), the delineation of these conversions has become routine work from space.

Restored mangroves and afforestation areas. Recently accreted inter-tidal zones bearing dense vegetation often correspond to mangrove restoration sites (Field 1996) or to afforestation activities. The monitoring of such areas has become effective from space at the mouths of the Ganges (Bangladesh) where the rate of survival and growth of *Sonneratia apetala* Buch-Ham is distinctly different from one island to another (Saenger and Siddiqi 1993; Blasco et al. 1997). Another interesting example is found in the Mekong delta (Viet Nam), near the capital city Ho-Chi-Minh, where *Rhizophora apetala* Bl. has been extensively replanted. In any case, dense monospecific planted stands have a high photosynthetic activity causing high absorption of photons and low response in the wavelength 0.6–0.7 μm , which make them rather easily discriminated from space.

Accuracy assessments carried out so far for each classification algorithm or for each discriminated class, lead to extremely variable results (Hudson and Ramm 1987). In a vast mangrove area like the Ganges (India and Bangladesh), the “dense mangrove class” which is the largest class, covering an area exceeding 6,000 km^2 , may present a classification performance of 90%, although some restored mangrove areas are often assigned either to fish ponds or to rice fields or to salt marshes or even to algal deposits, because the high physiognomic diversity of planted mangroves (age of the plantation, density, species diversity, substrate, etc.) induces very diverse spectral responses. For small areas, the existing possibilities to improve the accuracy of the maps and to increase the number of discriminated classes have been described by Ramsey et al. (1998).

Conclusion and Issues

According to the latest mangrove resource assessment (Aizpuru et al. 2000) carried out with space products, the total extent is about 170,000 km^2 and the world's mangrove regression during the last decade has been about 1,030 $\text{km}^2 \text{y}^{-1}$. A critical analysis shows that remote sensing utilization for mangrove studies is still limited for several technical reasons. It appears from what has been said that the enormous commitments made to promote remote sensing technology during the last 30 years have not yet given access to the data actually needed by modern ecological researchers, that is, identification of trees, structure of stands,



Mangroves, Remote Sensing, Fig. 4 Mangrove studies from space (part of SPOT data KJ 022/333 dated 21/10/93). The fragmentation of the habitat appears clearly. Dark patches correspond to mangrove types. ((c) cnes 1993-distribution Spot Image.)

physical or biological stresses, sediment load and geochemistry of brackish waters, etc.

Regarding the inventory and monitoring of mangrove ecosystems using satellite data, there is no worldwide standard method which could be applied straight away. At local levels, a periodic survey of mangroves from space is operational (Fig. 4). The permanent monitoring of these ecosystems, at local and regional levels, is premature.

Oil spills detection from space is by no means an easy task. Such pollution events are causing great damage every year in mangrove ecosystems of Nigeria, the Caribbean, the South China Sea, etc. either by oil tanker accidents, by illegal tanker cleaning, or by oil leaks from platforms or wells, etc. Recent advances have been achieved primarily because hydrocarbon compounds at the surface of the sea reduce water surface roughness which can be detected on SAR images.

The accuracy of each mangrove map and statistics is extremely difficult to assess rigorously. It is probably included between 50% and 90% in most studied cases. Combining data sets generated from field surveys, aerial photographs, especially color infrared, and high-resolution satellite products (TM or SPOT and SAR) lead to a more accurate appraisal of mangrove resource and local ecological conditions than does the analysis of high resolution satellite data alone. The merger of these sources of data produces classified maps, which include many features not separable solely by existing space data.

Finally, remote sensing specialists recognize that the technology has so far been less successful in coastal areas than in

continental zones. The reasons lie in questions of spatial and temporal scales and on the physics of coastal signals often distorted by marine aerosols and warped by the proximity of the ocean (Cracknell 1999). However, the synergy, between optical and SAR data and the new data provided by hyper-spectral scanning systems and by very high-ground resolution tools, leave little doubt that a new breed of remote sensing is emerging.

Cross-References

- ▶ Altimeter Surveys, Coastal Tides, and Shelf Circulation
- ▶ Geographic Information Systems (GIS)
- ▶ Global Positioning Systems (GPS)
- ▶ Mangroves, Ecology
- ▶ Mangroves, Geomorphology
- ▶ Monitoring Coastal Ecology
- ▶ Shoreline and Coastal Terrain Mapping
- ▶ Photogrammetry
- ▶ RADARSAT-2
- ▶ Remote Sensing of Coastal Environments
- ▶ Remote Sensing: Wetlands Classification
- ▶ Synthetic Aperture Radar Systems

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Marine Debris-Onshore, Offshore, and Seafloor Litter

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Definition

The source of marine litter is mainly land (c 80% and c 20% the sea) and the bulk is plastic, which ultimately ends up on

the sea bed (70%) in a variety of guises and size (mega to nano). It has a deleterious effect on humans, marine organisms, tourism, and coastal economics.

Introduction

Marine litter is any persistent, manufactured, or processed solid material discarded, disposed of, or abandoned in the marine and coastal environment. This may be from land, e.g., direct from beach users, rivers, or from the ocean itself, e.g., ships/offshore installations. Once in the marine environment, debris may remain for many years/centuries, particularly if it is plastic, and numerous world-wide studies have recorded plastic as the dominant material (Bergmann et al. 2015). The increased use of plastics and its global manufacture has led to ubiquitous and vast levels of abandonment; this, coupled with plastic's persistence and its propensity to degrade into smaller and smaller fragments has meant that it has accumulated in enormous quantities in the world's oceans. Degradation mainly takes place through photodegradation leading to surficial cracking followed by embrittlement and ultimately complete disintegration to powder. Performance is generally measured through changes in tensile strength and viscosity although UV and laser spectroscopy are other approaches. Plastics sinking to the deep seafloor are subject to photodegradation and if resistant to biodegradational processes will tend to be preserved. Enhanced photodegradable polyethylene degrades slower under marine conditions but expanded polystyrene foam deteriorates more rapidly in seawater than atmospheric exposure (Bergmann et al. 2015).

The plastics industry gives direct employment to over 1.5 million people in Europe, is ranked 7th in industrial value, has a turnover of more than 340 billion euros in 2015, and involves nearly 60,000 companies which contributed close to Euros27.5 billion to public finances and welfare in 2015 (www.plasticseurope 2015). A global population of >7 billion people produced over 320 million tons of plastic, which is estimated to double by 2034. The massive plastics industry has given rise to an issue of waste debris, one that affects beaches, coastlines, the pelagic zone, the sea surface through all depths to the seafloor, and its impact is of global significance. Marine debris is also often termed marine/beach litter, and these terms will be used interchangeably.

The problems created are now universally accepted as chronic and global, rather than acute and local as was often perceived in the twentieth century. It has been established as a serious pollutant for over 40 years but has only gained widespread recognition in the past two decades, with attention becoming far more focused within the last 10 years, especially with micro- and nanoplastic research. Reports of the "Great Pacific Garbage Patch," increasing public participation in cleanup schemes, awareness of the microplastics volume

and their impact on ecosystems, and focus on environmental issues, plus the social-media information transfer have led to greater public awareness. This has not led to any reduction in the amounts of debris being found on beaches or in the oceans.

Plastic Groupings

Definitions re grade size have proved to be a problem, especially when related to microlitter, an area of recent intense interest. For these, Andrady (2011) suggested three micro-divisions each having its own unique physical characteristics/biological constraints: mesoplastics (500 μm –5 mm), microplastics (50–500 μm), and nanoplastics (<50 μm). Primary microplastics input to the ocean has been estimated as 1.5 million tons per year (Mtons/year, with a range between 0.8 Mtons and 2.5 Mtons; Boucher and Friot 2017).

Sources

Marine debris sources can be broadly classified into two groups: seaborne sources and land-based sources. Until the early 2000s, most concern focused on debris discharged from vessels; however, there exists extensive evidence that some 80% is land-based. Of the global figures for marine debris, the bulk is deposited upon the seafloor (70%), with the remainder split between beaches and the water column.

Land-based debris can enter the sea through rivers or be blown, washed, or discharged directly from land. The absence of sewage treatment installations, the presence of combined sewer overflows, storm water discharges, runoff from landfills sited nearby rivers and in coastal areas, absence of waste services or landfills in rural areas, recreational beach users, and fly tipping all contribute to debris ending up on beaches/oceans. Identifying from which source debris has originated is a difficult task.

By comparison to land-based sources, the contribution of garbage from shipping and offshore installations may not be as large as previously thought, although it remains a concern (Galgani et al. 2015). It is also one of the few inputs of plastic and other debris which is directly controlled by international treaty. Annex V of MARPOL 73/78 covers garbage from ships and partly from offshore structures. It entered into force on 31 December 1988 and its aim is to eliminate and reduce the amount of rubbish being dumped into the sea from ships. Garbage includes all kinds of food, domestic, and operational waste generated during the normal operation of the vessel and governments are obliged to ensure port reception facilities to accept ship garbage. Annex V explicitly prohibits the disposal of plastics anywhere into the sea.

Extent of the Problem

Onshore

Marine debris has been encountered on beaches across the globe (Fig. 1). The historic focus for recording litter has been on beaches – some 15% of marine litter is found her – due to their relative easy access and the low-cost measurement methodologies and techniques and no coastal region exists that is not impacted or fouled by litter, e.g., Table 1, even remote islands. For example, Laversa and Bond (2017) at Henderson Island in the South Pacific found litter that had the highest density of debris reported anywhere in the world, with some 671.6 items/m² on beach surfaces. An estimated 37.7 million debris items weighing a total of 17.6 tons are currently present on Henderson, with up to 26.8 new items/m² accumulating daily.

The Offshore: Pelagic Marine Debris

Systematic investigations to establish quantities and distribution of pelagic plastic litter have been sporadic and are based on either surface-towed neuston (or pleuston) nets or have

used sighting surveys from vessels on passage. The former have focused primarily on mesolitter, mostly plastic pellets or nibs, and the latter on macro- and megalitter items identifiable with the naked eye from a vessel’s deck or bridge. Microlitter amounts are vast, globally dispersed and in certain locations constitute the most numerous litter type.

Nanolitter is derived from, e.g., packaging, propriety hand cleaners, and cosmetic preparations (in 2017, Oriel banned usage of these). Plastic mesolitter, mainly in the form of nibs or pellets of virgin polystyrene and polyethylene, has a universal presence in oceanic surface waters. The greatest densities of plastic litter, whether measured by weight or item count, are consistently found in coastal and shelf waters adjacent to and down drift from major urban and manufacturing regions. On the open ocean distant from land-based sources, it tends to concentrate along oceanic fronts and in large eddy systems or gyres, e.g., the Great Pacific Garbage Patch (Bergmann et al. 2015). Concentrations of macro- and megalitter are also present along many shipping routes particularly those of the North Atlantic and North Pacific. They are much less across the South Pacific where shipping traffic is sparser and industrial developments are fewer and distant.



Marine Debris-Onshore, Offshore, and Seafloor Litter, Fig. 1 Typical beach litter – N. Ireland, UK

The Seafloor: Benthic Marine Debris

The seafloor from intertidal and shallow sublittoral to outer shelf, slope, and abyssal depths globally has been identified as an important sink for marine debris, holding some 70% of marine litter. The problem is now appreciated to be global with many observations verified by divers, through video footage from ROVs as well as sampling by bottom trawls (Bergmann et al. 2015). Data has been obtained from varying depths and at many widely separated places. Latterly, there have been several studies presenting substantial data on types, amounts, and distribution of marine debris on the seafloor (Bergmann et al. 2015).

Quantities of sunken litter being reported are high with strong spatial variations. Moore and Allen’s (2000) shelf survey of the Southern California Bight ranked quantities of anthropogenic and natural debris into four broad categories (trace, low, moderate, high) on the basis of number and weight of items determined from standardized trawl times along isobaths between depths of 20 m and 200 m. Bathymetrically, the proportion of area with anthropogenic debris increased with increasing distance along a broad offshore front, from inner to outer shelf. This suggested a source that lies in disposal practices from boating activities. The most comprehensive and thorough reports are from European and western Mediterranean waters (Galgani et al. 2015).

“The abundance of plastic debris shows strong spatial variations, with mean densities ranging from 0 to more than 7700 items km^{−2}” (Galgani et al. 2015, 43). Near metropolitan areas, they could exceed 100,000 items/sq. km but elsewhere maximum values are lower (50,000 items/sq. km in the

Marine Debris-Onshore, Offshore, and Seafloor Litter, Table 1 International Beach Ocean Conservancy, Cleanup – main items found (OC 2016)

Item	Numbers
Cigarette stubs	2,127,565
Plastic beverage bottles	1,024,470
Food wrappers	888,589
Plastic caps	861,340
Straws	439,571
Plastic bags	424,934
Glass bottles	402,375
Metal caps	381,669

Bay of Biscay; 600 items/sq. km in the North Sea 200 km west of Denmark). Concentrations of debris (to densities >50,000 items/sq. km) have been encountered at depths of >2000 m on floors of canyons along the Mediterranean coast of France (Galgani et al. 2015). Variations in distribution patterns have been attributed to geomorphologic factors including little tidal flow, local anthropogenic activities, and coastal shipping. The blanketing effects of sheeting may damage biotas of both soft sediment and rocky hard ground substrates at all depths from intertidal to the abyss, leading to anoxia and hypoxia induced by inhibition of gas exchange between pore water and seawater.

Problems

Effects on Humans

Numerous studies of beach litter have commented on the potential danger to visitors, mainly from foot lacerations caused by stepping on glass or discarded ring pull tabs (Bergmann et al. 2015). Other, more dangerous items have been encountered on beaches, such as munitions and containers of corrosives have been found washed ashore along with pyrotechnics and packaged hazardous goods (Fig. 2). There are also less obvious health risks that can feature on beaches, i.e., medical waste and sewage-related debris. Although the risks are considered to be relatively low, any contact with infected sanitary products or fluids in syringes or other medical equipment or ingestion of any of these could cause disease.

In the twenty-first century, low concentrations (due to the hydrophobicity) of persistent organic pollutants (POPs) have been found in meso/micro plastics, which when ingested enter the food chain. Waste water from washing machines can

produce many thousands of microfibers and recently even tap water has been shown to be contaminated. From 14 countries, 159, 500 cc samples of tap water were analyzed, and 83% was found to have microfibers with diameters some 10 microns, which standard water treatment systems cannot filter out. These very small particles can harbor chemical/pathogens, which are small enough to penetrate cell walls (www.orbmedia.org). Foraging behaviors in anchovy schools have shown that they are attracted to odored (pollutant) plastics rather than clean plastics, which have an important bearing on fish as a staple food source (Savoca et al. 2017).

Economic Effects

The problem of litter leads not only to potential health risks but also economic losses (Mouat et al. 2010). Stranded debris has direct and indirect social and economic costs to shoreline communities, with the financial strains imposed by such debris not always easy to quantify or appreciate. The cost each year in removing beach litter for UK municipalities are approximately €18 million – a 37% increase over the past 10 years. In the Netherlands and Belgium, it is some €10.4 million per year (Mouat et al. 2010). Of Scottish fishing boats surveyed, 86% had experienced a restricted catch due to litter, 82% had their catch contaminated, and 95% had snagged nets on seabed debris. Fouled propellers/blocked intake pipes cost the Scottish fishing fleet between €11.7 million and €13 million on average each year, 5% of total revenue of affected fisheries (Mouat et al. 2010). Additionally, fouled fishing nets via plastic sanitary liners took 3 h to clean (Table 2).

Aesthetic Quality and Tourism

A clean and unspoiled coastline is the principal attraction for tourists, and marine litter can threaten the image, the reputation, and ultimately the economy, as the public may avoid certain beaches if they find the appearance unacceptable. For



Marine Debris-Onshore, Offshore, and Seafloor Litter, Fig. 2 Dangerous beach litter, UK

Marine Debris-Onshore, Offshore, and Seafloor Litter, Table 2 Number of plastic panty liners caught in 30 fishing nets at Swansea, UK

Net	Amt	Net	Amt	Net	Amt
1	7	11	7	21	3
2	6	12	0	22	4
3	3	13	5	23	2
4	2	14	3	23	4
5	8	15	13	25	4
6	3	16	3	26	10
7	5	17	7	27	44
8	19	18	2	28	8
9	2	19	4	29	4
10	8	20	2	30	3

most municipalities, the main incentive for removing beach litter is the potential economic impact on marine and coastal tourism as degraded coastal areas negatively affect tourism (McIlgorm et al. 2011). Many of these problems are manifest in developing regions where natural resources may be limited and economic development is largely dependent upon coastal tourism.

Biologic Interactions

The impacts of marine debris on wildlife are generally divided into two groups:

Entanglement: Large items of debris trap animals, which may result in the drowning of air-breathing species, asphyxiation of fish species that need constant movement to respire, or death by starvation or predation. Smaller debris items can become snagged on the seafloor trapping animals, or entangling birds and other animals on land, additionally, entangled objects can tighten around the animal leading to restrictions in growth. This can lead to death or inhibit the ability to reproduce and can also affect feeding (Bergmann et al. 2015).

Ingestion: Some species may be able to regurgitate or excrete debris, but some plastics do not appear to pass through the intestines of certain seabirds as there is a marked absence of debris from droppings. Rummel et al. (2016) analyzed 290 gastrointestinal tracts of demersal and pelagic fish species (herring and mackerel) from the North and Baltic and found plastic particles in 5.5% (74% being microplastics i.e., <5 mm) and 40% were polyethylene (PE) in 3.4% of demersal fish and 10.7% of pelagic fish. Similarly, Rochman et al. (2015) found anthropogenic debris (all plastic) in 55% of all market fish species from Makassar, Indonesia, and 66% from Californian markets (mainly fibers); these fish were sold for domestic consumption.

Summary

Marine litter, mainly land-sourced (80%), affects the environment, economy, human health, and wildlife, with some 70% ending up on the seafloor.

Cross-References

- Coastal Scenery
- Estuaries
- Marine Litter
- Rating Beaches
- Tourism and Coastal Development
- Tourism, Criteria for Coastal Sites

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Marine Litter

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Definition

Plastic litter is one of the biggest issues in the seas and oceans. Suggestions are given to combat this problem, i.e., innovation (making of cloths, chairs etc.), education, usage of economic instruments, beach cleans, recycling, and product design.

Introduction

Globally, marine debris, composed of all solid waste materials, litter (ML) including plastics together with microplastics (<5 mm in size), are ubiquitous in all oceans (Cheshire et al. 2009; Fig. 1).

At least 80% is land based, a result of poor waste management, recycling and recovery with plastics being the dominant component (50–80%; Bergmann et al. 2015). Surprisingly only 15% ends up on beaches; while the same amount remains afloat in the water column, the rest sinks to the seabed (OSPAR 2007; Galgani et al. 2000). Surveys however tend to be undertaken for specific locations (Table 1), but despite this beach litter globally tends to follow very similar patterns (Fig. 3) although amounts can vary considerably.

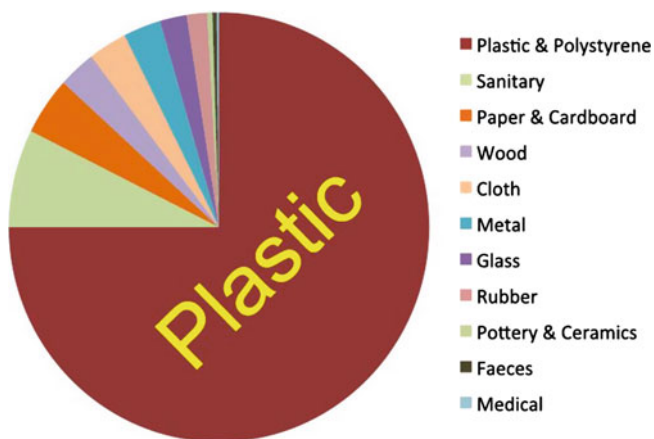
Over the past two decades, global litter studies showed that more of the modal plastic litter type occurs in the northern rather than southern hemisphere (Poeta et al. 2014). Estimates are of 275 million metric tons of plastics produced from 192 countries, with 4.8–12.7 million metric tons entering the oceans each year (Jambeck et al. 2015). ML is a social phenomenon, the visible evidence of rampant capitalism. ML is but a symptom of a problem and the problem is what to do with it, especially the plastic issue, as 60–90% of marine litter is made of different plastic polymers and most (40%) are disposables that have only been used once.

In 1950, the world's population of 2.5 billion produced 1.5 million tons of plastic; in 2016, a global population of >7 billion people produced over 320 million tons of plastic, which is estimated to double by 2034 (Fig. 2). One ton of plastics equals 20 k two litre drinks bottles or 120 k bags. Daily, approximately eight million pieces of plastic pollution enter the oceans. In the UK, approximately 5,000 items of marine plastic pollution occur per mile of beach; 150 are plastic bottles (www.sas.org.uk). As a result of an £750,000

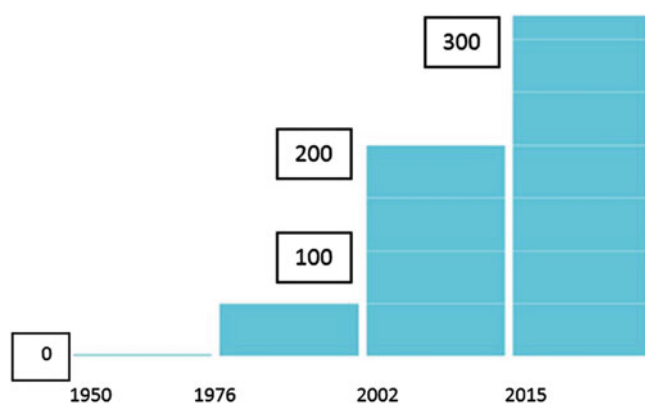
investment > UK 400 beaches will soon be checked weekly by the Environment Agency (EA) in order to tackle the “scourge” of discarded plastic. The new unit will log data from water sampling teams around England regarding the

Marine Litter, Table 1 Top 10 litter items found in recent surveys: Ocean Conservancy, 2016 (97 countries); Great British Beach Clean 2014 (GBBC, 364 beaches); Northern Ireland 2017 (14 beaches)

Item	Ocean Conservancy 2016 Rank	GBBC 2014 Rank	Northern Ireland Williams et al. 2017 Rank
Cigarette stubs	1	2	–
Plastic drink bottles	2	10	8
Food wrappers	3	3	6
Plastic caps/lids	4	4	5
Straws	5	–	–
Plastic bags	6	–	–
Glass	7	7	10
Metal cans	–	–	5
Plastic/polystyrene pieces	–	1	2 and 3
Plastic grocery bags	8	–	–
String and cord	–	5	1
Cotton bud sticks	–	6	9
Rope	–	–	7
Drinks	–	–	4
Wet wipes	–	8	–
Fishing lines	–	9	–
Metal bottle caps	9	–	–
Plastic lids	10	–	–



Marine Litter, Fig. 1 Marine litter composition (OSPAR 2007)



Marine Litter, Fig. 2 Global annual plastic production, in million tonnes (www.Statista.com, PlasticsEurope)

amount and type of plastic found. It hopes to indicate the worst-hit beaches and use its findings to better regulate the problem. Most plastic bottles found on beaches are used for soft drinks and water and made from polyethylene terephthalate (PET), which is highly recyclable, but collection and recycling is poor. Global PET plastic bottle production has risen: 2004, 300 bn. units; 2014, 470bn; 2016, 490bn.; and 2021 (estimate), 500bn (www.Euromonitor.com). The expected bio-plastics growth for the period 2015–2022 is 15.4%, with an 80% global share of the food and drink market.

Plastics are very versatile and can be tailored to meet specific technical needs, as they are/have:

- Lighter in weight than competing materials, reducing fuel consumption during transportation
- Extremely durable and relatively inexpensive to produce and are resistant to chemicals, water and impact
- Good safety/hygiene properties for food packaging
- Excellent thermal/electrical insulation properties
- Mobility: travel huge distances around the world with ocean currents and winds; found across the entire water column

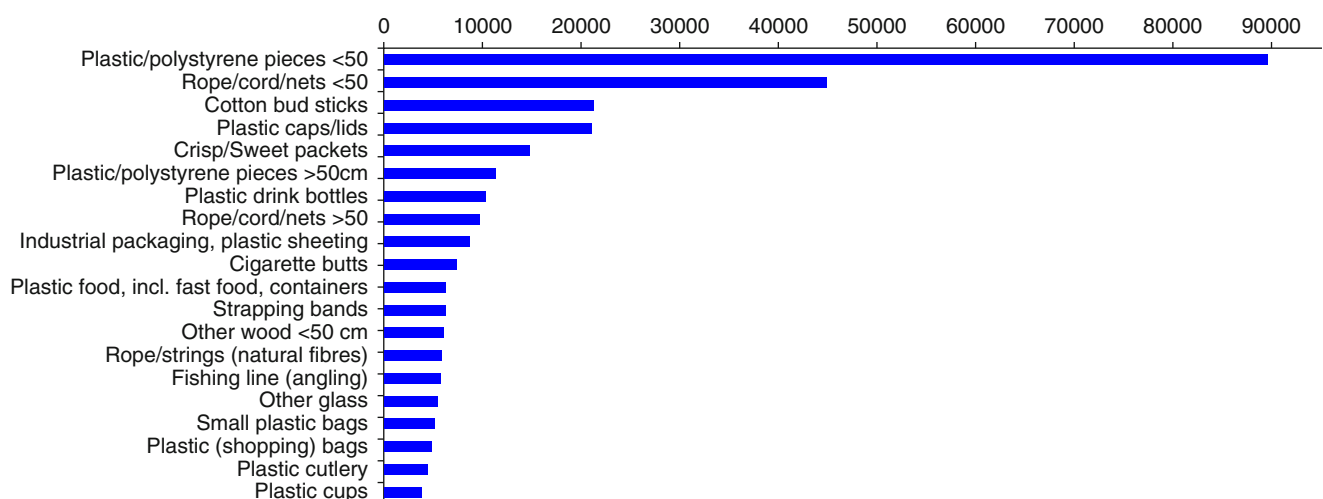
ML is multisectoral problem plus a trans-boundary one. It is probably the most “exposed” environmental problem to the public and the millions of people who are economically affected. No other marine pollution phenomena mobilises such public participation and readiness to act. ML is not an environmental problem to be solved ONLY by means of legislation, law enforcement, beach cleaning campaigns, and technical solutions. It is also a cultural problem, and sterling efforts to change attitudes, behavior, management approaches, education, and involvement of all sectors and interests need to be undertaken.

What We Know

- Legislation alone appears ineffective in reducing marine litter.
- Despite a lack of research, there is much evidence that the economic impact of ML is significant.
- Apart from a few organizations, e.g., Marine Conservation Society, KIMO, Ocean Conservancy, who specialize in beach litter; there seems little co-ordination of effort to tackle ML (Fig. 3).
- A significant amount of beach litter items are small (<30 cm).
- 50–90% of beach litter comprises plastics, which persists in the sea for a long time.
- Much SRD still originates from combined sewer overflows (CSOs), which has a huge negative impact on public perception of water quality and fitness for bathing and other uses.
- ML has led to many volunteer-based surveys and clean-up schemes, which are cost-effective for gathering data. Estimated potential aggregated costs from the consequence of adverse ML effects on marine and maritime industries are >£54 m; the costs of clean-up and prevention costs are >£34 m per annum. Clean-up costs of ports, harbors, and beaches and marinas are the major costs to the maritime industry in the UK (Lee 2015). It is difficult to estimate the value of the marine and wildlife which can result in direct harm and death.
- Large volumes of litter can accumulate in shoreline and seabed sinks (Galgani et al. 2000).

What We Do Not Know/Have

How ML affects populations or indeed whole *ecosystems*. Plastic litter either as a whole manufactured product, such as



Marine Litter, Fig. 3 Most common (total numbers) items on reference beaches in the NE Atlantic (Lozano and Mouat 2009)

cartons, bottles, or as fragments/pellets, comes with a high socio-economic cost (the later in particular), constitutes a huge threat to biota (Mouat et al. 2010). Additionally, plastic polymers contain additives to give the plastic marketable qualities. These additives are able to leach out of the polymer matrix into the environment, e.g., DEHP a phthalate “plasticizers” used in PVC manufacture, PBDE used in flame retardants.

- Assessing economic damages associated with litter impact. UK municipalities spend approximately €18 million each year removing beach litter, while there is approximately a €10.4 million per year cost to municipalities in the Netherlands and Belgium (Lorenzo and Mouat 2009). The main motivation and incentive for beach litter removal for most municipalities was the potential economic impact on marine/coastal tourism, as degraded coastal areas negatively affect tourism.
- The *long-term* effects of persistent breakdown particles. Is it 100 or 1000 years?
- Enough monitored data to *source* litter items accurately.
- How *effective* is MARPOL, since there is no monitoring system to measure any effects of legislation.

For oceans, a litter breakdown is shown in Fig. 4. Many academic papers have been written on plastics found, but the thrust of twenty-first-century research has been on microplastics. The global release of primary microplastics into the ocean was estimated at 1.5 million tons per year (Mtons/year) with a range between 0.8 and 2.5 Mtons/year according to an optimistic/pessimistic scenario (Boucher and Friot 2017).

Bergmann et al. (2015) showed that microscopic plastic fragments (<5 mm in diameter) and fibers are ubiquitous marine pollutants. Plastic never decompose fully, but breaks down into tiny microplastics pieces. Much large and small marine life unknowingly ingest these microplastics and when eaten as seafood, humans end up consuming the plastics; as many as 51 trillion microplastic particles litter the oceans, seriously threatening marine wildlife, e.g., vulnerable bivalves, i.e., mussels (*Mytilus edulis*) (www.oceanconference.un.org, 2017). Washing machines release 100,000 s of microfibers each time they are used. Chemical transfer from plastics to organisms is a possibility, as plastics can absorb toxic chemicals, e.g., PCBs (polychlorinated biphenyls) and DDE (dichlorodiphenyldichloroethylene) obtained from pesticides, which are known to be endocrine disruptors. Their interaction is with entire ocean ecosystems. The growing presence of microplastics in the oceans requires new thinking about how to mitigate both primary and secondary releases across the supply chain.

To Curb the Inexorable Rise in ML: Some Suggestions!

Solutions exist but there is no, “*one size fits all*”

1. **Reduce plastic input and change to degradable plastics.** For example, bottles made from polylactic acid (PLA) but there is a problem re degradation rates:
 - Produce 60% less greenhouse gases.
 - Uses 50% less fossil fuels in its production.
 - 100% of the bottle’s material comes from plants like corn, cassava, and sugar beet, so it is fully biodegradable.
2. **Use Economic Instruments** (Table 2).
3. **Continue to clean beaches**, e.g., Ocean Conservancy, Ca (global), Marine Conservation Society (UK). By and large, this is preaching to the converted, but it all helps to gather much needed data and hopefully gets a message across to other members of the public.
4. **Have an appropriate methodology** to classify ML. The Ocean Conservancy, Ca, USA, OSPAR have suggested guidelines, but a plethora of methods exist that can be seen in countless academic papers. An exhaustive account of this issue can be found in Cheshire and Adler (2009).
 - Beach survey from water’s edge to splash zone
 - Varying width transects to find the optimum
 - Transect line quadrats
 - Strand line counts
 - Postal surveys
 - Offshore water column
 - Number of plastic bin bags/trucks, etc. (many global beach cleans)
5. **Enhance the circular economy** (Fig. 5), which can “close the loop” of product lifecycles (more recycling/reuse) to obtain maximum value and use from all materials, products and waste. The plastic flow must be redirected and cleaned up in order to stop it ending up in the oceans. This needs a circular economy:
 - The “R” Message: Rethinking products for Reuse, Refurbish, Remanufacture, and Recycling, the aim being on reusing refurbishing, etc.
 - Biomimicry, i.e., learning from nature and designing systems analogous to natural algorithms.
 - Selling performance instead of products and changing from the old linear economic model that has existed since the eighteenth century. With declining GDP growth trends, pressure for higher consumption has been increasing, promoting under-utilization of value of goods bought and all the resources embodied in those goods.

This would hopefully eliminate the concept of waste, as systems collect and recover the value of these materials, maximize renewable energy sources, and celebrate

diversity in all its meanings. The current economy is linear, where we *take, make, use, and dispose*. A “closing of this loop” is needed by *taking, making, using, and recycling* and not simply taking anymore. After a first-use cycle, 95% of plastic packaging material

(US\$80–120 bn. /annum) is estimated to be lost to the economy (www.ellenmacarthurfoundation.org). Working with finite resources and renewable energy, it is essential to understand the role of primary/secondary resources. A recycling target of 75% of all waste and

Marine Litter, Table 2 Suggested economic instruments

Market-based instruments	Explanation
Effluent charges	Payment of fees to pollute
Fines	For littering. These exist but are not often enforced. This is a MUST
Deposit-refund system	Return an item, get deposit back. Iceland has this for cans/bottles. The Scottish Government has just (2017) commissioned detailed work on how a potential “deposit return scheme” might operate in Scotland
User/admin charge	Product disposal and municipal charges
Auction/tender	Of the right to create pollution
Tax incentives	Tax rebate for firms etc. for collection/litter prevention
Sales tax	Specific items, e.g., plastic bags. Six billion fewer plastic bags were taken home by shoppers in England and over £29 million donated to good causes thanks to a 5p charge. It reduced plastic bags found on English beaches by half. One trillion plastic bags are produced each year, (two million every minute), each bag with an average use time of just 12 min
Tradeable permits	Right to trade with set pollution outputs
Tourist taxes	“Voluntary/ Mandatory” or imposed by Authorities, e.g., Great Barrier Reef visits. Many USA States have these automatically added to tourist bills, e.g., Florida.
Award incentives	Encourages local communities, e.g., Green Flag award in Wales, UK.
Port reception facilities	Low/no cost, which encourages disposal.

landfill maximum of 10% of all European waste by 2030 was set in 2015 by the European Commission. Plastics are now recycled and turned into clothes, furniture, etc.

6. **Regional Cooperation**, especially with Fisheries, would cover: damage/direct loss of fishing boats, loss of production via ghost nets, loss of fishing gear, and down time (US\$10 m. only for retrieval in Canada), propeller entanglement (\$US792 m. in the USA), human injuries (US\$440,000 in the UK), loss of amenity and aesthetics to beaches, and assessment of world wildlife. Debris-related damage to marine industries has been estimated at US\$1.26 bn./annum in 2008 for the 21 countries in the Asia Pacific rim (Mellgorm et al. 2008).
7. **Education** in ALL sectors. This would include industry, public, seafarers, schools, and universities. Cutting litter off at source should be a priority and it can be done. People’s attitudes can be changed, e.g., the successful “do not drink and drive” campaign in the UK. “The US

Plastic Pollution Coalition” (www.plasticpollutioncoalition.org/) with >500 members; 90 NGOs under the Surfrider umbrella (www.surfrider.eu/en/) are organizations focused on aiming to stop plastic pollution, both aiming to empower more people and organizations to take action to stop plastic pollution and live in a plastic-free environment.

8. **Product design**. Design for the product life and end-of-life and label accordingly. Why does one need colored plastic soft drink bottles? Pigmentation is purely brand convenience and marketing and frequently coloring makes plastic recycling void. Orange, blue, etc., on a PET bottle vastly reduces its recycling potential. The business model is outdated and inefficient and product redesign will produce synergistic benefits, which enhances society in all aspects and cuts waste. With coffee cups, the problem is the plastic lamination lining that makes cups impermeable. Some five billion are in circulation; the paper can be recycled, but the polyethylene lining is resistant to degradation, at least 30 years to break down if a reusable cup is used usually made by the particular chain. Other reusable cups (and prices) are Rice Way £18; SAS bamboo £12 and Keep Cup £10. Starbucks currently gives a 25p discount if their re-usable cups are used; Costa gives a 25p donation to charities. Currently FRUGALPAC (www.Frugalpac.com) is close to developing a 100% one.
9. **Thinking “outside the box”**. Estimates to clean up the Pacific garbage patch are of the order of 7.25 million years (www.boyanslat.com). Slat argues that via 24 platforms (manta ray shapes) about 40% of the plastic in the North Pacific Gyre can be harvested within 10 years. Multiple long drifting boom collectors, 1 km in length, cut the per-unit cost to a few million dollars and by 2018 trials should be ready. It is estimated that the moored design would collect about 40% of the plastic in the North Pacific Gyre within 10 years. The approach uses weighted 12 m. tall metal towers suspended for a few hundred meters below the surface, where currents are about one-fifth the surface speed. Weights drag in these slower currents, while faster moving surface water will move plastic debris, which will collect on drifting booms.
10. **Research**. Research of all kinds is needed from plastic chemistry, manufacturing process design, correct governance, sourcing, methods, etc. Waste dynamic is little understood and this is necessary to monitor mitigation efficiency. What are the transport rates between biota and the environment and what if the environment changes? A whole range of questions can be asked but answers are only coming slowly.

Marine Parks

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There is a wide and growing range of marine parks. They include designated natural marine areas, such as the coral reefs off the coast of Al Fujayrah in Abu Dhabi, which were declared the country's first marine parks in 1995, the Great Barrier Reef, and artificial tourist facilities, such as Seaworld California in San Diego and Marineland Canada near Niagara Falls. The latter often include captive marine life, including marine mammals, as well as recreational features and structures such as water-slides. They attract large numbers of visitors and for many people are their first introduction to the marine world. They are, however, controversial because of their use of captive mammals, such as dolphins, although concern for animal welfare means that the best operate with very high standards of care and carry out important research. Some natural marine parks use underwater observatories to provide dry access to the undersea world, for example, in the Marine Park and Underwater Observatory at Coral World on the northeast coast of St. Thomas in the US Virgin Islands. The majority, however, depend upon boat and diving access to the marine world, for example, in the Pulau Weh Marine Park in western Indonesia and the Malindi Marine Park on the coast of Kenya.

The terminology used to describe Marine Parks varies from country to country depending upon the specific legal designation used, and the term "Marine Park Area" is also commonly used. Many marine parks also form parts of a worldwide pattern of Marine Protected Areas (MPAs), which include a wide range of legislative and conservation practice. However, not all marine parks are marine protected areas and not all marine protected areas form parts of marine parks.

Historical Background

Although the First World Conference on National Parks held in 1962 expected governments of coastal countries to address the creation of marine parks or reserves to protect underwater areas as urgent, the response was very mixed, and in many countries very little action occurred.

Ray (1976) suggested that marine areas could be reserved to achieve several objectives:

1. Conservation of habitat
2. Protection of species habitat
3. Conservation of important breeding areas
4. Conservation of aesthetic values
5. Protection of cultural and archeological areas
6. Provision of sites for interpretation, education, tourism, and recreation
7. Provision of areas for research
8. Provision of areas for monitoring the effects of human activities, and
9. Acting as areas in which to train personnel in protected area management.

In 1988, the International Union for the Conservation of Nature (IUCN) defined a MPA as "any area of intertidal or sub tidal terrain, together with its overlying water and associated flora, fauna, historical and cultural features, which has been reserved by law or other effective means to protect part or all of the enclosed environment." The IUCN Commission on National Parks and Protected Areas (1994) recognized that protected areas were managed for six main reasons:

1. Strict protection (i.e., strict nature reserve or wilderness area, e.g., Laut Banda, Indonesia);
2. Ecosystem conservation and recreation (i.e., National Park e.g., Ras Mohammed National Park, Egypt);
3. Conservation of natural features (i.e., Natural Monument);
4. Conservation through active management (e.g., habitat or species management area, e.g., Galapagos marine reserve, Ecuador);
5. Landscape or seascape conservation and recreation (protected landscape or seascape, e.g., Northern Sporades, Greece);
6. Sustainable use of natural resources (managed resource protected area, e.g., Kiunga Marine National Reserve, Kenya).

IUCN proposed that marine parks should be large (more than 1,000 ha), and have both high marine conservation value and high recreational potential. Educational and recreational use in these areas should be controlled at a level which ensured that the area's natural or near natural state is maintained. In contrast, marine reserves should be areas which have some outstanding ecosystem feature and/or species of flora or fauna of natural or scientific importance or they should be representative of a particular marine area. They may contain fragile ecosystems or life forms, be areas of biological or geological diversity, or be of particular importance for the conservation of genetic resources. These reserves should be large enough to ensure the integrity of the area, to protect communities and species, and to maintain natural processes. However, IUCN also expects that they can accommodate a wide range of human activities which are compatible with their primary goal in marine and estuarine settings (Box 1).

When MPAs are selected, their special importance is usually because (Norse 1993) they are:

1. areas of high diversity;
2. areas of high endemism;
3. areas of high productivity;
4. spawning areas that serve as a source of recruits;
5. nursery grounds; and
6. migration stopover points and bottlenecks.

Thus, although there is strong support for the co-establishment of marine parks and MPAs, the reasons for the designation of MPAs can be in conflict with the expectation that marine parks are for recreation as well as conservation. Furthermore, as Gibson and Warren (1995) stress, legislative requirements for the creation of MPAs depend upon a combination of national and international law. For example, the Law of the Sea Convention (articles 192 and 194) imposes an “obligation on States to protect and preserve the marine environment and requires them to take action to prevent, reduce or control pollution from any source.” In addition, there are obligations under the International Convention for the Prevention of Pollution from Ships 1973 (MAR-POL) and the Ramsar Convention (1971), and through the International Maritime Organisation and the UNEP Regional Seas Programme to identify areas which are sensitive and should be avoided by shipping. It is not surprising, therefore, that there is so much variety amongst the world’s marine parks.

Box 1 IUCN recommends that a national system of marine protected areas should have the following objectives

- To protect and manage substantial examples of marine and estuarine systems to ensure their long-term viability and to maintain genetic diversity
- To protect depleted, threatened, rare or endangered species and populations and, in particular, to preserve habitats considered critical for the survival of such species
- To protect and manage areas of significance to the life cycles of economically important species
- To prevent outside activities from detrimentally affecting the MPAs
- To provide for the continued welfare of people affected by the creation of MPAs
- To preserve, protect, and manage historical and cultural sites and natural aesthetic values of marine and estuarine areas, for present and future generations
- To facilitate the interpretation of marine and estuarine systems for the purposes of conservation, education, and tourism
- To accommodate within appropriate management regimes a broad spectrum of human activities compatible with the primary goal in marine and estuarine settings

(continued)

Box 1 IUCN recommends that a national system of marine protected areas should have the following objectives (continued)

- To provide for research and training, and for monitoring the environmental effects of human activities, including the direct and indirect effects of development and adjacent land-use practices.

Based on Kelleher and Kenchington (1992, p. 9)

Marine Parks in Practice

Practice varies from country to country, but many of the fundamental principles for marine park conservation and management were developed in the Great Barrier Reef of Australia. Different approaches were adopted by Japan, but both countries provide good examples of approaches to marine park establishment.

Australia’s Great Barrier Reef Marine Park Act 1975 is one of the earliest examples of applying the conservation philosophy of the World Conservation Strategy. The Great Barrier Reef is the world’s largest system of corals and associated life forms. It extends along the north-east coast of Australia for almost 2,000 km from the Tropic of Capricorn to about 9°S. It is a complex maze of 2,500 individual reefs ranging in size from under 1 ha to over 100 sq. km. The Reef Region exceeds 350,000 sq. km (135,000 sq. miles) – an area slightly larger than Poland. In the south, the Reef is wider than 100 km and is characterized by patch reefs separated by narrow winding channels or open water. To the north, the Reef is much narrower, occasionally less than 30 km, with a series of linear ribbon reefs at its eastern edge.

The Great Barrier Reef has a long history of use for fishing, with Australian aboriginal fishermen using it 15,000 years ago. With the arrival of Europeans, it became widely used for commercial fishing, including pelagic and demersal fish, prawns, scallops, turtles, and beche de mer. In addition, it is a prime location for scientific research and for a wide range of recreational activities. Between 1960 and 1983 the number of trawlers fishing within the Reef region grew from 250 to 1,400 with a turnover of about \$(A) 40 million in 1983 (Kelleher 1985). Tourism is a very important economic activity throughout the region with more than two million visitor trips to the Reef, island, and adjacent mainland in the 12 months ending in March 1980 (Kelleher 1985). By 1997, the annual value of tourism was estimated to exceed \$(A)1 billion and the direct economic value of commercial fisheries at about \$(A)200 million. There are an estimated 24,300 privately registered boats involved in recreational fishing as well. In 1997, the Great Barrier Reef Marine Park Authority (GBRMPA) recorded 1.6 million

visitor-days. Accessibility has improved greatly as high-speed catamarans with speeds between 25 and 35 knots have brought the Reef within 1 h travel time from mainland towns.

The Great Barrier Marine Park Act 1975 (Section 7(1)) established the GBRMPA with the following functions (Kelleher 1985):

- (a) To make recommendations to the Minister in relation to the care and development of the Marine Park including recommendations from time to time, as to
 - (i) the areas that should be declared to be parts of the Marine Park and
 - (ii) the regulations that should be made under this Act
- (b) To carry out, by itself or in co-operation with other institutions or persons, and to arrange for any other institutions or persons to carry out, research and investigations relevant to the Marine Park;
- (c) To prepare zoning plans for the Marine Park in accordance with Part V;
- (d) Such functions relating to the Marine Park as are provided for by the regulations;
- (e) To do anything incidental or conducive to the performance of any of the foregoing functions.

The Park is divided into six Sections, each with a zoning plan, which aims to allow any reasonable activity, but is designed to ensure that

1. the natural qualities of the Park are conserved for users, both today and in the future;
2. suitable areas are identified where reasonable uses are permitted;
3. incompatible activities are separated; and
4. these activities do not cause unacceptable damage (Kelleher 1985).

GBRMPA followed five phases in establishing zoning plans for each Section of the park (Kelleher 1985):

1. All the available information on the section in industry and government reports and the scientific literature and maps was collected, collated, and studied.
2. A public participation program, using both specially prepared information as well as extensive use of the media, advised of the intention to prepare a zoning plan and sought information from the public.
3. A draft zoning plan was prepared using both the information review and the information provided by the public.
4. A second public participation program using the same approaches was used to gather comments on and reactions to the draft zoning plan.
5. After taking all comments received into account the final zoning plan was submitted to the Minister to seek approval

before laying the plan before Parliament. After 20 days without any motions against the plan in either House of the Australian Parliament, the Minister published the date upon which the zoning plan would come into force. The first zoning plan, for the Capricornia Section covering over 11,800 sq. km (4,558 sq. miles), went into force in July 1981.

The GBRMPA thus came into being after extensive public consultation and was firmly established within the legal structures of Australia. Marine Parks need to have both the support of the wide variety of stakeholders in them and it is essential that appropriate legal arrangements are in place to maintain their protected status. However, the legal processes used may not be consistent with the customary practices of indigenous peoples who live and use them. Their interests must be safeguarded, not least because they are often custodians of information about the ecology and heritage of these sites.

Monitoring and research are very important for both the establishment of Marine Parks and their sustainable management. The GBRMPA identified three information and research needs (Kelleher 1985):

1. Information management, to ensure that information is available for analysis and interpretation.
2. Analysis of use to understand how the area is used, the effects of these uses on physical, biological, and economic processes, and their intensity, patterns, and rates of change. Changes which may be due to external factors such as sea-level rise or climate change and effects which are related to the uses themselves can be identified.
3. Resource analysis building up an inventory of the physical, chemical, human, and biological processes.

This approach to resource analysis, use analysis and information management has provided a model for the approach to marine parks. It acknowledges that marine parks are valued economic and social resources where the use of the resources must be monitored, managed, and if necessary changed to ensure the longer-term survival of the park's habitats and economy. The process of marine park management is a continuous one and in July 1998 the GBRMPA was restructured in order to strengthen its focus upon four major critical issues:

1. fisheries;
2. tourism and recreation;
3. water quality and coastal development; and
4. conservation, biodiversity, and World Heritage.

These four themes have increasingly become concerns for many marine parks and protected areas, for sometimes they are victims of their own success. As they attract visitors, so

the pressures both within the parks and along adjacent coasts have grown. At the same time, there has been a growing recognition that management of the water quality of these areas depends upon management of the catchments that drain toward them. The emphasis has shifted toward more holistic management of the coastal zone within which the marine parks are sited.

Whereas, the GBRMPA approach applies to a specific very large feature, the Japanese approach is designed as a framework for the development of marine parks, many of which are small in comparison, throughout Japan. In 1931, Japan passed a National Parks Law to allow the creation of parks for protection of scenery and for recreation (Marsh 1985). Between 1934 and 1936, 12 areas were designated, including some coastal sites, the largest being the Inland Sea (Seto Naikai). The Natural Parks Law of 1957 defined the purpose of parks as

the protection of places of scenic beauty, and also, through the promoted use thereof, as a contribution to the health, recreation and culture of the people.

The Law provided for three types of parks: National Park, Quasi-National Parks, and Prefectural Nature Parks. By the mid-1980s, Japan had 27 National Parks, 54 Quasi-National Parks, and 297 Prefecture Nature Parks, many of which were coastal in location. Following the First World National Parks Conference in 1962 which drew international attention to marine parks, the Nature Conservation Society of Japan (NCSJ) set up a marine parks investigation committee in 1964. In 1966, the Ministry of Health and Welfare financed the investigation and in 1967, the NCSJ committee became an independent foundation: the Marine Parks Center of Japan. In 1970, the Natural Parks Law was amended to allow the creation of marine parks within national parks. Ten were designated immediately and a further 12 in 1971. By 1985, there were 23 Marine Parks adjacent to 10 National and 13 Quasi-National Parks. They included 57 Marine Park Areas (special zones within marine parks) totaling 2,387 ha and ranging in size from 3.6 ha to over 233 ha. Marine Parks aim at “preserving beautiful underseascapes.” Criteria for selection required (Marsh 1985) that:

1. both land and sea areas surrounding the Marine Park Area are designated as National or Quasi-National Park and that nature conservation on land can be fully ensured;
2. the seabed topography is typical of the area and undersea fauna and flora abundant;
3. seawater should be transparent and unlikely to become turbid or polluted;
4. water depths are not more than 20 m;
5. there are slow currents and the park is sheltered from violent waves;

6. on land, there should be enough space to construct such facilities as landing piers, rest houses, visitor centers, and car parks;
7. there should be coordination with local fisheries concerning their use of the Marine Park, and;
8. the slightest risk of destruction of under-seascapes by all kinds of industrial exploitation should be prevented.

A marine park is defined as the total area comprising three zones:

1. Ordinary Area – including that part of the park within 1 km of the coastline.
2. MPA: the areas within the park having specific features warranting the most protection.
3. Ordinary or buffer zone including the area within 1 km of the MPA.

In practice, most Japanese marine parks are divided into two zones: the Ordinary Area zone including the sea within 1 km of the MPA or the coastline and the MPA core zone.

For some countries, the criteria are simpler, but explicit. For example, the 38 Malaysian marine parks were established in order to protect and conserve (1) the marine ecosystem, especially the coral reefs, for the management of the fisheries resources in the coastal waters in order to maintain/increase fish landings and (2) the coral reefs for research on biodiversity, and for the purposes of education and recreation/ecotourism. There is thus a very strong emphasis on the economic importance of these areas for fisheries and tourism.

In contrast, in some regions where mass-tourism was established early and has ceased to grow rapidly, marine protected areas may be an effective way to restore damaged environments. For example, after tourism to Spain slowed down in the 1990s, it decided to establish a protected area every 30 km along the coast to ensure the maintenance of marine and terrestrial fauna and flora. Many sites that have been partly damaged could be restored, in particular around the main areas of interest for tourism (Gubbay 1995). Spain has created some 25 protected areas on the Mediterranean coastline, 6 related to the coast or sea. MPAs come under the jurisdiction of the Ministry of Agriculture, Fisheries and Food and include a wide range of sites both in reasons for designation and in size (Table 1).

In contrast, Finland has designated MPAs within the context of the needs of a Regional Sea: the Baltic. Finland is a signatory to the Helsinki Convention that requires its members to protect the marine environment and biodiversity (HELCOM 1996). Both marine and coastal biotopes are included within the framework of HELCOM initiatives for the protection of species, habitats, and ecological processes including the establishment of coastal and marine Baltic Sea

Marine Parks, Table 1 Spanish MPAs and other Spanish-protected areas of international

MPAs (Mediterranean coast)	Biosphere Reserves	World Heritage Site
Medas Islands Marine Reserve (40 ha) S'Arenal Regional Protected Landscape (400 ha) Tabarca Marine Reserve (1463 ha) Columbretes Natural Park and Marine Reserve (5,766 ha) Archipelago de Cabrera National Park (Balearic Islands) (10,000 ha) Cabo de Gata Nature Park and Marine Reserve (26,000 ha)	Donaña National Park Island of Minorca (including protection of the sea adjacent to protected core areas)	Salinas de Ibiza y Formentera Nature Reserve (protecting oceanic <i>Posidonia</i> sea grass beds)

Source: World Conservation Monitoring Centre (WCMC)

Protected Areas (BSPAs). This is unusual in that it recognizes that MPAs are often natural units which cross national marine boundaries and that cooperation within Regional Seas is essential. Finland has eight proposed BSPAs, most of which are already protected as National Parks. For example, the Archipelago National Park, established in 1983, forms the core of a larger Archipelago Sea Biosphere Reserve established by UNESCO in 1994. It consists of around 1,000 islands and islets with important bird areas forming clusters separated by large expanses of open water. Although it is open to the public, there are restrictions, including prohibition of landing, especially during the nesting season.

Italy has designated more than 70 marine and coastal protected areas, including Marine Natural Reserve (MNR), Nature Reserve (NR), Special Protection Area (SPA), Fisheries Reserve, Biogenic Reserves, Recreation Parks, State Reserves, Marine Reserves and Parks, and Marine Sanctuaries. The designations take into account such treaties as the Mediterranean Action Plan and the Ramsar Convention (WCMC 2001). As well as subscribing to internationally recognized initiatives, due to its position sharing borders and waters with seven states, Italy importance has agreements with them for protection of the marine and coastal environment. For example, an international marine reserve was established between Sardinia and Liguria (Italy), Corsica (France), and Provence (France and Monaco) in 1993. There is thus considerable variety in the approach taken within Europe as areas that have often been exploited for centuries are given protection and restoration is undertaken.

In North America, both the United States and Canada have identified a wide range of areas which should be designated as MPAs, including marine parks. MPAs are frequently protected primarily to conserve important biological or historical

features, but many are also marine parks although they are not so designated because they allow access for recreational activities. In Michigan, for example, the Great Lakes Bottomland Preserves are designated in order to protect shipwrecks of historical importance. The Michigan Department of Natural Resources Underwater Salvage Committee has emphasized that, whereas underwater parks would provide physical facilities for recreational divers such as slipway access and interpretation programs, bottomland preserves are intended to be set-aside areas where little or no salvage will be allowed and they are recognized as areas of distinctive historic and/or recreational interest.

The United States has a very large number of MPAs, but only one area specifically designated as a Marine Park (Julia Pfeiffer Burns MP in California) and there are 13 State Marine Parks (2 in Washington and 11 in Alaska). There is a single Aquatic Park at Estero Bay in Florida. This is not uncommon. South Africa, for example, has only two Marine Reserves and the United Kingdom has no Marine Parks, although it has a very large number of MPAs.

Canada has a long-established system of national parks, the first being established in 1885 in part of what is now the Banff National Park. However, its marine national park system only came into effect in 1986 with the National Marine Parks Policy. In contrast to many countries, Canada has based its marine parks system on the fundamental principle that it should protect the country's major biogeographical provinces, specifically the Arctic, Atlantic, and Pacific marine environments and the Great Lakes, by sampling their representative, outstanding, and unique characteristics. This is based upon a broad-scale hierarchical system of biogeographic units or "marine regions" (Mondor 1992). In establishing the marine parks, the emphasis has been upon ensuring that there is a steady movement toward each marine region being represented: for example, the St. Lawrence River Estuary marine region is represented by the Saguenay Marine Park. For countries that are working toward establishing marine parks, this well-trying approach is recommended by IUCN, but for many less-developed countries, there is neither the funding nor information to carry out such a comprehensive approach. Establishment of marine parks may be as a result much more ad hoc.

Box 2 IUCN criteria for inclusion within MPAs and delimiting boundaries

- Naturalness
- Biogeographic importance
- Ecological importance
- Economic importance
- Social importance
- Scientific importance

(continued)

Box 2 IUCN criteria for inclusion within MPAs and delimiting boundaries (continued)

- International or national significance
- Practicality/feasibility

Seven-stage sequence of decision-making in establishing and managing an MPA

- (1) Legal establishment of boundaries
- (2) Zoning
- (3) Enactment of zoning regulations
- (4) Specific-site planning
- (5) Specific-site regulation
- (6) Day-to-day management
- (7) Review and revision of management

At each stage, the following should be taken into account

- Geographical habitat classification
- Physical and biological resources
- Climate
- Access
- History
- Current usage
- Management issues and policies
- Management resources

The variety of existing practice in selection of MPAs has led IUCN to draw upon best practice and experience to identify criteria for deciding upon areas within MPAs or for determining their boundaries (Box 2).

Economic Value of Marine Parks

The economic value of marine parks has been the key driver for much of their development especially in tropical areas where coral reefs are often the main attractions and where, as a result, considerable tourism income is generated. Marine parks both attempt to provide appropriate levels of protection to the natural resources of the parks and to maximize tourist income. These objectives, however, are often incompatible. However, organizations such as the Marine Conservation Society have played a role in developing codes of conduct for divers, which point out the risks of permanent damage to coral. There may be conflicts between traditional users of newly designated parks, especially fishermen, and in some areas, for example, the Seychelles, steps were taken to allow traditional uses to continue. Dixon and Sherman (1990) showed that the Virgin Islands National Park produced considerable profits for the local economy.

Involvement of the Community

Ultimately, the success and sustainability of marine parks depends upon the people who use them as visitors and the local community that often depends upon them for its livelihood. So, for example, in the Balicasag Island Municipal Marine Park in the Philippines key elements in its development included involving the project community worker into the community and baseline data collection, education, organization of a resource management core group, building responsibility around beneficial projects, and formalizing the community in management of the park. Continuity of management was essential for success – it means that the local community has a consistent point of liaison within the park (White and Dobias 1990). If visitors enjoy the parks, but cause irretrievable damage to the ecosystems and disrupt the livelihoods of the human communities which depend upon food from the sea, they will destroy for future generations the same opportunity to enjoy and wonder at the beauty of the undersea world. Users are expected to respect and to help preserve the parks. This depends upon responsible use and it is now common to inform recreationists, especially divers, of good practice. Typically, this information is available both on websites as well as at the parks themselves. It usually focuses on

1. *Good diving practice*, such as obtaining local diving information from qualified local diving operator before diving and maintaining proper buoyancy control. This includes being aware of and obeying all the marine parks regulations.
2. *Prevention of direct damage* to coral reefs, by using rest floats provided by the park authorities, and using only the mooring buoys in the parks. Dropping anchors on reefs causes serious damage.
3. *Avoidance of impacts* on the underwater plants, animals, and fish. This means being aware of potential impacts on aquatic life through interactions with it; leaving litter, feeding fishes or provoking them, and collecting corals and shells for souvenirs should be avoided. Underwater photography is encouraged instead. There is also an expectation that any environmental disturbances or destruction observed during dives are reported to marine park centers so that park staff can prevent further damage or reduce its impact.
4. *Involvement in and support* for marine park activities, such as education and practical conservation work to help conserve the marine habitats and species.

Note. The International Union for the Conservation of Nature (IUCN) World Conservation Monitoring Centre (WCMC) in Cambridge, England, is the most important source of information regarding protected sites. Further reading on this topic may be found in Cognetti (1990), Forest and Park Service (2000), Halsey (1985), Kenchington (1991), Kenchington and Hudson (1984), and Salm and Clark (1984).

Conclusion

Marine parks, MPAs, and similar designations have a common goal: better understanding of the marine environment for the benefit of all who depend upon them. They depend upon sustained resources and incomes, the latter often from tourism, but this can be a source of conflict. Fortunately, there is a growing awareness, often because of the excellent public education programs that many parks run, that these are areas which must be nurtured and managed with the highest quality of understanding of their natural and human systems. Lien and Graham (1985) recognized that the establishment of marine parks would be difficult but rewarding. Marine ecosystems would present new challenges, partly because of the natural characteristics and partly because legal and constitutional problems would confront park managers and planners. They anticipated that research, management, and education would present technological problems which would need new and creative approaches. It would be untrue to say that all those challenges have been overcome, but the spread of marine parks, the common acceptance that these represent a window on the ocean commons and the increased use of new technologies to explore, understand, explain, and nurture the marine ecosystems mean that the challenges which Lien and Graham foresaw have been identified and confronted. As these areas come under greater pressure during the early twenty-first century, the challenges will remain.

Cross-References

- [Archaeology](#)
- [Coastal Management Practices](#)
- [Conservation of Coastal Sites](#)
- [Coral Reefs](#)
- [Economics of Beaches](#)
- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Tourism and Coastal Development](#)

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Marine Terraces

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A marine terrace is any relatively flat, horizontal, or gently inclined surface of marine origin, bounded by a steeper ascending slope on one side and by a steeper descending slope on the opposite side. In temperate regions, marine

Paolo A. Pirazzoli: deceased.

terraces often result from marine erosion (abrasion or denudation) (*marine-cut* terraces, or shore platforms) or consist of shallow-water to slightly emerged accumulations of materials removed by shore erosion (*marine-built* terraces). In inter-tropical regions they may also result from bioconstruction by coral reefs and accumulation of reef materials (reef flats).

The development of a series of elevated, stepped marine terraces usually corresponds to the superimposition of eustatic changes in sea level and of a tectonic uplifting trend. Uplifted terraces act in this case as a continuous tape recorder, each step developing when the rising sea level overtakes the rising land. Each raised terrace, eventually capped by marine and/or alluvial materials, is a fossil counterpart of the present-day terrace, platform, or reef flat. For relatively rapid uplift rates (e.g., >1 mm/yr), each marine terrace corresponds to a different interglacial period or stage and ages usually increase with elevation, the uppermost levels being also the less well-preserved. For slower uplift rates, marine terraces may have a polycyclic origin, sea level returning again at the terrace level after a period of exposure to weathering.

When reconstructing past sea-level changes from the study of datable raised marine terraces, two assumptions are usually made: (1) that the eustatic sea-level position corresponding to at least one raised terrace is known and (2) that the uplift rate has remained essentially constant in each section. From these assumptions, the eustatic sea level can be calculated for each dated terrace (Pirazzoli, 1993).

Bloom and Yonekura (1990) proposed a calculation procedure in which the assumption of a constant rate of dislocation becomes unnecessary when a sequence of dated emerged terraces is found at different heights on several transects.

Example of a Sequence of Stepped Quaternary Marine Terraces: Sumba Island, Indonesia

An exceptional sequence of raised coral reef terraces is preserved at Cape Laundi, on the north coast of the island, between the present sea level and an ancient patch reef now at 475 m above sea level. At least 11 terraces are wider than 100 m, 6 of them even being over half-a-kilometerwide. Most of these terraces are polycyclic in origin and can be subdivided into a number of secondary levels by more or less continuous scarps or by alignments of fossil reef ridges. ESR, Th/U, and ^{14}C date estimations of almost unrecrystallized corals, most of them in growth position, have made possible the identification of.

Stage 9 (ca. 330 ka ago) and of those corresponding to isotope Stage 15 (ca. 600 ka ago). When the most likely uplift trend deduced from the present altitude of dated terraces (ca. 0.5 mm/yr) is extrapolated to the whole raised section, most geomorphological features appear to correspond to interglacial stages, up to Stage 27 (ca. 0.99 Ma) for the uppermost patch reef (Pirazzoli et al., 1993) (Fig. 1).



Marine Terraces, Fig. 1 The uppermost sequence of marine terraces at Cape Laundi, Sumba Island, Indonesia. Behind the very wide Terrace II (in the foreground, at elevations between 135 ± 10 m and 110 ± 5 m,

developed during isotope Stages 9 and 7), six major steps of coral reef terraces, visible in the background up to 475 m, developed between about 330 ka and 1 million years ago (Photo B466, September 1988)

Cross-References

- [Changing Sea Levels](#)
- [Coral Reefs, Emerged](#)
- [Eustasy](#)
- [Paleocoastlines](#)
- [Pleistocene Epoch](#)
- [Sea-Level Indicators, Geomorphic](#)
- [Tectonics and Neotectonics](#)
- [Uplift Coasts](#)

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Mass Wasting

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Mass wasting, the movement of material downslope by gravity, occurs as slopes evolve toward stable, equilibrium forms. Active coastal slopes are often in short rather than long-term stability, because of wave undercutting, oversteepening, and the removal of basal debris. For convenience, a distinction is made in this discussion between the types of mass wasting that occur on rock and cohesive clay coasts. Some slope movements occur on both types of coast, however, and many cliffs consist of variable combinations of rock and clay. Nevertheless, translational slides and the free fall of material from steep slopes tend to be typical of rock coasts, whereas deep rotational slumps and shallower slides and flows of wet material are more common on cohesive clay coasts (Trenhaile 1987, 1997; Sunamura 1992; Viles and Spencer 1995).

Rock Coasts

Fresh rock surfaces and the presence of debris at the foot of cliffs testify to the importance of rock falls on many coasts. Although they occur more often than deep-seated slides,

most falls are much smaller, and they are therefore less frequently reported. Rock falls are particularly common in well-fractured rocks where there is undercutting by wave action, bioerosion, or chemical solution, or differential erosion of weaker rocks at the cliff base. In cool environments, rock falls tend to occur when frost is most active, although this may reflect high water pressures in rock clefts generated by snow melt and precipitation in spring and autumn, rather than the effects of frost action. It has also been suggested that tides can generate damaging fluctuations in water pressure in rock cliffs. This could explain why the size and frequency of rock and slab falls in the Bay of Fundy, for example, increase eastwards, as the tidal range increases. Many falls are caused by the reduction in confining pressures resulting from cliff erosion and retreat, and the formation of tension cracks parallel to the cliff face. Deep tension cracks in massive, cohesive rocks allow large slabs, as opposed to rock rubble, to be released. Slabs can also topple or overturn by forward tilting, particularly in rocks that consist of columns defined by joints, cleavage, or bedding planes which dip into the rock mass. Topples are able to occur on slopes whose gradients are lower than those associated with landslides.

Rock and slab falls, sags, and topples and other surficial failures are generally the result of weathering, basal erosion, and hydrostatic pressures exerted by water in rock clefts, whereas deep-seated slides occur in rocks that have been weakened by alternate wetting and drying, clay mineral swelling, or deep chemical weathering. Translational movements occur along straight slide surfaces. They are structurally controlled, and are usually associated with seaward dipping rocks, alternations of permeable and impermeable strata, massive rocks overlying incompetent materials, or argillaceous and other easily sheared rocks with low bearing strength. The failure surface is frequently a bedding, cleavage, or joint plane. Deep-seated rotational slumps are most common in thick and fairly homogeneous deposits of clay or shale, but more irregular slumps can occur in rock, where they tend to combine the characteristics of slumps with those of translational slides.

Coastal landslides have been particularly well-documented in the Chalk along the southern coast of England and along the Pacific Coast of the United States. In southern England, most slides have occurred where chalk and other arenaceous and calcareous sedimentary rocks are underlain by clay, shales, marls, or mudstones. Most slides are deep-seated rotational movements, although they also failure involve mudslides, slab, surficial sliding, toppling, sagging, and rock falls. Many slides have occurred in seaward dipping igneous and sedimentary rocks on the coast of northern Oregon, and along seaward-plunging synclines in southern California. A bentonite bed in shales and limestones provided an incompetent horizon for sliding at Portuguese Bend, near Los Angeles, whereas the large block glide at Point Firmin,

about 8 km to the east, occurred along the bedding planes of shales that dip seawards at between 10° and 22° .

Cohesive masses fail when there is a critical combination of slope height and angle, in relation to the strength and bulk density of the material. The occurrence of joints, bedding planes, faults, and other discontinuities are more important, however, than the strength of the rock *per se*. Terzaghi (1962) considered the effect of the joint pattern on the critical slope angle of cliffs – the steepest gradient which can be maintained in long-term stability. His analysis showed that slope stability in jointed rocks largely depends upon the joint pattern and orientation, and to a lesser extent upon the effective cohesion of the rock mass. The principles of rock mechanics are increasingly being applied and incorporated into simulation models to determine failure mechanisms and to account for erosion rates and variations in the form of coastal cliffs, and to account for the formation and shape of marine caves (Davies and Williams 1986; Allison and Kimber 1998). In southern Wales, for example, it has been shown that translation failures commonly result from the formation of tension cracks near the cliff face, and toppling where there are tension cracks and basal undercutting. Translational failure is dominant where mudstones are prevalent at the cliff base, but toppling is more common where limestones provide a fulcrum at the foot of the cliff. The safety factor is inversely related to the ratio of the depth of undercutting at the cliff base to the distance of the tension crack from the cliff face. Numerical modeling, based on a range of geo-technical parameters, has also been used to predict the most likely modes of failure in the cliffs of southwestern Wales (Davies et al. 1998).

Landsliding events are generally triggered by the build-up of groundwater, and they tend to take place during or shortly after snowmelt, or prolonged and/or intense precipitation. Groundwater reservoirs provide a water supply that is less dependent on seasonal rainfall, and further instability can result from the ponding of water in surface depressions that were created by older landslide events. Water washes out beds of sand in potential slide masses, causing swelling and the generation of pressures in argillaceous rocks, it generates high pore and cleft water pressures, and it softens colluvial materials allowing them to flow. The lubricating effect of water is not considered to be important, however, and the increase in weight of the potential slip mass as a result of water absorption is only of minor significance. Water can initiate slides in permeable rocks that overlie impermeable materials, and where massive, heavy rocks are on top of wet, impermeable clay or shale. Water from septic systems, irrigation, runoff disruption, and other human activities are playing increasingly important roles in some areas (Griggs and Trenhaile 1994). For example, septic tanks and cesspools associated with residential development may have exacerbated slope instability at Portuguese Bend in southern California, and irrigation may have played a similar role at Point Firmin.

There is also a close cause-and-effect relationship between wave action, which undercuts and steepens coastal slopes, and the occurrence of landslides. Therefore, the degree of protection afforded to a cliff by beach material can be of enormous importance in determining its stability. The damming of rivers and the building of groins and other coastal structures has reduced longshore sediment transport in some areas, depleting beaches and exposing the cliffs to more vigorous wave action. For example, at Newhaven, in southern England, the build-up of coarse clastic material behind a jetty has provided protection to the cliff behind, whereas interruption of the littoral drift by the development of Folkestone Harbor may have been responsible for the increased activity of slides in the late nineteenth and early twentieth centuries.

Changing sea level, whether on a geological or tidal time scale, causes the water table to fluctuate within potential slide masses. Pore and cleft water pressures vary with tidal ebb and flow, and cliffs may be most unstable when there are high water tables and low tides. Landslides can occur during high tidal periods, however, because of the increased ratio between pore pressures and the total overburden at the base of the slide (i.e., increased buoyancy at the toe of the slope) (Muir-Wood 1971). In addition to the buoyant effect of the water, it has been suggested that the Portuguese Bend landslide in southern California is most unstable during high tidal periods because of increased pore water pressure, and saturation of the slide material and the basal shear plane at the base of the slide.

Cohesive Clay Coasts

The erosive resistance of a cohesive material is complex – it depends upon the water content and the properties of the pore water and eroding fluid, as well as the clay content and its plasticity, structure, consolidation pressure, and compressive or shear strength. Small pieces of clay can be eroded by pitting and flaking, but larger units can be removed by spalling along fractures and sandy planes.

Cliffs of sand and gravel tend to erode more rapidly than silt and clay, although this can be partly offset by the formation of protective, coarsegrained beaches at the cliff base. The accumulation of coarse sediment determines the accessibility of the cliff base to wave action. Rates and modes of cliff erosion can therefore vary according to the amount of beach material and the morphology of the beach profile (Jones and Williams 1991; Shih and Komar 1994). Eroded fine-grained cohesive sediment is generally carried offshore as suspended load, and as it provides no protection to the cliff, there is generally no relationship between the height of a clay cliff and its rate of erosion.

Because of variations in wave energy, temperature, precipitation, and groundwater pressures, the subaerial and marine

erosion of cohesive coasts in temperate regions is often markedly seasonal in nature (Brunsden 1984). The strength and stability of clay materials is affected by variations in the groundwater level. Mass wasting also occurs where grains are removed by seepage or piping of outflowing groundwater, which causes back sapping and collapse of the free face, often by toppling. Although seepage is generally associated with coarse silts to fine sands, it can also occur in clays, and it often develops at the base of water bearing fine, predominantly cohesionless, material underlain by an aquiclude.

There is a strong relationship between the frequency and mobility of slope movements in cohesive sediments and the presence of clays with a high proportion of swelling minerals. For example, mudslides in Denmark occur in clays that are dominated by swelling montmorillonite minerals, and the most mobile and furthest advanced portion of a mudflow in northeastern Ireland has the highest montmorillonite and illite to kaolinite ratios. Although slides occur in materials with the highest proportion of swelling clay minerals and the highest plasticity on St. Lucia and Barbados, soil plasticity is also sensitive to the types of exchangeable metal cations held in association with the clay minerals. The occurrence of large amounts of sodium in montmorillonite-rich clays, for example, increases their plasticity.

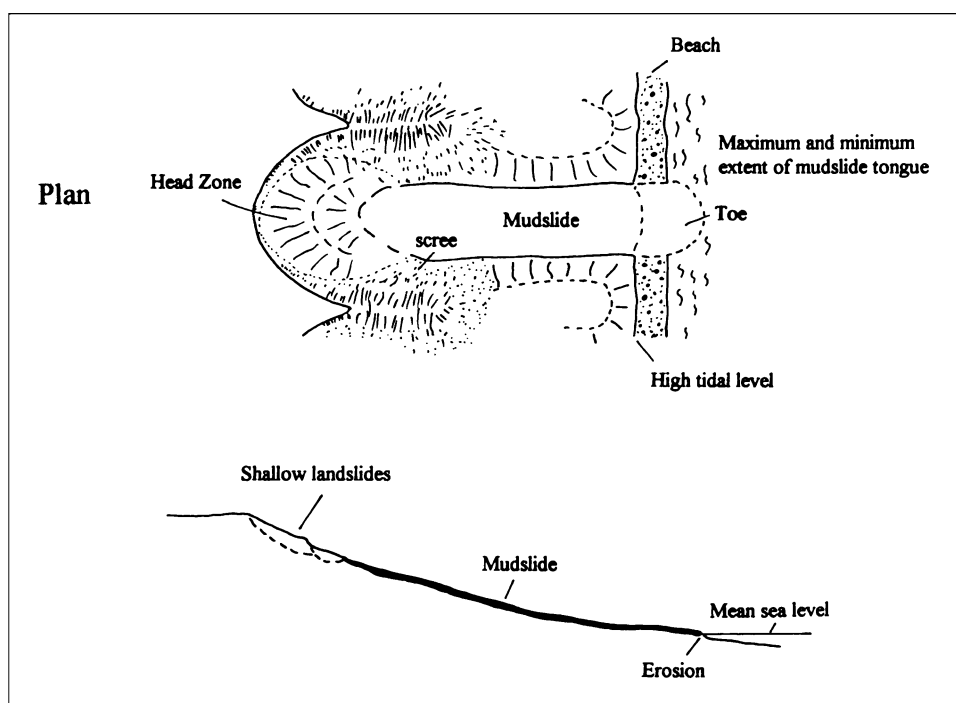
Shallow sliding occurs in cohesive sediments as mudslides or, if the water content is very high, as mudflows (Fig. 1). High moisture content is not required for flow initiation in fine-grained, cohesionless materials, but it must be above the liquid limit if there is a high proportion of clay. Mudslides have a bimodal shape, consisting of fairly steep feeder flows and gently sloping accumulation flows

composed of single or overlapping lobes of clay and other debris. Shallow rotational sliding produces miniature slump blocks along the frontal edge, which has a steeper slope than the accumulation area. The feeder zones begin to move with the seasonal rise in groundwater levels. Surface and subsurface movements in the accumulation zone are largely by shearing at its boundaries, and material can be incorporated in the mudslides from below. Other material is added to the flows by falls and slides from the bowl-shaped head zone and from the sides, and it may help to trigger rapid surges, usually following periods of heavy rainfall. Deep-seated rotational landslides also occur where basal erosion is rapid enough to remove the mudflow debris and then steepen the coastal slope.

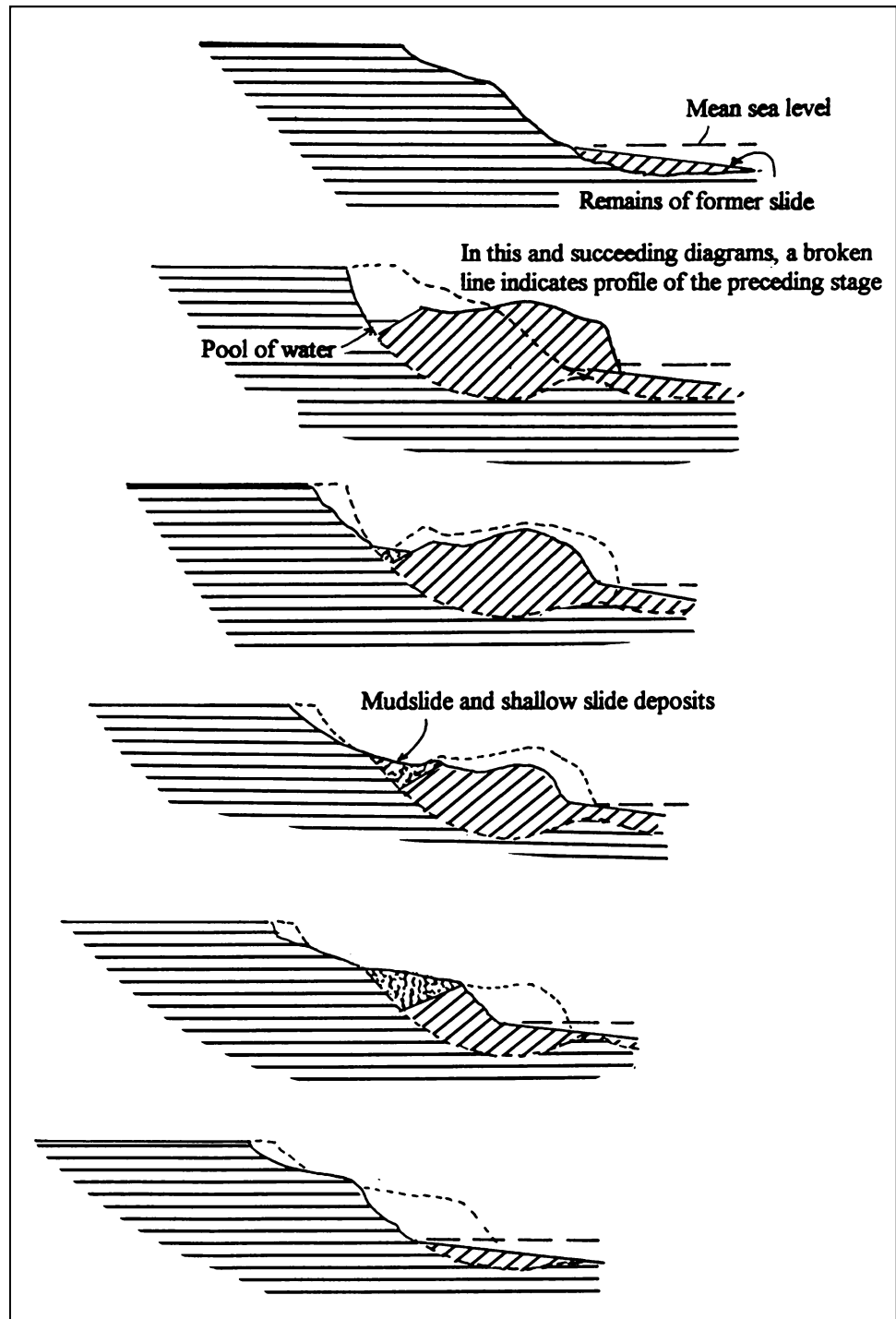
Hutchinson's (1973) form/process classification of cliffs in the London Clay of southeastern England is based on the relative rates of basal marine erosion and subaerial weathering (see "► [Cliffed Coasts](#)"):

1. Type 1 occurs where the rate of basal erosion is broadly in balance with weathering and the rate of sediment supply to the toe of the slope by shallow mudsliding. The slope undergoes parallel retreat and erosion only removes slide material, as opposed to *in situ* clay (Fig. 1).
2. Type 2 cliffs occur where basal erosion is more rapid than weathering. The cliff experiences cyclical degradation, with slope steepness varying between upper and lower values corresponding to toe erosion and degradation and deep-seated, slump-dominant modes, respectively (Fig. 2).
3. Type 3 cliff slopes develop when a cliff is abandoned by the sea and debris is carried to its foot by a series of shallow

Mass Wasting, Fig. 1 Mudslide in London Clay cliff with moderate wave erosion. (From Hutchinson 1973 by permission of Building Research Establishment Ltd., UK)



Mass Wasting, Fig. 2 The cyclical behavior of London Clay cliffs with strong wave erosion. (From Hutchinson 1973 by permission of Building Research Establishment Ltd., UK)



rotational slides. Abandoned cliffs therefore often have a steeper upper slope on which landsliding is initiated, and a flatter lower slope where colluvium accumulates.

Spatial variations in cliff type along the shore of Lake Erie reflect variations in wave intensity and the type of glacial sediment, whereas temporal changes occur in response to

changes in lake level. Type 1 cliffs develop during periods of normal lake levels, when there are moderate rates of cliff retreat; type 2 during periods of high lake levels and rapid cliff retreat; and type 3 during periods of low water levels and slow basal erosion. Quigley et al. (1977) distinguished four types of cyclical instability in this area, according to variations in cliff height, wave height, and rates of cliff retreat. Although Hutchinson's classification is also generally applicable to

cohesive cliffs in Denmark, glacioisostatic recovery and changes in relative sea level are responsible for some significant differences.

Hutchinson's model is generally applicable to a wide range of cohesive sediments, but alternate models of cliff retreat are relevant to specific areas. For example, shallow, planar landslides have been found to be more important than larger but less frequent rotational landslides in glacial sediments in northeastern Ireland. Bromhead (1979) considered that the transition from one type of cliff to another largely depends upon the nature of the material at the crest of the slope, and to a lesser degree on its groundwater hydrology, rather than on the relative rates of subaerial and marine erosion. He suggested that mudsliding dominates if the material at the crest of the slope is similar to that in the rest of the slope. Mudsliding is inhibited, however, if there is stronger or better drained material at the crest. The lack of protective debris at the cliff foot then allows wave action to erode and oversteepen the cliff, ultimately precipitating a deep-seated rotational landslide. These slides tend to be of the multiple rotational type when the cliffs are capped with a thick, hard, and jointed caprock.

Cross-References

- [Cliffs, Erosion Rates](#)
- [Cliffs, Lithology Versus Erosion Rates](#)
- [Hydrology of the Coastal Zone](#)
- [Klint](#)
- [Rock Coast Processes](#)

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Meteorological Effects on Coasts

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The most dramatic and long-lasting meteorological impact on many coasts is in response to storms. Virtually, every continent on earth is variously impacted by storms, the degree to which being a function of many factors including storm intensity, duration and path, as well as antecedent geology of the inner shelf and coast. Cyclones that exert important controls on coasts are generally categorized as hurricanes, tropical, and extratropical storms. Land-and sea-breezes are observed along many coasts and are in response to differential temperatures during day and night; onshore winds during the day develop nearshore sea state, whereas offshore flow in the evening causes wave decay close to shore. Neither effect can equal the impacts of waves, currents, and winds generated during cyclones. The low latitudes are dominated by tropical storms and hurricanes, whereas the mid-and higher-latitudes experience extratropical storms and weather fronts. Frontal systems are associated with extratropical cyclone development.

Mid-latitude cyclones and their associated fronts significantly impact all of the coasts of the United States, although the frequencies of major storms decrease from north to south along both the Atlantic and Pacific coasts. Along the Pacific northwest and New England coasts, mid-latitude cyclone frequencies can range upwards to two or three storms per week during winter months, with average frequencies decreasing to one or less a month in southern Florida and

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southern California. Seasonally, mid-latitude cyclones occur year round along northern coasts, although frequencies and intensities are much lower in the summer. Local and regional coastal configuration and orientation strongly modify exposure to storm processes. For example, Cape Hatteras on the Atlantic and Point Reyes on the Pacific coast are particularly vulnerable given their orientation. Similarly, the northern Gulf of Mexico experiences significant coastal change during frontal events given its general east–west orientation. As an example of rapid and intensive cyclogenesis, the “Storm of the Century,” an intense extratropical cyclone, comparable in strength to a category 1 hurricane, formed in the western Gulf of Mexico in March 1993 and inflicted major damage along the Gulf Coast’s beaches and low-lying areas (Schumann et al. 1995).

Cold fronts are usually initiated by outbreaks of cold air (Polar air masses) from Canada, which advances south, and southeast where they ultimately encounter warm air. Each year up to 30 fronts move south and significantly impact the Gulf of Mexico and southeast Atlantic beaches. As the frontal systems push south, strong southerly winds ahead of the fronts generate deepwater waves whose significant wave height may exceed 5 m. Considerable water level set up occurs along the ocean-facing beaches. A considerable degree of beach erosion and dune breaching accompanies this pre-frontal stage, particularly along low-lying coasts such as Louisiana. This pre-frontal period can last up to several days preceding the arrival of the front. During this period, a considerable volume of water is funneled into the bays, lagoons, and sounds behind the barrier coasts causing elevated water levels. As the front moves across the coast a dramatic wind shift from the south to the north occurs over a matter of hours. While nearshore waves are suppressed in the Gulf and Atlantic during this period of strong northerly winds, high frequency, steep waves are generated in the bays and lagoons where fetch lengths permit. During this post-frontal phase significant wave heights approaching 3 m have been measured in some bays fringing the northern Gulf. Research now suggests that these frontal passages cause dramatic changes to the soundside beaches of the southeast and Gulf bay shores and is the primary process behind chronic coastal erosion.

Each year several of the mid-latitude cyclones evolve into powerful coastal storms with typical storm tracks from southwest to northeast along the Atlantic coast. These storms have traditionally been called Nor’Easters because of the strong winds that blow from the northeast ahead of the warm fronts. The most severe occur in October and November and again in March and April. These storms usually develop and intensify as they move parallel to the coast toward the northeast. The Ash Wednesday storm of 1962 was one of the most powerful Nor’Easters to impact the East coast of the United States. Generating deepwater

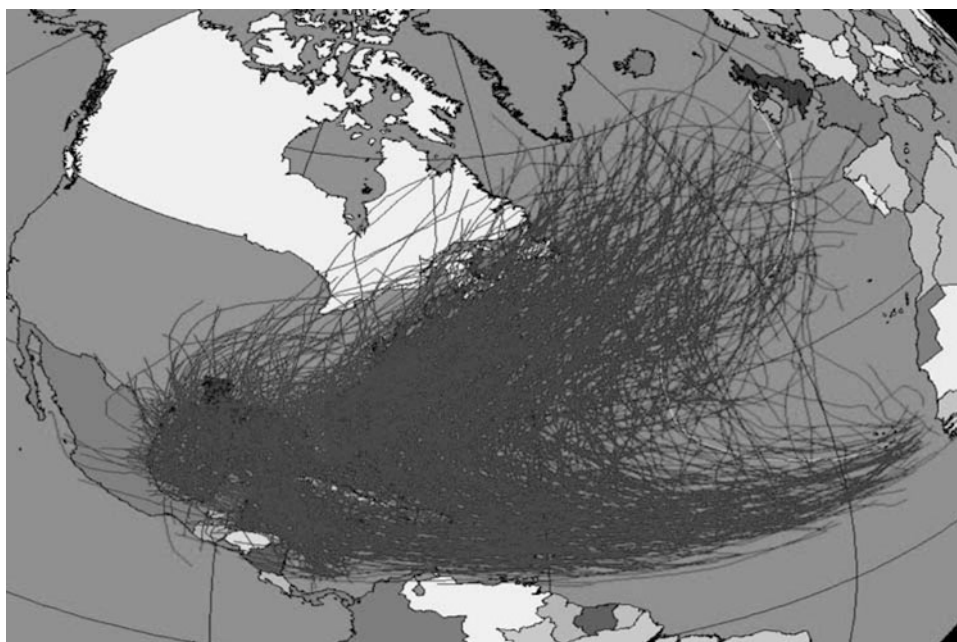
significant wave height in excess of 10 m, the Ash Wednesday storm caused severe erosion along a 1,000 km stretch of coast from North Carolina to Long Island, New York. Because of the danger to life and property and the importance of predicting the explosive dynamics of storm development, a concerted research effort has been directed toward enhancing our comprehension of these systems. After the Ash Wednesday storm, a detailed compilation of powerful coastal storms was compiled (Mather et al. 1964) and more recently, the major coastal storms from New England to Florida have been classified by synoptic weather situations and evaluated in terms of changing frequencies of severe events through many decades (Dolan and Davis 1992).

Tropical storms and hurricanes also impact the entire US mainland coast from New England to Texas, and to a much lesser extent, southern California. The season extends from June through November and thus there is overlap with mid-latitude cyclones in the fall. Unlike mid-latitude cyclones, tropical storms and hurricanes originate over the eastern tropical Atlantic, Caribbean Sea, and Gulf of Mexico. Tracks are generally westward over the tropical Atlantic then recurve northward to the North American coast (Fig. 1).

Hurricanes are much smaller than mid-latitude cyclones, and unless the track is close to and parallel with the coast, the severe impacts are restricted to much shorter segments of the coastline (Muller and Stone 2001).

A direct strike by a major hurricane is more devastating than mid-latitude cyclones because wind speeds, storm surges, and waves are generally much higher. Because of the counter-clockwise circulation of surface winds in a hurricane in the Northern Hemisphere, wind speeds, surge, and waves are generally higher to the right of the system than to the left. Severe morphological change and destruction to infrastructure along coasts is most often limited to a coastal segment no more than 25–50 km to the right of the location at which the storm makes landfall (Muller and Stone 2001). At many locations in the southeastern United States, tropical storms impact the coast as frequently as once every three to 4 years on average, whereas the recurrence of major hurricanes is much more infrequent. As an example, in south Louisiana and south Florida the recurrence interval for these storms is every 10–15 years on average, and only once every 100 years along the more sheltered coast of Georgia (Elsner and Kara 1999; Muller and Stone 2001). In recent years, there has been considerable research conducted on decadal or longer storm-track patterns which result in long runs of years with very limited storm activity followed by shorter runs of years with frequent storms (Gray 1999). Since the mid-1990s, the clustering of storms has resulted in a marked increase in beach erosion along significant portions of the southeastern United States (Stone et al. 1996, 1999). In turn, this has resulted in a marked increase in new beach nourishment and renourishment projects costing tens of millions of US dollars.

Meteorological Effects on Coasts, Fig. 1 North Atlantic hurricane tracks from 1886–1996. (Data obtained from National Oceanic and Atmospheric Administration, National Hurricane Center)



The coast of southern California and Baja California are also susceptible to tropical storm and hurricane impacts that move from southern Mexico northwestward parallel with the coast. Each hurricane season there are numerous tropical storms and hurricanes generated off the southwestern coast of Mexico, however, most storms eventually move to the west out to sea. The rare storm that affects southern California is feared more for heavy rains and subsequent flooding rather than high winds, surge, and waves. One of the most memorable of these events caused much unexpected flooding in the Los Angeles basin in September 1939 (Hurd 1939).

The coasts of the Great Lakes are subject to frequent mid-latitude cyclones, normally moving from southwest to northeast across the lakes and eventually down the St. Lawrence River valley to the Atlantic. Frequencies of mid-latitude cyclones and associated fronts can be as high as six or more per month from October to May, with frequencies and especially intensities much lower in summer. Similar to the Atlantic coast, the truly destructive great storms occur most frequently during fall and spring. Cold fronts in the late fall and early winter, when lake waters are much warmer than cold polar air behind the fronts, are also associated with Great Lakes snow squalls occurring along the eastern and southeastern lake shores. This is pronounced where the orographic effects of hilly terrain amplify the lake effect snow squalls. These are most apparent over the Keweenaw Peninsula in Upper Michigan, and southeast of Lake Erie and Ontario in New York.

Along the Pacific coast from the Canadian to Mexican border, mid-latitude cyclones are again the primary storm events, with the season extending normally from October through April. There are more storms at northern rather than southern locations, with southern California experiencing

approximately one to two storms at most per month. These storms tend to originate along the eastern coast of Asia or over the Gulf of Alaska, so that the storms arrive in an occluded stage, with the warm and cold fronts not particularly noticeable at the surface. Instead, the warm tropical air in these systems has typically been lifted above the surface, resulting in very large storms with gale-force winds along the coasts, and widespread precipitation occurring as snow in the mountains. Precipitation and melted snow can generate mudflows and floods that endanger and destroy homes in the narrow canyons near the coast. These storms tend to stall along the coast so that the duration of storm weather can often last for two to 3 days, with high waves generated by southwesterly winds causing significant erosion of beaches along the Pacific coast. This problem is exacerbated when the storms coincide with high astronomical tides.

Cross-References

- ▶ [Beaufort Wind Scale](#)
- ▶ [Climate Patterns in the Coastal Zone](#)
- ▶ [Storm Surge](#)
- ▶ [Wave Climate](#)
- ▶ [Wave Environments](#)

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Microtidal Coasts

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Coasts where the tidal range (difference between successive high and low tide levels) does not exceed 2 m are commonly referred to as microtidal (Hayes 1979; Davies 1980; Cooper 1994). Such coasts may be composed of a variety of materials and occur in all latitudinal zones and in a variety of energy settings. Their common characteristics are derived from the fact that their small tidal range focuses marine action (via waves and tidal currents) into a relatively narrow vertical range. Hayes (1979) identified a generalized link between tidal range and coastal morphology. Microtidal coasts were characterized by long, narrow barriers with abundant washover features, well-developed flood-tidal deltas, and small ebb-tidal deltas. These characteristics have often led to application of the term “wave-dominated” to such coastlines, although Davis and Hayes (1984) demonstrated that this is not universally true. Some microtidal coasts do exhibit dominance of tidal currents over waves and some macrotidal coasts exhibit wave-dominance. The west peninsular coast of Florida is a good example of a microtidal coast on which wave energy (and sediment supply) is low and yet large tidal deltas develop that are more characteristic of tide-dominated conditions (Davis and Hayes 1984). Microtidal coasts are widespread in oceanic areas (most Indo-Pacific islands), in some semi-enclosed seas (Mediterranean, Baltic, Gulf of Mexico), on parts of continental margins (S.W. Africa, S. Australia, central Brazil, Antarctica) and on lake shorelines (US Great lakes, African Rift lakes).

Open Coast Morphology

Wave processes on microtidal coasts are generally the principal agent of morphological change and sediment transport. Waves constantly expend their remaining energy at the same elevation and thus their effectiveness in coastal modification and sediment transport is greatly enhanced in comparison with areas of greater tidal range.

On coasts where little sediment is available, wave action that is concentrated in a narrow zone may give rise to a well-developed intertidal shore platform, an erosional feature produced by undercutting and removal of rock or semi-consolidated material. In tropical or warm temperate zones, the production of an erosional notch and shore platform on limestone coasts may be aided by chemical processes of weathering. Trenhaile (1980) has established that tidal range does exert a strong control on shore platform morphology.

Where sediment is available, wave processes are effective in reshaping sediment bodies (beach and beachface) by cross-shore and longshore transport. Depositional shores on microtidal coasts thus tend to be rapidly modified by wave processes to attain a form of equilibrium. Examples of such forms include static equilibrium (where wave energy is equalized along the shore), dynamic equilibrium (where sediment input is balanced by sediment losses), and graded equilibrium (where sediments are sorted along the shore by size such that at all points they just exceed the transport capacity of waves) (Carter 1988). Concentration of energy in a narrow vertical band enhances the development of coastal cells that develop in response to wave and current conditions. Some of the best literature on microtidal coasts is based on studies in the US Great Lakes shores and the lack of tidal influence has enabled some basic concepts in coastal geomorphology to be investigated (Dubois 1975).

Because of the limited intertidal area and relatively narrow beaches that limit the potential for entrainment of sand by wind, sand dunes tend to be poorly developed on microtidal coasts and those that are present often owe their existence to oblique onshore winds and/or a falling relative sea level tendency. In contrast, the lack of regular wetting of the beach surface by tidal action may enhance the potential for deflation by wind on microtidal beaches.

River Influences

Rivers that enter the sea on microtidal coasts experience limited tidal influences in their lower reaches. Thus, estuaries tend to be shorter than those of areas with greater tidal range. In addition, if sediment is available in the vicinity of the river mouth, this may be reworked by waves to form a rivermouth barrier that encloses the estuary mouth temporarily or permanently. In a review of the effects of tidal range on tidal inlet

morphology, Hayes (1979) noted that microtidal examples tended to have well-developed flood-tidal deltas and relatively poorly developed ebb-tidal deltas as a result of high wave energy. Deltas in microtidal coastal areas exhibit a range of morphologies in accordance with the level of wave energy. The Mississippi delta, for example, has a distinctive birds foot morphology indicating the relative ineffectiveness of waves in modifying it, whereas the Sao Francisco in Brazil comprises a series of beach ridges that represent the wave modification of deltaic sediment delivered to the coast. In few instances, however, does a single process operate to the exclusion of all others and it is worth noting in this regard that the Mississippi delta does contain chains of barrier islands that are the wave-reworked sections of the delta and which form after fluvial dominance has diminished.

Human Modification

Microtidal coastlines do not contain appreciable intertidal areas and their relative constancy of water level has rendered them favorable for certain forms of human exploitation, particularly for shipping and more recently for recreation. Microtidal coastal areas have recently seen an increase in marina development, particularly in milder climatic regions such as the Mediterranean and Gulf of Mexico.

Development of recreational facilities on microtidal beaches has the advantage of a relatively constant beach area at all tidal stages and thus avoids the regular variability of available space that characterizes beaches of high tidal range. This apparent stability, is however, often deceptive in the longer term. Storms on microtidal coasts may produce elevated sea levels through surges that temporarily raise water levels and which may enhance erosion of backshore areas. Since such storms may be relatively infrequent, a number of such areas have been deemed stable and have been developed. Subsequent erosion of these areas by enhanced wave action has threatened developments and given rise to a succession of shore protection works.

It is interesting to note that human modification of tidal inlets (for example, by barrage construction or inlet closure) may diminish tidal range and produce a change in coastal type in the vicinity of the inlet to microtidal conditions.

Cross-References

- [Atolls](#)
- [Barrier](#)
- [Beach Erosion](#)
- [Beach Processes](#)
- [Beach Ridges](#)
- [Bioerosion](#)

- [Cliffed Coasts](#)
- [Deltas](#)
- [Dune Ridges](#)
- [Eolian Processes](#)
- [Erosion Processes](#)
- [Human Impact on Coasts](#)
- [Longshore Sediment Transport](#)
- [Meteorological Effects on Coasts](#)
- [Rock Coast Processes](#)
- [Sandy Coasts](#)
- [Shore Platforms](#)
- [Storm Surge](#)
- [Tides](#)
- [Wave-Dominated Coasts](#)

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Middle America, Coastal Ecology and Geomorphology

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Introduction

There are several reasons why geomorphology, ecology, and integrated coastal management, have gained attention globally in Mesoamerica (Middle America, Central America). From the coasts of this region, energy, materials, and food are extracted, cities are developing, industries and port are growing, transport and tourism are in expansion, and general infrastructure and environmental technology marketing are increasing (Windevoixhel et al. 1999; Yáñez-Arancibia 1999, 2000). Under this approach, during the 1990s more

than 75 million US dollars have been destined for initiatives and regional projects on integrated coastal zone management, only in Central American countries, by international agencies (Rodríguez and Windevoxhel 1996).

In this framework the “coastal zone” is a broad geographic space of interactions between the sea, the land, freshwater drainage, and the atmosphere, in which the principal interchanges of materials and energy are produced between the marine and terrestrial ecosystems.

Traditionally, Central America includes the following countries: Belize, Guatemala, Honduras, El Salvador, Nicaragua, Costa Rica, and Panama. Nevertheless, from a “mesoamerican” point of view and from an ecological approach because of tropical latitudes, and littorals in two oceans – Atlantic and Pacific – southern Mexico and northern Colombia will be considered, in this paper, as part of the complex of Middle America, because all are integrated components of the functional structure of coastal zone in Central America. The central America coastal zone possesses extensive scenic and geographical wealth as well as great ecological biodiversity (ecosystems, habitats, forcing functions, biological species, functional groups).

Middle America is a “neotropic mosaic.” A mosaic of authors’ training and experiences; a mosaic of roots and cultural heritage resources; a mosaic of social development; a mosaic of ecosystems and ecological approaches; a mosaic of biogeographical regions and biodiversity; a mosaic of climatic zones; a mosaic of pristine areas as well as highly degraded zones. Because of this, to deal with common terms of reference, the purpose of this is to focus on the tectonic history, the regional geography and geomorphology, and to analyze the major ecological systems, as the key focus for understanding the geomorphology/ecology of coastal zones in Middle America (i.e., continental shelf, coastal lagoons and estuaries, mangroves, and coral reefs).

Central America: Antilles Tectonic History

The tectonic history of Central America and the Antilles is extremely complex (Burke et al. 1984; Buskirk 1992). Geological studies are fragmentary, and the application of plate tectonic theories to detailed areas has been controversial (Donnelly 1985; Coates et al. 1992; Suárez et al. 1995). Measurements of present-day plate motions can give a picture of recent movements but the older history of the area is more difficult to interpret; much geological evidence is missing.

The tectonic hypothesis of Duncan and Hargraves (1984) provides the best framework for the Buskirk (1992) zoogeographic analysis for understanding Mesoamerican ecological relationships. Based on plate tectonics determined from the geometry of hotspots on the earth’s mantle, the Duncan and Hargraves model reconstructs the evolution of the Caribbean

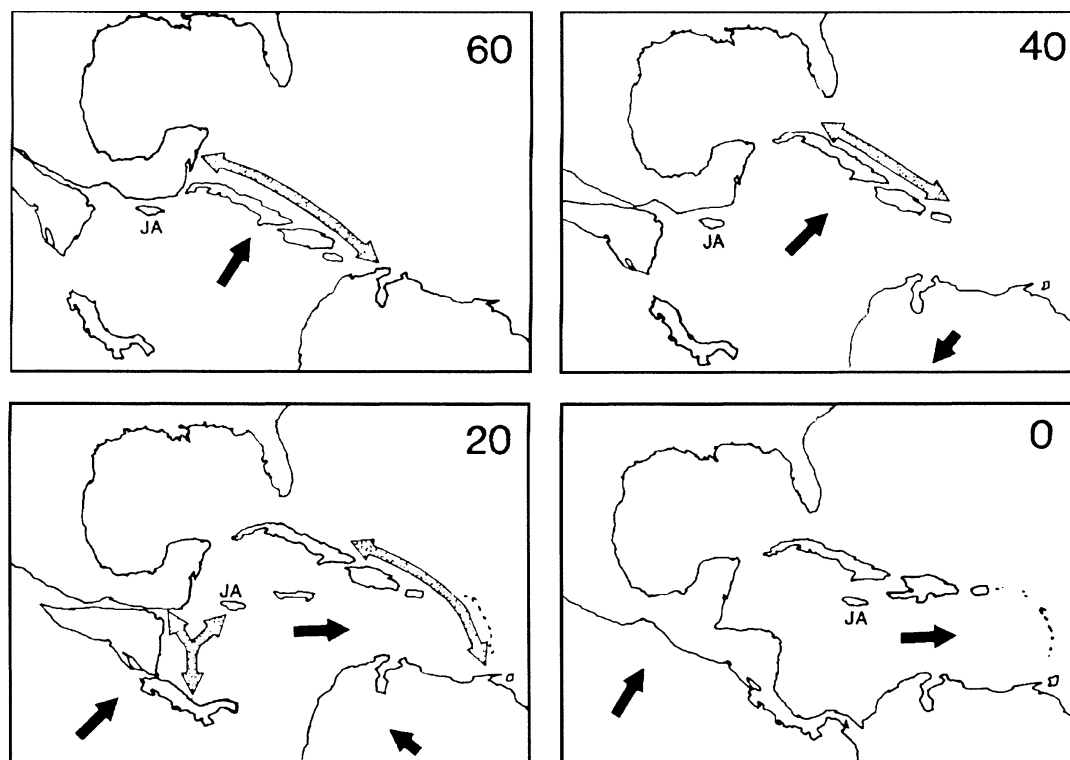
region as follows. In the Cretaceous, as the North and South American plates moved farther apart, subduction movements of the Pacific ocean floor under the west side of the Americas formed the proto-Antilles island arc. In the late Cretaceous, an oceanic plateau formed behind that arc and began to push it northeastward between the continents. This motion was halted at about the Eocene and Oligocene when the arc collided with the Bahamas platform. At this time, subduction of the Pacific ocean floor under the west side of the oceanic platform, in combination with uplift of older Cretaceous island arc features, formed the lower Central American arc. Since the Oligocene there has been eastward movement of the Caribbean plate. Subduction of the Atlantic ocean crust beneath the eastern margin of the Caribbean plate has formed the Lesser Antilles island arc. A subduction zone along the Pacific side of Mesoamerica forms the western boundary of the Caribbean plate.

To this framework, Buskirk (1992) added information about the Greater Antilles used in preparing the sketches in Fig. 1. Solid arrows indicate the direction of major tectonic movements. Land outlines do not indicate coastlines at the time, but are merely for recognition of the relative location of present-day terrains. The position of the Honduras–Nicaragua and Costa Rica–Panama blocks is based on the model of Duncan and Hargraves (1984). Although each of the blocks might be considered a tectonic unit for the last 50 million years, both have been changing extensively during that time with much shearing and internal deformation.

It is suggested that the Greater Antilles island arc (Cuba, northern Hispaniola, and Puerto Rico) moved northeastward during the late Cretaceous after fragmenting its western end (to form Jamaica) in a collision against Mexico–Guatemala. Pollen data from the Eocene and Oligocene of Cuba and Puerto Rico indicated that there was frequent dispersal between the island of this northern Caribbean arc (Buskirk 1992). Jamaica was largely inundated during this time. Since the mid-Tertiary it has moved as the eastern end of the Nicaraguan block.

In Fig. 1 (after Buskirk 1992), probable prevalent pathways of dispersal at 20, 40, and 60 million years ago are suggested by the stippled arrows. Probably, there was a period of dispersal between North and South America in the Cretaceous or early Tertiary, despite the lack of geological evidence for an emergent island arc at that time. In about the Miocene (beginning 23 million years ago), as the Lesser Antilles island chain became prominent, there was dispersal from South America into Puerto Rico, Hispaniola, and Cuba. As Jamaica become emergent, it initially was colonized mostly by the Central American terrapins.

Following the emergence of Jamaica, therefore, geological changes altered the potential for dispersal of terrestrial organisms between Jamaica and Mesoamerica. In the Miocene, the emerging island was situated closer to Central America than



Middle America, Coastal Ecology and Geomorphology, Fig. 1 Hypothetical sketch of Mesoamerican and Caribbean history at 60 (Paleocene), 40 (late Eocene), 20 (early Miocene), and 0 (Present) millions years before present (after Buskirk 1992). Land outlines do

not indicate shoreline at those times but are merely for recognition of the present-day terrains. Solid arrows indicate the predominant direction of tectonic movement. Stippled arrows indicate suggested dispersal pathways

presently, and shallow reefs or emergent limestone banks could have provided stepping-stone conditions for dispersal between Jamaica and the Mesoamerican island chain. During the Pliocene, as Jamaica uplift increased, tectonic activity slowly moved the island eastward as the rise between it and the Nicaragua subsided.

Regional Geography and Geomorphology

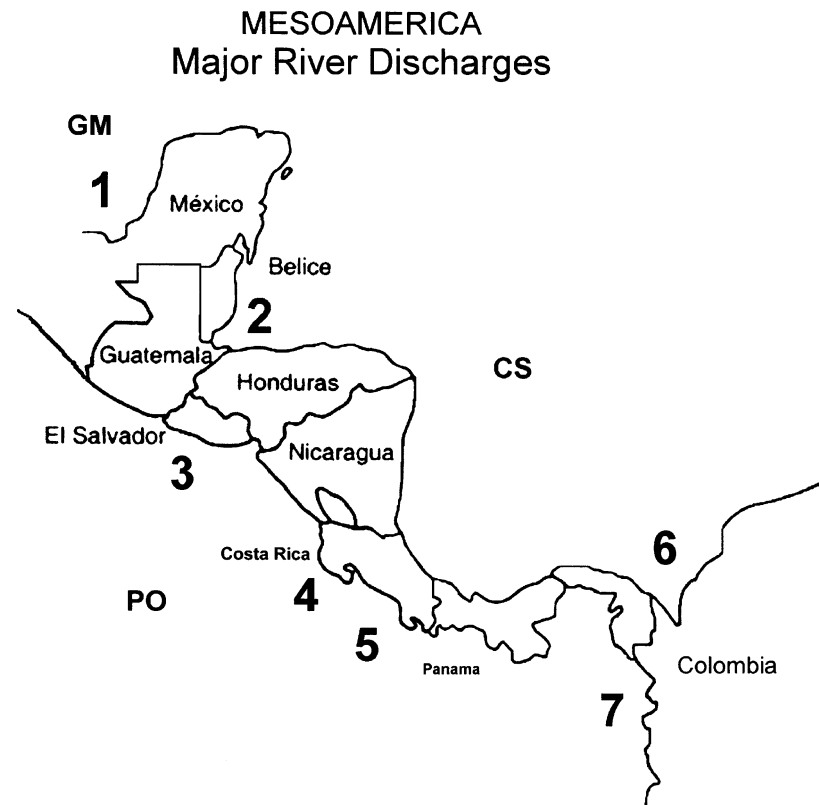
It is not only the tectonic history or the relative areas of continental shelves and the open tropical oceans we need to quantify, if we want to understand the basis of Mesoamerican coastal ecosystem productivity. Rather, we must also know the geographical and geomorphological characteristics of continental shelves, shallow bays, and coastal lagoons and estuarine system features. Are they narrow or wide, flat or steep, what are their superficial deposits, what is the geology of the coastline that lies behind them, and do they lie on the east or the west of the Mesoamerica continental mass? All these characteristics – quite apart from the circulation of water masses over them and the seasonality of river discharges – determine the kind and quantity of ecosystem productivity and biomass production (Deegan et al. 1986; Longhurst and Pauly 1987).

Tectonic activity causes shelves to be narrow on active, subducting margins as along the Pacific coast of Mesoamerica, and allows them to be wider along passive margins, as on the eastern coast of the American continent (Kolarsky et al. 1995; Parkinson et al. 1998). The insular arc in the Caribbean protects shelves from wave action and allows the development of relative wide shelves, such as the Campeche Sound (Mexico), the Magdalena basin (Colombia), or the Orinoco delta (Venezuela) in the northern and southern end of the Atlantic Mesoamerican coastal zone. But earthquakes associated with coastal subduction has been clearly correlated with benthic organisms (Cortés et al. 1992).

Major rivers can modify the morphology of continental shelves: the Grijalva/Usumacinta (Mexico), Dulce (Guatemala), Magdalena (Colombia), and Orinoco (Venezuela) rivers have all widened the adjacent continental shelves and have built fans of sediment down the continental slope on the Atlantic coast of Mesoamerica, even the influence of the Amazon from Brazil must be considered as important in the coastal ecology of Middle America. Figure 2 shows the astonishing manner in which a few major tropical river systems dominate the global transport of sediments from the continents to the ocean coast in Middle America. Recent references for Colombia are (Restrepo and Kjerfve 2000a, 2000b). This amount of sediment discharged to the coastal

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Fig. 2 River transport of water and silt to the continental shelf from the major drainage basins for Middle America. Water discharge in $\text{m}^3 \text{s}^{-1}$, and sediment yield in million tons per year. (1) Mexico, Grijalva-Usumacinta $4400 \text{ m}^3 \text{s}^{-1}$; $126 \times 10^6 \text{ tons yr}^{-1}$. (2) Guatemala, Rio Dulce $1100 \text{ m}^3 \text{s}^{-1}$. (3) El Salvador/Honduras/Nicaragua, $834 \text{ m}^3 \text{s}^{-1}$. (4) Costa Rica, Gulf of Nicoya Tempisque River $50 \text{ m}^3 \text{s}^{-1}$. (5) Costa Rica, Terraba-Sierpes Deltaic River $338 \text{ m}^3 \text{s}^{-1}$; $2 \times 10^6 \text{ tons yr}^{-1}$. (6) Colombia, Magdalena Deltaic River $7200 \text{ m}^3 \text{s}^{-1}$; $144 \times 10^6 \text{ tons yr}^{-1}$. (7) Colombian western discharge $8020 \text{ m}^3 \text{s}^{-1}$; $16 \times 10^6 \text{ tons yr}^{-1}$. GM = Gulf of Mexico, CS = Caribbean Sea, PO = Pacific Ocean



zone is reflected in the distribution of the sediments of tropical continental shelves and seas, and in the high turbidity of coastal seas in the monsoon-type regions of Mesoamerica (i.e., Tabasco-Campeche Mexico, Costa Rica, Panama-Colombia).

The superficial geology of continental shelves is determined by processes that also determine coastal morphology, working in concert with several important biogenic processes of sediment production. Shelf sediment in the Middle America coastal zone includes inorganic components such as sand derived from weathering of rock, and transported by wind, longshore currents, or rivers, and also silts and clays that are mostly river borne. The organic components of shelf sediments include terrestrial plant debris in all stages of decomposition which is either riverborne as particulates, or flocculated from dissolved organic material at the fresh/salt-water interface in the estuaries (i.e., Gulf of Fonseca shared by El Salvador, Honduras, and Nicaragua; Sierpes delta in Costa Rica, both areas in the Pacific coast). The remaining shelf deposits are oolite sand formed by precipitation from dissolved calcium; shelf-sand and large particles of calcareous material derived from molluscan benthos; reef corals and coralline algae, produced largely in situ and not transported great distances; and finally, small particulate organic material from planktonic communities of plants and animals.

Contrary to general conceptions, tropical continental shelves dominated by the effects of mangroves or reef

corals are more the exception than the rule (Longhurst and Pauly 1987). Mangroves are important only where deltaic or a low-lying coastal plain occurs (Kjerfve 1990; Wolanski and Boto 1990; Jiménez 1994; Kjerfve 1998; Yáñez-Arancibia and Lara-Domínguez 1999); and coral reefs are important only where negligible amounts of terrigenous material reach their habitat (Wolanski and Choat 1992; Kjerfve 1998; Cortés 2002).

Major Ecological Systems

Continental Shelf

The eastern Mesoamerican shelf is much more complex, and dominated by terrigenous deposits principally near the mouths of the Grijalva/Usumacinta, Dulce, Magdalena, Orinoco rivers, where spectacular mud regions dominate. This is also true in the Terraba-Sierpes river delta on the western coast of Costa Rica. The continental coastline of Central America is much folded and fractured by tectonic activity, and has basins formed by calcareous deposits. The wide, very shallow Yucatan shelf has offshore banks (notably Campeche Bank), at least one being of atoll form, and the shelf sediments are predominantly calcareous. Along the northern South America coast (the southern end of Middle America), which itself is much altered by large tectonic faults, a shelf is lacking in the Pacific, but because of the outgrowth of the Magdalena

estuary in the Colombian Atlantic a significant portion of shelf exists. Off Venezuela, the shelf deposits are terrigenous and the shelf itself is complex, with the deep Cariaco Trench intruding. Offshore, the Antilles and Caribbean islands are rocky, or bear coral reefs, and the geological environment resembles the western Pacific, an impression strengthened by the occurrence off Belize and Cancun (Mexico) of a Barrier reef complex similar in structure if not in extent to the Great Barrier Reef of Australia.

The western coast of Middle America is dominated by an active subduction zone almost the full length of the continent. Coral reefs and banks are largely absent, and mangroves dominate only along parts of Central America (Chiapas Mexico, Guatemala, Gulf of Fonseca El Salvador/Honduras/Nicaragua, Terraba-Sierpes Costa Rica, and Colombian coasts). Despite the absence of major rivers, terrigenous material is very important. The continental shelf is very narrow. In the Panama Bight, silt and terrestrial organic material contribute to muddy deposits, especially closer to the coast; here, the outer shelf tends to have sandier deposits. The shelf is very narrow or absent off Middle America, where depths fall away down the continental slope to 6000 m in the Middle America Trench. Only off Panama and the Gulf of Tehuantepec in southwestern Mexico is there a significant width of shelf (Tapia García 1998); elsewhere, only in isolated bays and gulf are there shallow areas resembling continental shelf habitats.

Coastal Lagoon Systems

Coastal lagoons here include lagoons, estuaries, and deltaic plains, in the sense of Day and Yáñez-Arancibia (1982), Kjerfve (1986), Yáñez-Arancibia (1987). These systems occur in all tropical oceans (Kjerfve 1994). They are a specially prominent topographic feature on both the Atlantic and Pacific coasts of Middle America, including the northern (Mexico) and southern (Colombia) end (Yáñez-Arancibia 1999). Many stretches of coastline are backed not by solid land, but by coastal lagoons behind barrier island, open to the sea to various degrees, with or without rivers, and of great significance in some coastal fisheries as nursery grounds (Lasserre and Postma 1982; Kjerfve 1994).

Lagoons not associated with coral reefs are most strongly developed on coasts with a history of submergence during the Holocene post-glacial sea-level rise during the last 10,000 years or so (Kjerfve 1986, 1990). Lagoons may also form behind the cusped spits that accrete at the mouth of open estuaries and delta mouths, as in Tabasco/Campeche Mexico (Yáñez-Arancibia and Day 1988). Special evaporative mechanisms may also build lagoons in arid or semiarid regions, characterized by ephemeral inlets (Kjerfve 1994). The formation and maintenance of a lagoon's coastal barrier depends on a balance between the supply of sedimentary material and its removal to deeper water by wave action. Sediment may be

supplied to the outer side of the barrier by longshore currents and normal wave action, or may enter the lagoon itself by beach "washover" during storms as in Amatique Bay Guatemala; Terminos Lagoon, Mexico, Gulf of Fonseca El Salvador/Honduras/Nicaragua (Yáñez-Arancibia and Day 1988; Yáñez-Arancibia et al. 1999). It may also accumulate by the settlement of riverborne silt, by flocculation of organic material at the saltwater-freshwater interface, or by local estuarine plant production especially by a mangrove ecosystem and associated coastal wetlands (i.e., La Encrucijada Chiapas, Mexico; Tortugero, Costa Rica).

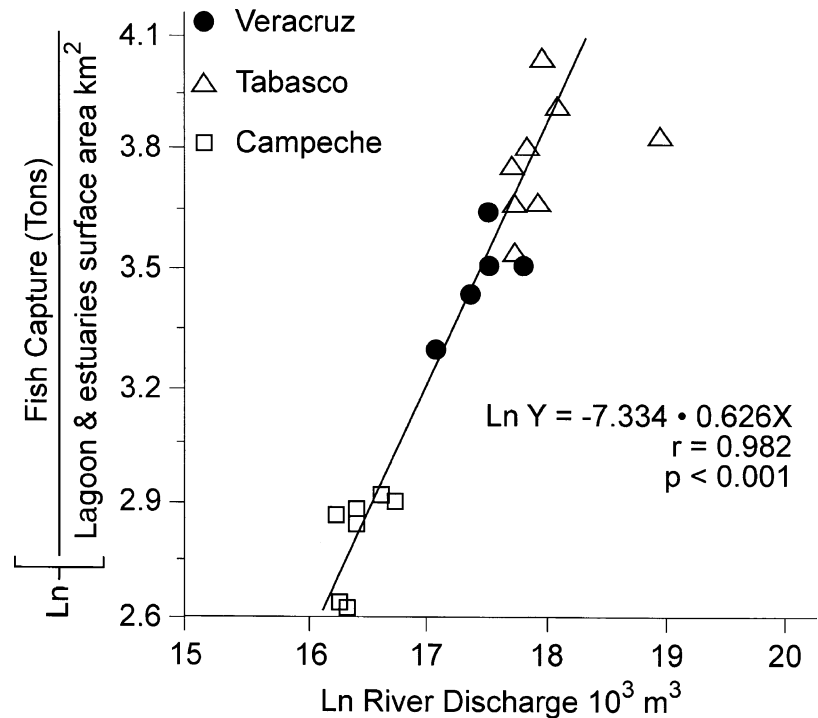
The primary determinant for the existence of lagoon-barrier coasts is a relative small tidal range (Kjerfve 1986). In microtidal (<2 m) environments as in Middle America, wave action is especially important in maintaining linear barrier islands. There is a relationship between tidal range and the width and number of the mouths of tidal channels by which lagoons open to the sea. In the Middle America region, there are few, narrow entrances, and barrier islands are long and narrow. Also of great significance for lagoon geomorphology is the pattern and amount of river discharge that they receive and must carry to the sea; in many places, remains of terrestrial plant cover destroyed by deforestation are having important effects on the sedimentary regime within lagoons. Precipitation/evaporation rate is a further important factor in the evolution and environmental behavior of lagoons once formed. Lagoons in Middle America are in a state of continuing evolution, and many are already filled in by sea-sand transport and the accumulation of both terrigenous sediments and biogenic material. Nevertheless, the remaining complex dominates the coastal ecology in Central America and it is an important factor in the coastal productivity and subsequent fisheries in the region. This family of coastal lagoons differs from those that are formed in association with coral reefs as in Yucatan Peninsula, Mexico, or Belize.

Shelf-Estuary-Lagoon Relationships

There is a general relationship between the amount of water and terrigenous sediments discharged onto continental shelves, and the production there of fish that has been studied particularly well in the Gulf of Mexico (Deegan et al. 1986; Sánchez-Gil and Yáñez-Arancibia 1997), confirming the linkages between shelves, estuaries, and coastal lagoons, coupled with river discharge budgets, estuary area, intertidal area, coastal vegetation area, and fishery harvest (Fig. 3). Such relationships probably can also occur in the Gulf of Fonseca influence area, the Gulf of Nicoya and Gulf Dulce influence areas, and Terraba-Sierpes estuarine plume, in the Pacific coast of Middle America, assumed by analyzing the reference of Jesse (1996), von Wangelin and Wolff (1996), Wolff (1996), Vargas (1995), Leon-Morales and Vargas (1998), Dittmann and Vargas (2001). Seasonally, there is a natural alternation between dry-season conditions

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Fig. 3 Linear logarithmic regression between fishery harvest per unit open waters (lagoons and estuaries) and average river discharge. The data are from the states of Veracruz, Tabasco, and Campeche in Mexico



with the river discharge, and rainy season when the river discharge dominate the deltaic conditions, as well as the winds regime and tidal range throughout the year.

As we might expect from the greater relative area of reef, rock, and especially soft bottom habitats off deltaic systems in the Atlantic coast of Middle America, Longhurst and Pauly (1987), based on selected references, can distinguish four species assemblages of fish that comprise the coastal demersal resources in the tropical west central Atlantic coastal ocean:

1. *Lutjanid community*. Fauna of rock, coral, and coral sand from Florida to Brazil, which is dominant on the Bahamas, the Antilles, and the other Caribbean islands; and on the coast of Yucatan to Panama (Balistidae, Lococephalidae, Lutjanidae 14 spp., Pomadasysidae 3 spp., Serranidae 11 spp., Synodidae), see Longhurst and Pauly (1987), Lowe-McConnell (1987), Arreguín et al. (1996).

2. *Subtropical sciaenid community*. Fauna of soft deposits from the Gulf of Mexico to at least Cape Hatteras, and especially well developed near river mouths influence, including the Mississippi and Grijalva/Usumacinta deltas (Branchiostegidae, Clupeidae, Gerreidae, Polynemidae, Serranidae, Bothidae 19 spp., Sciaenidae 19 spp.), see Longhurst and Pauly (1987), Sánchez-Gil and Yáñez-Arancibia (1997).

3. *Tropical sciaenid community*. Fauna of the soft inshore muddy and soft sand deposits from the southern coast of the Caribbean to Cape Frio in Brazil (Dasyatidae, Ariidae 6 spp., Clupeidae, Gerreidae, Heterosomata 3 spp., Ehippidae, Pomadasysidae, Sciaenidae 19 spp.), see Longhurst and Pauly (1987), Lowe-McConnell (1987).

4. *Sparid community*. Fauna of the sandy and muddy sands of the subtropical regions north of Cape Hatteras and south to Cape Frio with very attenuated representation through the tropical region, that is, Yucatan to Venezuela (Ariidae, Carangidae 9 spp., Clupeidae, Mullidae, Sciaenidae, Sparidae 10 spp., Synodidae or Synodontidae), see Longhurst and Pauly (1987), Lowe-McConnell (1987), Sánchez-Gil and Yáñez-Arancibia (1997).

In the Pacific of Middle America important progress has been done for fish and benthos species assemblages, and after analyzing Jesse (1996), Wolff (1996), Vargas (2001), and Dittmann and Vargas (2001), some analogies can be found in the coastal demersal resources in the tropical coastal zone in the Mesoamerican Pacific in comparison with the Caribbean coast.

Because relatively few fish species are wholly adapted to a life cycle within a lagoon–estuarine system sensu stricto, Longhurst and Pauly (1987) pointed out a controversial discussion on the topic. But it is clear – nevertheless – that in the tropic of Middle America, and especially in monsoon-type areas where estuarinization of the continental shelf occur, the distinction between estuaries and the neritic sea is slighter than in non-monsoon type areas (Yáñez-Arancibia 1985; Yáñez-Arancibia et al. 1994). This approach can be easily applied in Tabasco/Campeche, Mexico; Amatique Bay, Guatemala; Magdalena delta shelf Colombia; Gulf of Fonseca, El Salvador/Honduras/Nicaragua; Gulf of Nicoya, Costa Rica.

There are also sufficient descriptions of specialized estuarine fish fauna for it to be clear that this is often a reality in the tropics of Middle America and adjacent sub-tropical regions,

where the size of the estuary and its tidal regime permit a relatively long flushing period. If the Guntherian hypothesis of estuarine-dependence of continental shelf fish stocks cannot be valid for the tropical regions as a whole (Longhurst and Pauly 1987), at least it can be applied without any doubt in tropical America especially in soft bottom communities on the shelf associated with extensive deltaic systems, river discharge, and broad areas of coastal vegetation (Deegan et al. 1986; Pauly 1986, 1998; Pauly and Yáñez-Arancibia 1994; Sánchez-Gil and Yáñez-Arancibia 1997; Baltz et al. 1998; Chesney et al. 2000; Lara-Domínguez 2001, i.e., Fig. 3).

As a consequence, some important points have emerged from those studies; (1) The utilization of the lagoon–estuarine environment is an integral part of the life cycle of numerous fishes, particularly in the Neotropics (Middle America). (2) The lagoon–estuarine environment is mainly utilized by juveniles and young adults. (3) There are a greater number of fish species in tropical and subtropical lagoon–estuarine ecosystems than in comparable temperate or boreal systems. (4) Second-order consumers are more abundant and diverse than first-order consumers or top carnivores. (5) Functional components are (a) freshwater spawners occurring in waters <10 ppt salinity, (b) brackish water groups limited to 10–34 ppt salinity, (c) marine spawners occurring in waters > or = 35 ppt salinity. (6) Three groups of fish occur in lagoon–estuarine systems (a) resident species, those which spend their entire life cycle within coastal lagoons, (b) seasonal migrants, those which enter the lagoon during a more or less well-defined season from either the marine or the freshwater side and leave it during another season, (c) occasional visitors, those which enter and leave the lagoon without a clear pattern within and among years. To these, two other groups may be added: (d) marine, estuarine-related species, those which spend their entire life cycle on the inner sea shelf under the estuarine plume influence, and (e) freshwater, estuarine-related species, those which spend their entire life cycle in the fluvial-deltaic river zone, associated with the upper zone of the estuarine system. Finally, if the “recruitment” both biological and fishery, utilize coastal lagoons and estuaries, those resources are “estuarine-dependent,” but if that resources utilize regularly the option of the estuarine plume on the continental shelf, they are “estuarine-related” or estuarine opportunistic.

An advanced approach, following shelf–estuary–lagoon relationships, and the compressive models of trophic fluxes in key coastal Mesoamerican ecosystems, has recently been done. Construction of mass-balance trophic models of aquatic ecosystems, including complete food webs, from primary producers to top predators, allows numerous inferences on ecosystem status and function and can serve as a base for dynamic simulation models, as well as comprehension of shelf–estuary–lagoon relationships and the coastal ecological processes evolved. In Mesoamerica, this is a clear coastal

ecological perspective, that is, in the southern Gulf of Mexico (Pauly et al. 1999), in the Dulce Gulf (Wolff et al. 1996), and in the Gulf of Nicoya (Wolff et al. 1998).

For instance, the Dulce Gulf acts differently from most tropical coastal ecosystems, as it is dominated by biomass and energy flow within the pelagic domain and resembles rather an open ocean system than an estuarine one. Due to its low benthic biomass and low overall productivity there seems no potential for a further development of the demersal and semi-demersal fishery inside the Gulf; and an increase of the fishing pressure on pelagic fish would seriously threaten the large resident predators, such as dolphin, sharks, and large birds species (Wolff et al. 1996). In the Gulf of Nicoya, the model shows that shrimps occupy a central role within the Gulf as converters of detritus and other food into prey biomass for many predators, that seem to be simultaneously affected by the over exploitation of this resource (Wolff et al. 1998), as occurs in classical tropical coastal estuary–shelf relationships.

In the southern Gulf of Mexico, the trophic model shows the estimation of the primary production required to sustain fisheries offshore, coupled with tropical lagoon–estuarine systems (Terminos Lagoon), which can be determined given knowledge of fishery catches, from the trophic levels of the exploited groups represented in those catches, and the transfer efficiency between trophic levels (Pauly et al. 1999).

These examples will be more useful the smaller the scale that is used for the spatial stratification of the overall models; also more local models will mean more consideration of local data and constraints and hence a more realistic overall model and more realistic aggregate statistics (i.e., primary production required to sustain the fisheries of those key coastal ecosystems in Mesoamerica).

Mangroves, Coral Reefs, Islands, and Other Related Ecosystems

Not only coral reefs, or river basins and associated wetlands, are a common coastal landscape in Middle America, but the most important critical habitats are the mangroves in the coastal zone (Jiménez 1994; Kjerfve 1998; Yáñez-Arancibia and Lara-Domínguez 1999). The relative importance of mangroves for each Central America country is illustrated by Windevoixhel et al. (1999). This rough representation shows that mangroves are the most important forest formation in certain countries and they should be given priority in management and conservation. The entire coast is characterized by the presence of mangroves, with nine species present in five genera on the Pacific, and four genera on the Caribbean and the Atlantic coast of Mesoamerica. Central America possess 8% of the world's mangroves, and represents 7% of natural forest in the region (Jiménez 1994; Suman 1994). The greater extensions of mangrove on the Pacific are found

Middle America, Coastal Ecology and Geomorphology, Table 1 Biophysical characteristics of the Central American coastal zone (from Windevoxhel et al. 1999, with permission from Elsevier)

Biophysical aspects	BEL	GUA	HON	E.S.	NIC	C.R.	PAN	Total
National Territory (km ²)	22,965	108,889	112,088	20,935	118,358	50,900	77,082	511,217
Population (millions) 1994	0.209	10.322	5.497	5.641	4.275	3.334	2.611	31,889
Density (pop/km ²)	9.1	94.8	49.0	269.5	36.1	65.8	33.9	62.4
1994% population in the CMZ	39	26	15	13	24	7	50	21.6
Length of coast (km)	250	403	844	307	923	1376	2500	6603
Coast/territory rate Contin. Platform, 200 m (km ²)	0.01	0.003	0.007	0.001	0.008	0.03	0.03	0.01
EEZ area (thousand of km ²)	8250	12,300	53,500	17,700	15,800	57,300	237,650	
	n.d.	99.1	200.9	91.9	159.8	258.9	306.5	1117.1
Mangrove areas (ha)	11,500	16,000	14,5800	26,800	15,5000	41,000	170,800	566,900
Coral reefs (km)	474	1	364	1	455	2.5	320	1617.5
Surface drainage Pacific (%)	0	21	18	100	10	53	69	39
Surface drainage Carib. (%)	100	79	82	0	90	47	31	61

off the coast of southwestern Mexico, Guatemala, Costa Rica, Panamá, and the Gulf of Fonseca (Table 1). Pacific mangroves maintain lesser biodiversity associated with root systems than those in the Caribbean, where the small changes of tides provide conditions of environmental stability for root-associated organisms, including important sea grass beds associated with mangrove systems, as in Campeche and Yucatan Peninsula Mexico, Honduras, Nicaragua, and Costa Rica. However, Pacific mangroves are more extensive due to the greater tides and coastal topography. These conditions offer greater areas that exclude competitors and favor facultative halophytes with saline intrusion. On the Caribbean and the Atlantic coast of Mesoamerica, the most extensive areas of mangrove are found in Honduras and Nicaragua, and also southeast México.

In general terms, Pacific coast coral reefs are less extensive and diverse than those on the Caribbean (Cortés 2002). Live coral formations have been described in El Salvador, Costa Rica, and Panama. Coral communities in the Pacific are richer off the southwestern coast of Costa Rica, and particularly in Panama, where at least 21 species have been reported (Cortés 1997a). On the Caribbean side, coral reefs can be found in all of the countries in the Atlantic coast of Mesoamérica (Cortés 1997b). The Belize coral barrier or reef system (Mexico, Belize, Guatemala, Honduras) runs more than 220 linear kilometers and contains atolls and other formations practically unique in the Caribbean Sea. More than 60 species of coral have been reported to exist on the coast (Woodley 1995; Gibson and Carter 2002), but the total number of species associated with the coral reef system is still uncertain. Offshore the Antilles and Caribbean island are rocky, or bear coral reef, and the geological environment recalls that

the western Pacific, an impression strengthened by the occurrence off Belize and Cancun (Mexico) of a barrier reef complex similar in structure, if not, in extent, to the Great Barrier Reef of Australia.

The Pacific coast has long sandy beaches with a broad range of texture and color, and predominant clear coastal waters. Beaches on the Caribbean and the Atlantic coast are less extensive as a result of current patterns and sea cycles, as well as the oceanographic and geomorphologic factors, and because the presence of important river basins, predominant turbid coastal waters, with the exception of Yucatan Mexico, Belize, and southeastern Costa Rica.

Islands and islets abound on the Caribbean coast. There are some 2400. In Campeche Mexico, Isla del Carmen is a barrier sandy island on the Usumacinta delta; but mostly associated with coral formations, as in Yucatan Peninsula Mexico (Cayo Arcas, Alacranes, Chinchorro, Isla Mujeres, and Cozumel), Belize (the Keys), Honduras (Bay Islands and Cochinis Keys), Nicaragua (Miskitos, Cisne, and Maiz Keys), and Panama (Bocas del Toro and San Blas Archipelago). In comparison, there are few islands on the Pacific coast except for Panama, with some 200. There is a small group of islands in the Gulf of Fonseca (including Meanguera, Meanguerita, Amapala, and El Tigre). A group of eight islands are found in Costa Rica's Gulf of Nicoya, with the Chira island and Murcielago island located to the north, and Isla del Caño 11 km in front of Peninsula de Osa. Coco Island, 500 km to the southwest of the continent, marks the most distant territorial point of the Central America region.

Physiographic variability is also present on the sea bottoms. The Mesoamerican Trench extends all along the Central

America Pacific and has a maximum depth of 6662 m. The Caribbean Cayman Trench has a maximum depth of 7680 m, with depths up to 2000 m off Belize.

The largest regional outcrops of ocean waters are found off the Pacific gulfs of Fonseca, Panama, and Papagayo. These are caused by the Caribbean's seasonal winds, which push water out to the sea causing outcrops of colder water richer in nutrients. This effect is also true in the Gulf of Tehuantepec in the southwestern Mexico caused by the seasonal winds coming from the Gulf of Mexico throughout the Tehuantepec Itsmo. As it is typical of tropical seas, the Caribbean's surface water is mixed very little with colder deep waters which are richer in nutrients, and as a result the open seas are low in primary productivity. The greatest wealth, in terms of Caribbean productivity, is associated with the presence of coral reefs, mangroves, and other important ecosystems (i.e., deltas) on which regional fishing depends.

Global Ecological Scenario: Synthesis

The physiographic, hydrological, climatic, physiochemical, bathymetric characteristic, and habitat heterogeneity, littoral currents, and extension of continental shelf, previously described determine productivity, as well as the quantity and distribution of Central America's coastal-marine resources. Likewise, this distribution has historically conditioned utilization of resources and their relation to socio-economic development in the region. Traditional numbers indicate that Central America possess 6603 km of coast, representing approximately 12% of the Latin America and Caribbean coast. These contain some 567,000 ha of mangroves, 1600 km of coral reef, and about 237,650 km of continental platform where multiple activities of economic and social importance are carried out (Windevoxhel et al. 1999). The region has potential for utilizing more than 1.1 million km² of exclusive economic zone (Table 1). Nevertheless, and following the "Mesoamerican" approach focused on in this entry, these numbers should be much more impressive considering the states of Chiapas, Campeche, Yucatan, and Quintana Roo in southern Mexico, as well as Colombia, and this is true for the length of the marine coast and its natural resources, shelf to 200 m depth, exclusive economic zone, urban population in coastal cities, average annual volume – metric tons – of goods loaded and unloaded as crude oil/gas products and dry cargo, tourism, and fish products (Yáñez-Arancibia 1999).

The Central America coast consists of numerous peninsulas, gulfs, and bays favoring a high degree of physiographic diversity. Extensive intertidal zones and well-developed coastal barriers encircle great coastal protected waters. On the Pacific, coastal cliffs are highly developed in Costa Rica, and partially developed in El Salvador, Nicaragua, The Gulf of Fonseca, Panama, and Colombia; while

Guatemala and Mexico have no coastal cliffs. On the Caribbean side, the coast tends to be quite flat and cliffs are nonexistent due to less drastic geological and geomorphologic processes, with some local exception in the Dulce River low basin in the Atlantic coast of Guatemala. There are very important mega-lagoon/estuarine systems in Mesoamerica, that is, Terminos Lagoon Mexico (Yáñez-Arancibia and Day 1988), Amatique Bay, Guatemala (Yáñez-Arancibia et al. 1999), Gulf of Fonseca, El Salvador/ Honduras/Nicaragua (Gierloff-Emden 1976), Gulf of Nicoya and Gulf Dulce Costa Rica (Voorhis et al. 1983; Vargas 1995; Lizano 1998), and the Panama Bay.

Climatic conditions vary latitudinal along the Pacific coast. A dry zone from northern Costa Rica to Guatemala experiences water shortage for at least five months; transition toward a system of greater moisture is presented in Guatemala; rainfall is extremely heavy in southwestern Costa Rica and Panama, where shortage occurs no more than one of two months (Jiménez 1999). The particular characteristics of the Caribbean coast determine differences between this area and the Pacific. For example, while tides on the Pacific reach up to 6 m (extreme), those on the Caribbean and the Atlantic of Mesoamerica coast are around 30 cm (extreme). Dominant winds on the Caribbean coast produce waves up to 3 m, higher than those on the Pacific. Pacific coast rivers are short and highly dynamic, and discharge significant volumes of sedimentation from May to November during the rainy season. Rivers on the Caribbean tend to be longer and discharge is greater and more stable as a result of topographical conditions and almost year-round precipitation, and in the case of the northern end of Mesoamerica (Pacific and Atlantic of Mexico) with a significant seasonal pulse in September–October.

Cross-References

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- [Coastal Climate](#)
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Mining of Coastal Materials

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Mining is the process of extracting rock and mineral material from near-surface sources of the earth. It may include small, local borrow pits with no supporting infrastructure or large-scale operations with permanent facilities for removing, loading, and transporting the resource. Considering only coastal areas, those adjacent to the shoreline and extending seaward through the breaker (surf) zone (U.S. Army Coastal Engineering Research Center 1966) and landward through the highest surfaces subject to modern processes of wave alteration, the area represented (based on estimates by Kuenen 1950) is a little less than 1% of the global surface. The coastal zone, as used here, includes all parts of barrier islands, lagoons, tidal flats, and mangrove swamps, but does not include embayments and fiords, permanently exposed parts of deltas, lands bordering estuaries (such as the Bay of Fundy), and other inland tide-influenced areas.

Energy conditions within the coastal zone, as evidenced by the land-forms, are very high. The zone typically encompasses a transition from terrestrial fluvial and eolian to marine processes, but in high latitudes may be transitional from glacial or periglacial to marine ice-shelf conditions. Everywhere coastal areas are subject to retreats and advances of the shore owing to deposition or erosion by oceanic currents, isostasy, and especially, change in sea level related to storage and melting of glacial ice. Recently, a relative rise in sea level has been augmented or induced in some low-lying coastal areas as a result of large-volume subsurface fluid withdrawal. The geologic history of these processes defines the changing positions of coastal zones relative to the rock and mineral resources stored in and supplied to them, and determines the volumes and concentrations of materials that may have economic value.

Most of the economically recoverable coastal-zone resources worldwide are sand and gravel deposits and heavy-mineral placers locally embedded in them. Occurrences of these resources are typically results of combined

rock-fragment transport by water and ice from terrestrial areas, reworking by both marine and eolian currents, and either sea-level stability or inundation by rising sea level. Conversely, where tectonics or marine-circulation patterns produce nondeposition of terrigenous clastics or erosion of rocky shores, limestone and building stone may be exposed and available for mining. Authigenic mineralization in near-shore areas, especially of phosphorite, may require specific conditions of upwelling of nutrient-rich seawater in relatively stable areas of continental margins.

Sand and Gravel

Deposits of sand and gravel are the most widespread and are probably the most economically important non-energy resource of global coastal zones (Cruikshank and Hess 1975; Williams 1986). Estimated global reserves of sand and gravel on continental margins are 2 orders of magnitude greater than those of terrestrial supplies; the proportion of these deposits that occur within coastal zones is uncertain but is no doubt substantial (Cruikshank 1974). Principal uses of the resource include aggregate in cement and asphalt for construction, road base, earth fill, beach restoration, and a variety of industrial products dependent on silica sand and other sand-sized minerals (Langer and Glanzman 1993). Increasingly during the latter twentieth century, deposits of coastal-zone sand and gravel near urban areas were mined as an alternative to the extraction and shipping of dwindling terrestrial supplies from greater distances. During this same period there was also a shift from onshore to offshore mining. Because sand with physical properties comparable to those of beaches is needed for beach replenishment, shallow offshore sand accumulations near the sand-deficient beaches, such as shoreface, tidal deltas, and shoals, have become preferred mining sites.

Movement of erosion products from land areas by late-Quaternary fluvial and glacial action, and the flooding of the deposits by elevated sea level related to deglaciation and eustatic change, are the main processes by which the enormous reserves of global sand and gravel have developed and been preserved (Williams 1986). Where wide continental shelves are tectonically inactive, these sand and gravel resources mostly remain concentrated in the nearshore depositional environments, generally being moved only short distances by marine currents and wave action.

Mining of coastal-zone sand and gravel only recently has become important on a global scale as fluvial sand and gravel deposits near coastal cities have been depleted. Harbingers of this trend especially have been urban areas of small-to medium-sized islands, where limited onshore supplies of fluvial sand and gravel have been nearly exhausted, leading to the dredging and use of coastal and offshore supplies. Since

1987, for example, an average of nearly 25 million tons of sand and gravel has been dredged annually from deposits off coastlines of Great Britain (Hitchcock et al. 1999), accounting for perhaps a fourth of the national demand. In Japan, about half of recent aggregate needs have been met by mining marine supplies, and almost all remaining sand and gravel reserves of Puerto Rico are offshore (Rodriguez 1994). Extending the estimates of Williams (1986), 20% or more of the sand and gravel needs of the Netherlands and Denmark are met by nearshore sources; adjacent to the New York metropolitan area about 1.5 million m³ of coastal-zone sand and gravel are extracted annually, in part as an aid to navigation.

Placer Minerals

Placers, heavy minerals (specific gravity >2.85) released by rock weathering and generally concentrated mechanically by fluvial or littoral processes, occur in coastal areas where the minerals are derived locally from bedrock or from rocks of inland drainage basins that supply detritus to the coastal zone. Following transport from the rock source to a coastal environment, placers may be concentrated by tides, waves, and nearshore currents wherever shore processes are neither strongly erosive nor depositional (Rajamanickam 2000). Thus, depending on provenance, the ability of water, ice, and wind to transport mineral particles varying in size and specific gravity, and fluctuations of sea level, coastal-zone placers occur as flooded fluvial and glacial deposits, disseminated beach deposits, or eluvial deposits (Garnett 2000a; Kudrass 2000; Yim 2000).

Fluvial placers include nodules of gold and cassiterite (SnO₂) that have been transported short distances from their sites of origin by high-energy streamflows and deposited among coarse alluvial deposits of stream channels. Although wave action leads to concentration of these high-density minerals, deposition typically occurs during periods of lowered sea level when fluvial processes moved the placers into what is now the coastal area. Coastal-zone deposits of gold have been mined in Alaska, the South Island of New Zealand, southern Argentina and Chile, and the Philippines (Garnett 2000a), and deposits of cassiterite have been extracted from coastal Indonesia, Malaysia, Burma, Thailand, China, Australia, and Cornwall, Great Britain (Yim 2000). Although of relatively low specific gravity (3.5), but having extreme hardness and thus resistance to abrasion, diamond also occurs as placers in flooded paleo-river courses of coastal (and deeper marine) zones, often having originated far inland (Garnett 2000b). Coastal diamond deposits are mined from Namibia and western South Africa; mining of diamonds off northern Australia, southern Kalimantan (Indonesia), and elsewhere generally has not proven profitable (Garnett 2000b).

Disseminated beach deposits, or beach placers (specific gravity 2.85 to about 5.5), are not sufficiently heavier than most sediment particles (specific gravity about 2.7) to permit concentration by processes of fluvial sorting. Instead, sand-sized fluvial deposits of these light heavy minerals, especially rutile (TiO_2), ilmenite (FeTiO_2), magnetite (Fe_3O_4), zircon (ZrSiO_4), garnet ($\text{Fe}_3\text{Al}_2(\text{SiO}_4)_3$), monazite ($[\text{Ce}, \text{Y}]\text{PO}_4$), sillimanite (Al_2SiO_5), and, to a lesser degree, diamond, become concentrated in surf zones through wave sorting during periods of stable or rising sea level (Kudrass 2000; Rajamanickam 2000). Coastal-zone deposits of these minerals have been mined in Australia (rutile, zircon, garnet, monazite), South Africa (diamond, ilmenite), New Zealand (magnetite), India (ilmenite, zircon, rutile, garnet, monazite, sillimanite), Brazil (zircon, monazite), and Florida (zircon).

Other than gold and cassiterite, which are very heavy and tend to remain near the host rocks from which they were derived, the principal eluvial mineral of economic importance is phosphorite, formed mainly of varieties of apatite ($\text{Ca}_5[\text{F}, \text{OH}](\text{PO}_4)_3$) and defined as having a P_2O_5 content of at least 5% (Riggs 1979). Unlike other heavy minerals, exploitable deposits of coastal-zone phosphorite can be formed authigenically by deposition of nutrient phosphorus during periods of low sedimentation in zones of upwelling submarine currents. The phosphorite becomes concentrated through microbial processes and later, if affected by sea-level change, further concentration of primary phosphate grains may occur by coastal-zone wave action (Riggs 1979; Burnett and Riggs 1990). Examples of economically minable phosphorite deposits occur along the coasts of Chile and Peru and in the Atlantic Coastal Plain of Florida and North Carolina (Riggs et al. 1991).

Limestone and Shell

The mining of coastal-zone limestone largely is limited to areas where terrestrial resources are otherwise unavailable within economical depth or transport distance, or are preferred owing to features such as shell content or purity. Examples are shallow-marine limestones bordering volcanic islands, coral and oyster reefs, and chalks and coquinas of nearly pure CaCO_3 . Along tectonically active coastlines where terrestrial sediment is poorly stored, carbonate rocks of any age may be exposed and available for extraction. In contrast, shallow-marine lime precipitates and organic carbonates, principally shells, accumulate mainly on stable continental platforms having virtually no detrital deposition. Exceptions are those lagoonal areas where shell concentrations are so large that the muddy matrix can be removed economically. Where shells of marine organisms are concentrated as unlithified calcareous clasts, the resource may be

mined as a source of lime for cement in addition to its aggregate properties (Cruickshank and Hess 1975). In the Bahamas, deposits of precipitated aragonite and aragonitic oolites are mined as a source of pure calcium carbonate. Although these sources of lime, measured in billions of tons along the Grand Bahamas Bank, for example, may be mined locally, the value of limestone and shell deposits extracted annually is small compared to those of sand and gravel and heavy minerals.

Building Stone and Crushed Rock

Coastlines of tectonically active areas, volcanic islands, and areas subject to isostatic rebound following glaciation typically have a narrow coastal zone bordered by bedrock. Where the mining of building stone and crushed rock for aggregate from inland sources is restricted owing to land use or environmental concerns, rock may be quarried at the landward edge of the coastal zone. Historically, these rock quarries have been small and the amount of rock extracted for building stone and aggregate has been of minor economic importance. Beginning about 1990, however, restricted access to available inland rock reserves in areas of high population density, such as western Europe, resulted in the establishment of superquarries at coastal exposures of granite, quartzite, gneiss, limestone, and sandstone. Examples of active superquarries, those yielding at least 5 million tons of rock annually, are in Scotland, Norway, and Mexico. Other coastal superquarries are planned for the Shetland Islands, the Western Isles of Scotland, Canada, Ireland, Norway, and Spain (Pearce 1994).

Evaporites

The mining of offshore evaporite deposits, particularly halite, gypsum, and anhydrite, is limited mostly to coasts of countries where inland supplies are not readily available and importation costs are prohibitive. Thick beds of evaporites are mined in several regions of the United States, for example, but the needs of the Philippines are satisfied partly by the extraction of about 0.6 million tons of marine salt annually (Lyday 1995). Similar quantities of salt are mined in Indonesia but the portion taken from coastal or marine environments is not specified (Kuo 1995). Particularly along coasts of arid lands, such as those of the Arabian Peninsula and the Caspian Sea, evaporation of runoff and near-surface ground water concentrates dissolved solids beneath sabkhas. Extensive resources of potassium and magnesium in these areas, both as brines and in bedded evaporites, are likely to be extracted from these areas in the early twenty-first century.

Potential Environmental Impacts

Mining is a human disturbance that can have adverse environmental consequences if steps are not taken to prevent alteration of physical processes and degradation of water quality. In some well-documented examples, coastal sand extraction clearly has lowered beach and dune elevations, which subsequently has caused increased storm flooding, washover, and beach erosion (Zack 1986; Nichols et al. 1987; Webb and Morton 1996). A well-documented example of damage due to reduced beach elevation, washover, and sea-cliff erosion occurred in the early 1900s at the fishing village of Hallsands, on the south Devon coast of England. The village was destroyed during intense storms following extensive dredging of tidal-zone sand and gravel (Worth 1923; Robinson 1961).

Where fluvial sand and gravel have been mined from stream channels entering the ocean or in lagoons, excavation pits filled with anoxic water may alter nearshore circulation patterns and degrade water quality. In other examples, subtidal sand extraction for beach nourishment has altered nearshore currents and wave refraction patterns, causing or accelerating erosion of the adjacent beaches (Combe and Soileau 1987; Rodriguez 1994). Diamond mining along the west coast of South Africa has disrupted the intertidal fauna, destroyed kelp beds, adversely affected lobster habitat, and vast areas of dunes have been destroyed to get to the underlying placer deposits. Controversy surrounds the possible adverse biological effects of hydraulic dredging in lagoons and other nearshore waters. Clearly, dredging of living reefs is deleterious, but the conclusions are less certain about the permanent damage to benthic organisms caused by increased concentrations of suspended sediment associated with offshore dredging (Blake et al. 1996).

In most countries, sand mining from beaches and dunes begins as small, local operations because the resource is abundant and extraction costs are low. These operations, now illegal in the United States, continue until government authorities recognize that the practices cause permanent environmental damage, and then policies are formulated to mitigate future damage from sand extraction. Depending on the government's ability to enforce the regulations, mining practices may continue illegally where the economy is depressed. Most countries with coastal rock resources suitable for super-quarry development are imposing standards for licensing designed to protect scenery and recreation, marine biota, and water and beach resources.

Cross-References

- Beach Nourishment
- Beach Processes

- Beach Sediment Characteristics
- Changing Sea Levels
- Environmental Quality
- Shelf Processes

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Modeling Platforms, Terraces, and Coastal Evolution

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Terraces and their associated platforms and sea cliffs are the wave-cut and wave-built features associated with the land-water interface of seacoasts and lake shores. Along ocean coasts, they are the primary signature of the stillstands in water level during the transgressions and regressions of Pleistocene and Holocene epochs. The mechanics of terracing are fundamental to understanding the evolution of today's coastlines with their platforms, sea cliffs, barriers, spits, and capes. Coastal evolution models must incorporate processes that treat sediment transport and deposition as well as the abrasion and cutting of bedrock formations. This can be accomplished by coupled models, one treating the mobile sediment and the other bedrock cutting.

Background

Early studies of terraces, platforms, and sea cliffs include the work of de Beaumont (1845), Cialdi (1866), Fisher (1866), and Gilbert (1885). Gilbert's study of the active topographic features along the shores of the Great Lakes, supplemented by the visually distinct lake levels of the Pleistocene fossil shores of Lake Bonneville in Utah, provided the most detailed insight into the formation of platforms and terraces. He describes the wave-quarried hard rock platforms as *wave-cut terraces* backed by sea cliffs and the depositional features comprised of littoral drift as *wave-built terraces*. He also studied the terrace relations to changing lake level. Emery (1960), Shepard (1963), and others used this nomenclature with *marine terrace* as a more general term for wave-cut and wave-built terraces along ocean coasts.

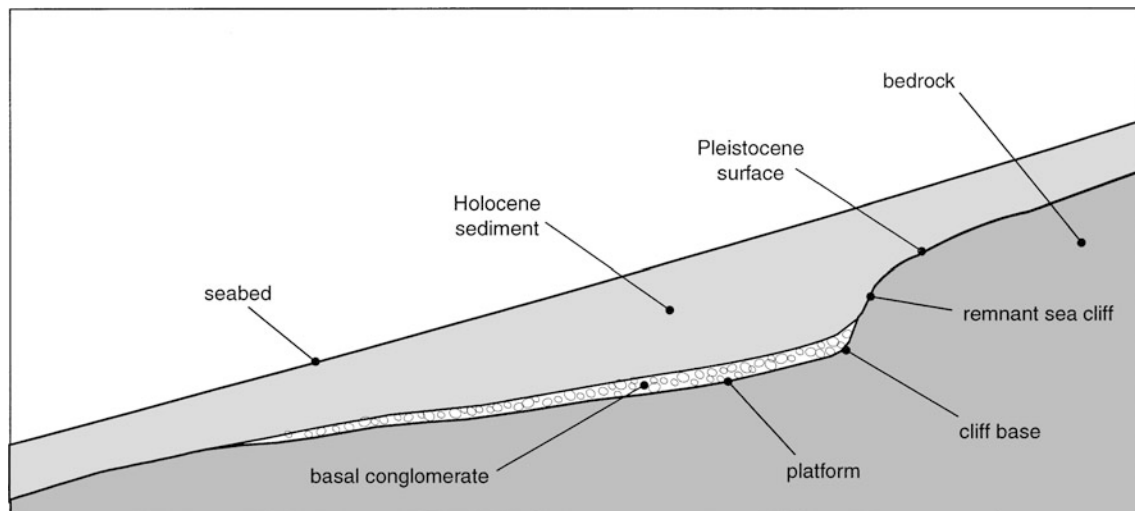
However, as pointed out by Trenhaile (1987), platforms may not be wave-cut but formed by other processes such as solution. He prefers *shore platform* as a more general term for rock surfaces of low gradient within or close to the intertidal zone. He uses the term *wave-cut terrace* to refer to the specific category of platform formed by waves (Trenhaile 2002). Sunamura (1992) classifies shore platforms developed during the present sea level as (1) sloping, (2) horizontal, and (3) plunging cliff. A more descriptive nomenclature for the latter two would be (2) step platform and (3) submerged platform. Generally, types (1) and (2) develop from cliff recession, with the step in platform (2) caused by differential erosion of rock strata. Plunging cliff platforms have sea cliffs that extend below the present water surface before joining a submerged platform. The submerged platform is a remnant feature from rapid sea-level rise and/or land subsidence. In what follows we will discuss numerical modeling of wave-cut terraces. These features consist of rock platforms backed by sea cliffs that were formed during the present stillstand in sea level as well as relic terraces now found on the continental shelf buried under Holocene sediment (Fig. 1).

The present configuration of coastlines and their associated terraces and sea cliffs retain vestiges of the previous landforms from which they have evolved. Coastal evolution is a Markovian process where the present coastal features are dependent on the landforms and processes that preceded them (Inman and Nordstrom 1971). This means that modeling coastal evolution must move forward in time from past known conditions and be evaluated by the present before proceeding to the future. Thus, paleocoastlines with their wave-cut terraces become time and space markers for modeling coastal evolution.

Numerical Modeling of Landforms

Physical models have provided insights and guidance to many of the processes leading to coastal evolution (e.g.,

Douglas L. Inman: deceased.



Modeling Platforms, Terraces, and Coastal Evolution, Fig. 1 Illustration of wave-cut terrace notched into the bedrock and now found on the shelf by seismic profiling below a cover of Holocene surface sediment

Inman 1983; Sunamura 1992). Generally, these efforts are limited by the uncertainties between laboratory experiments and the time and space scales of the landforms they represent. These uncertainties are circumvented by numerical models where the temporal and spatial scales of the landforms are applied to the laws governing geomorphology.

Numerical modeling of landform evolution is a rapidly expanding field driven by the need to understand the environmental consequences of climate change and sea-level rise. Numerical modeling has been enabled by the revolution in computational power, graphical representation, and ever-expanding digital databases of streamflow, wave climate, sediment flux, and landform topography (Inman and Masters 1994). Several two-dimensional models have been developed for the formation of wave-cut and wave-built terraces at various sea-levels. Storms et al. (2002) describe a process-response model for the development of barrier beaches during sea-level rise, and Trenhaile (2002) developed a model for the formation of rock platforms during changing sea level.

The Storms et al. (2002) model uses energy and mass flux balances to solve for incremental changes in the cross-shore profiles of mobile sediment in the Caspian Sea. On the other hand, the Trenhaile model uses a force-yield criterion to calculate the incremental erosion of steep, rocky submarine slopes. The latter model does not balance the budget of energy flux and, for certain selections of model parameters, requires a greater expenditure of energy in cutting rock than is available in the incident waves. This deficiency can be overcome by the energetics-based rock cutting model of Hancock and Anderson (2002). Developed for the formation of strath terraces in the Wind River valley during the Quaternary, the model includes sediment transport, vertical bedrock cutting that is limited by alluvial cover, and lateral valley-wall erosion.

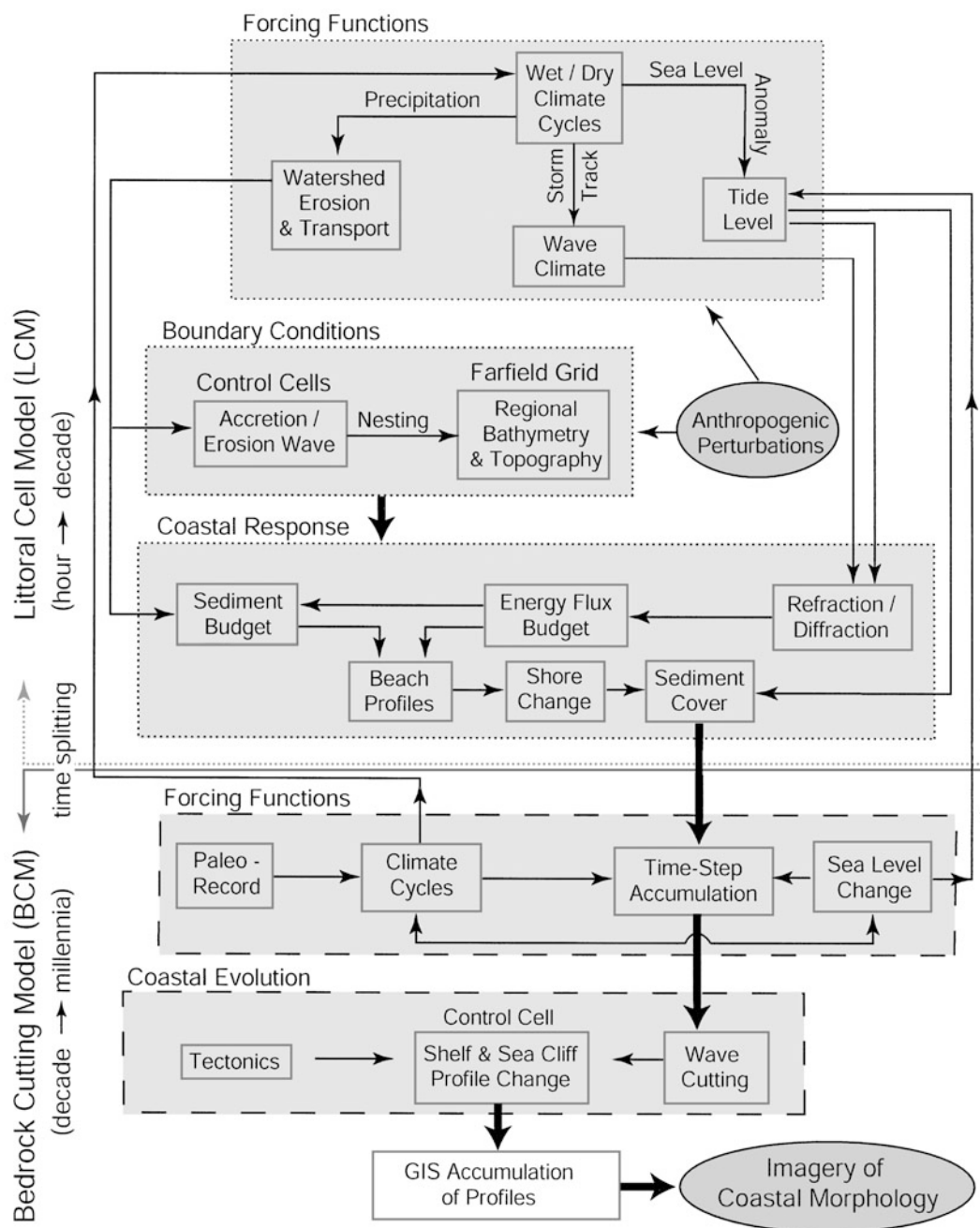
When reformulated for wave-forcing and sea-level change, their approach is applicable to wave-cut terraces.

Architecture of a Coastal Evolution Model

Here, we describe the broad outlines of a three-dimensional coastal evolution model developed under funding from the Kavli Institute (Inman et al. 2002). The model is functionally based on a geographic unit known as a littoral cell. A littoral cell is a coastal compartment that contains a complete cycle of sedimentation including sources, transport paths, and sinks. The universality of the littoral cell makes the model easily adaptable to other parts of the world by adjusting the boundary conditions of the model to cells characteristic of different coastal types (see entry on ► “Littoral Cells”).

The Coastal Evolution Model (Fig. 2) consists of a Littoral Cell Model (LCM) and a Bedrock Cutting Model (BCM), both coupled and operating in varying time and space domains determined by sea level and the coastal boundaries of the littoral cell at that particular time. The LCM accounts for erosion of uplands by rainfall and the transport of mobile sediment along the coast by waves and currents, while the BCM accounts for the erosion of bedrock by wave action in the absence of a sedimentary cover. During stillstands in sea level along rock coasts, the combined effect of bottom erosion under breaking waves and cliffing by wave runup carves the distinctive notch in the shelf rock of the wave-cut terrace (Fig. 1).

In both the LCM and BCM, the coastline of the littoral cell is divided into a series of coupled control cells (Fig. 3). Each control cell is a small computational unit of uniform geometry where a balance is obtained between shoreline change and the



Modeling Platforms, Terraces, and Coastal Evolution, Fig. 2 Architecture of the Coastal Evolution Model consisting of the LCM (above) and the BCM (below). Modules (shaded areas) are formed of coupled primitive process models (after Inman et al. 2002)

inputs and outputs of mass and momentum. The model sequentially integrates over the control cells in a downdrift direction so that the shoreline response of each cell is dependent on the exchanges of mass and momentum between cells, giving continuity of coastal form in the downdrift direction. Although the overall computational domain of the littoral cell remains constant throughout time, there is a different coastline position at each time step in sea level with similar sets of coupled control cells.

Time and space scales used for wave forcing and shoreline response (applied at 6 h intervals) and sea-level change (applied annually) are very different. To accommodate these different scales, the model uses multiple nesting in space and time, providing small length scales inside large, and short timescales repeated inside of long time scales.

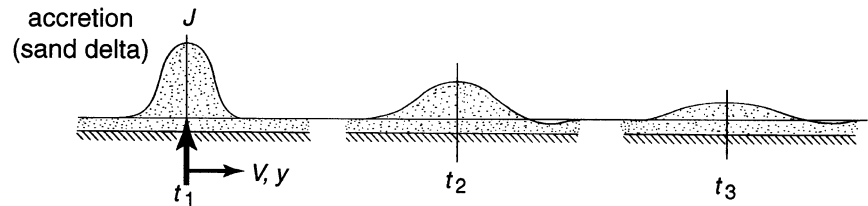
The LCM (Fig. 2, upper) has been used to predict the change in shore width and beach profile resulting from the longshore transport of sand by wave action where sand source

Modeling Platforms, Terraces, and Coastal Evolution,

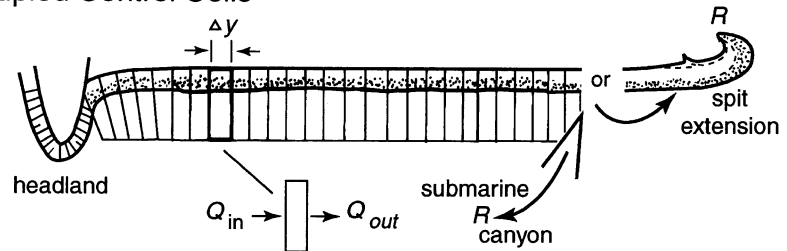
Fig. 3 Computational approach for modeling shoreline change (after Inman et al. 2002)

(a) Mass Balance

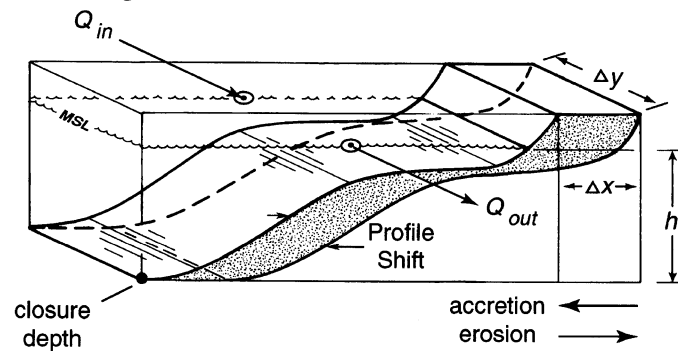
$$\underbrace{\frac{\partial Q}{\partial t}}_{\text{local time rate of change}} = \underbrace{\frac{\partial}{\partial y} \left(\epsilon \frac{\partial Q}{\partial y} \right)}_{\text{diffusion}} - \underbrace{v \frac{\partial Q}{\partial y}}_{\text{advection}} + \underbrace{J(t)}_{\text{river flux}} + \underbrace{R(t)}_{\text{spit / canyon sink}}$$



(b) Coupled Control Cells



(c) Profile Changes



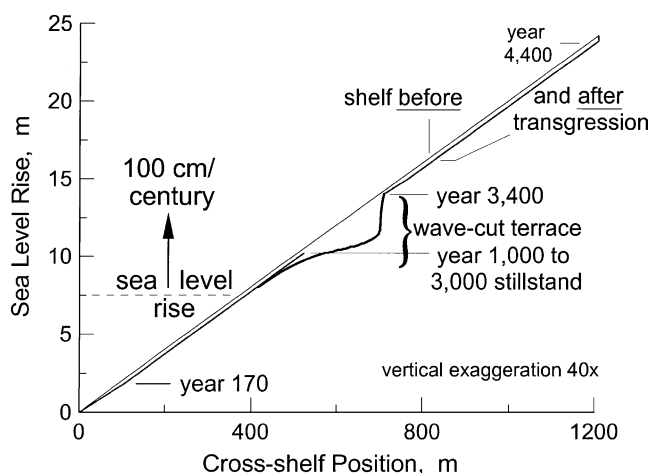
is from river runoff or from tidal exchange at inlets (e.g., Jenkins and Inman 1999). It has also been used to compute the sand-level change (farfield effect) in the prediction of mine burial (Inman and Jenkins 2002).

Bedrock Erosion

The BCM (Fig. 2, lower) models the erosion of bedrock by wave action during transgressions, regressions, and stillstands in sea level. Because bedrock cutting requires the near absence of a sediment cover, the boundary conditions for cutting are determined by the coupled mobile sediment model, LCM. When LCM indicates that the sediment cover is absent in a given area, then BCM kicks in and begins cutting. BCM cutting is powered by the wave climate input to LCM but applied only to areas where mobile sediment is absent. Time-splitting logic and feedback loops for climate

cycles and sea-level change are imbedded in LCM together with long runtime capability to give a numerically stable couple with the BCM.

Bedrock cutting involves the action of wave-energy flux to perform the work required to notch the country rock, abrade the platform, and remove the excavated material. Both abrasion and notching mechanisms are computed by wave-cutting algorithms. These algorithms provide general solutions for the recession of the shelf and sea cliff. The recession is a function of the amount of time that the incident energy flux exceeds certain threshold conditions. These conditions require sufficient wave-energy flux to remove the sediment cover, and a residual energy flux that exceeds the erodibility of the underlying bedrock. The erodibility is given separate functional dependence on wave height for platform abrasion and wave notching of the sea cliff.



Modeling Platforms, Terraces, and Coastal Evolution, Fig. 4 Example of BCM showing change in initial 2% shelf slope (thin line) due to wave cutting during a transgression/stillstand/transgression sequence (heavy solid and dashed line) (after Inman et al. 2002)

The erodibility for platform abrasion increases with the 1.6 power of the local shoaling wave and bore height, commensurate with the energy required to move the cobbles in the basal conglomerate that abrade the bedrock platform (Fig. 1). As a consequence, recession by abrasion is a maximum at the wave breakpoint and decreases both seaward and shoreward of that point. In contrast, the erodibility of the notching mechanism is a force-yield relation associated with the shock pressure of the wave bore striking the sea cliff (Bagnold 1939; Trenhaile 2002). The shock pressure is proportional to the runup velocity squared, and its field of application is limited by wave runup elevation. Wave pressure solutions (Havelock 1940) give a notching erodibility that increases with the square of the wave runup height above water level.

An example of terrace cutting by the BCM is shown in Fig. 4 where a constant sea-level rise of 100 cm/century over a continental shelf sloping 2% was interrupted by a 2,000-year stillstand. The wave cutting was driven by a two-decade continuous wave record reconstructed for the southern California shelf by wave monitoring (Inman et al. 2002). This data was looped 220 times to provide forcing over the 4,400-year long simulation. Inspection of the figure shows that the shelf slope receded about 15 m during the periods of rapid sea-level rise. During the 2,000-year stillstand, a wave-cut terrace was formed with about a 150 m wide wave-cut platform and a 3 m high remnant sea cliff. These dimensions are in approximate agreement with evidence of wave-cut terraces along the California coast (Inman et al. 2002). However, models of terrace cutting at paleo-sea levels will always require input of proxy wave climate appropriate for the location being modeled as well as the proper erodibility coefficients for the bedrock at that location (see entries on

“► Climate Patterns in the Coastal Zone”, and “► Energy and Sediment Budgets of the Global Coastal Zone”).

Cross-References

- Climate Patterns in the Coastal Zone
- Energy and Sediment Budgets of the Global Coastal Zone
- Littoral Cells
- Shore Platforms

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Modes and Patterns of Shoreline Change

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Definition

A mode of shoreline behavior is defined as a discrete pattern of shoreline movement identified by a recognition of a unique change in shape, a rate of movement, or a cycle of change.

Introduction

It is estimated that worldwide 70% of all beaches are eroding (Bird 1985); in the United States this percentage may approach 90% (Heinz Center 2000). Nearly every developed shoreline in the United States is retreating, and the coast is on a collision course with seaside development (Morton 2003). This problem is manifest in the United States where heavily developed barrier islands are experiencing sea-level rise and beach erosion. For example, Galgano (1998) demonstrated that more than 86% of US East Coast beaches are eroding, which is adversely affecting some 25 million people who live in areas now vulnerable to coastal flooding (FEMA 2008). With property values estimated at over \$3 trillion for the US East and Gulf Coast barriers, these policy decisions invariably have important economic consequences for coastal communities. Coastal activities, such as recreation and tourism along with transportation, offshore energy drilling, and resource extraction, and the fishing industry are integral to the nation's economy, generating 58% of the national gross domestic product (Moser et al. 2014). Consequently, a precise understanding of shoreline (defined as the high water line or wet-dry boundary) movements using accurate shoreline change models and geomorphic characteristics is paramount.

Managing beach erosion is a difficult challenge for government agencies. The costs of erosion are burdensome and are consuming an increasing percentage of Federal, state, and local budgets and may no longer be manageable using commonly accepted means (NRC 2010; Leatherman 2017). Furthermore, there is a paucity of long-term beach erosion data; and we still do not thoroughly understand how shorelines change over time. Shoreline change rates are typically based on data derived from only a few temporally spaced shoreline positions (Douglas et al. 1998). This problem is compounded because identifying trends (i.e., erosion or accretion)

demands forecasting future shoreline positions for effective coastal management, and simple forecast methods, such as endpoint rate and linear regression, are widely used. Yet, the matter of the error of forecasts has largely been neglected (Douglas and Crowell 2000). Consequently, policy decisions are regularly based on statistical modeling without a comprehensive understanding of patterns of shoreline change and the errors in computed change rates (Leatherman 1993). The methodology frequently employed is to determine shoreline change rates for an entire reach using simple statistical analyses of erosion.

Barrier islands are naturally dynamic landforms and inevitably change position and shape depending upon the changing relationship between coastal processes, the geologic framework, and human intervention. To date, a great deal of research has focused on understanding beach (profile) dynamics and defining shoreline (planform) change. Initially these studies focused on beach structure and beach-shoreface equilibrium models, generally over limited spatial and temporal scales – measured on the order of years and decades (e.g., Dean 1977; Bruun 1988). More recently, coastal geomorphologists have presented new theories for shoreline change and barrier island evolution over much longer time frames (e.g., Hayes 1979; Leatherman et al. 1982; Anders et al. 1990; List and Terwindt 1995; McBride et al. 1995; Sabatiera et al. 2009; Hapke et al. 2013).

Modes of Shoreline Change

It is generally accepted that there is geographic diversity in coastal landforms. This diversity is a result of the differences and combined effects of coastal processes, anthropogenic influences, and antecedent geology; shoreline change and coastal configuration are the integrative result. A “mode” of shoreline change is defined as a discrete pattern of shoreline movement identified through recognition of a unique change in shape, a rate of movement, or a cycle of change. For example, Morton (2003) illustrated the difference in shoreline change between bluff shorelines and those along sandy beaches. Bluffs retreat in a stair-step manner with periods of stability followed by rapid retreat associated with bluff failure. In contrast, sandy shoreline retreat is more erratic with interannual and decadal variations of eroding and accretion, although the shorelines ultimately move landward (Morton 2003; Galgano 2008). Kratzmann and Hapke (2012) illustrated the effects of beach nourishment and scraping on modes of shoreline behavior. Their Fire Island, NY, study indicated that from 1998 to 2007, developed beaches that employed beach scraping (i.e., using a bulldozer) and/or replenishment as erosion control measures experienced a measurable loss of volume, width, and dune elevation. Furthermore, areas affected by beach scraping as a form of

shoreline protection demonstrated distinct morphological differences in scraped areas relative to other areas of the beach. These areas also showed decreased beach width and volume and are more likely to experience overwash during storm events. Finally, the rapid erosion of sand from these beaches resulted in increased beach accretion downdrift (Kratzmann and Hapke 2012). These examples demonstrate natural and anthropogenically induced modes of behaviors that have to be understood to ensure the proper management of beach areas and coastal communities.

Modes of shoreline change may occur as two-dimensional (linear) or three-dimensional (alongshore) transformations operating over variable temporal and spatial scales, caused by the interaction of coastal processes within the constraints of geologic controls and sometimes with anthropogenic influence (e.g., inlet jetties). A mode of shoreline change can likewise depict a specific stage in an ordered sequence of shoreline development, whereby successive modes are dependent on preceding environmental factors and conditions for their existence (e.g., mesotidal barrier inlets). Finally, an examination of modes of shoreline behavior indicates that some beaches erode at much higher rates than adjacent beaches, sometimes as much as two to three times faster. These so-called erosion hotspots or erosion anomaly areas are caused by a variety of natural and anthropogenic factors. Kraus and Galgano (2001) defined and explained 18 different forms of erosion anomaly areas and classified them based on their duration, lateral extent, process–response factors, and predominant erosion mechanism (Kraus and Galgano 2001). Galgano (2007) explained, quantitatively, the persistent propagation of downdrift erosion adjacent to stabilized microtidal inlets. Modes of shoreline change identified to date are given in Table 1.

What relevance can be associated with the delineation and analysis of modes of shoreline change? Understanding modes of shoreline change provides a standardized geomorphic

framework upon which shoreline change data can be more accurately interpreted. Purely statistical methods of shoreline change analysis (e.g., the “black box” approach) often yield misleading results. Defining modes of change based on geomorphic principles and hence understanding the reasons for shoreline change yield a practical management tool. For many years, coastal scientists have been mapping historical shoreline positions to calculate rates of change (Heinz Center 2000). Nonetheless, traditional uses of these data (e.g., statistical prediction of future shoreline positions for building setback codes) have been rather confined and flawed (Douglas et al. 1998). Determination of an average shoreline rate of change and an understanding of the temporal and spatial variability associated with this change can be useful only if the causes of the change are recognized and understood.

Wave Versus Tide Energy

Processes that operate along the coast give rise to changes in barrier island size, shape, orientation, and position. Price (1953) first illustrated how wave and tidal energy were primary controls on shoreline processes and beach planform. Price (1956) later suggested that the bottom profile and geologic structure of the nearshore have important influences on beach morphology based on an analysis of low-energy coastlines in the Gulf of Mexico. He concluded that the most important geomorphic control on depositional coasts is the type and magnitude of hydrologic energy in the region. Davies (1964) illustrated that the two most significant hydrologic factors are wave energy and tidal current energy, both influenced by tidal range.

Hayes (1975) was the first to interpret the combined influence of wave energy and tidal range (TR) on barrier island morphology on a global scale. Thus, Hayes' (1975) work and later research by Nummendal et al. (1977) resulted in the acceptance of morphogenetic nomenclature of tide-dominated, wave-dominated, and mixed energy, which became widely accepted classifications for barrier systems (McBride et al. 1995). Hayes (1979) demonstrated how tidal range influences barrier island morphology because the effectiveness of wave action is attenuated, and tidal current activity is increased as vertical tidal range increases. Hence, in microtidal (i.e., $TR < 2$ m), wave-dominated conditions, barrier islands are long and narrow and are characterized by only a few ephemeral inlets. In contrast, short, stubby “drumstick” barrier islands are typically present in mesotidal (i.e., $TR = 2–4$ m) conditions. In this case, barrier morphology is linked to the presence of relatively stable inlets. Large ebb–tidal deltas are associated with these inlets and are common on mesotidal coastlines. The interaction of these large ebb–tidal deltas with wave energy combines to play a

Modes and Patterns of Shoreline Change, Table 1 Proposed shoreline classification types based on modes of shoreline change (After Leatherman 1993)

Type	Classification
1	Simple linear retreat
2	Alternating erosion and accretion
3	Progressive erosion alongshore
4	Storm-punctuated erosion
5	Inlet-induced erosion
6	Cyclic shoreline change
7	Mesotidal barrier behavior
8	Apparent island rotation
9	Pre-Holocene controlled configuration
10	Progressive erosion temporally
11	Cape-like features
12	Spit elongation

significant role in shaping the coastal configuration and morphology of adjacent barrier islands by accumulating and storing large quantities of sand-sized sediment which becomes available to the island on occasion and by inducing littoral current reversals driven by wave refraction. The end result is the characteristic “drumstick” shape (Hayes, 1979).

Davis and Hayes (1984) concluded that the “standard” definitions of wave-dominated (microtidal) and tide-dominated (mesotidal) coasts are based solely on the analysis of moderate wave–energy environments and are not adequate to explain the numerous exceptions to these broad classifications. Hayes’ (1975, 1979) classification is based largely on coasts with moderate wave energy. In their analysis of the Gulf of Mexico shoreline types, Davis and Hayes (1984) explained that tidal prism (the volume of water that must be exchanged through a tidal inlet) represents one of the more important but frequently discounted factors in determining the morphology of barrier islands. In the diurnal Gulf of Mexico, there is no relationship between tidal range and tidal prism because of the large lagoons and bays. Nonetheless, large tidal prisms, especially in areas of low wave energy, can explain large, well-developed ebb–tidal deltas and the characteristic drumstick shape. They suggest that a continuum between processes exists, and it is therefore possible to identify three types of coasts: (1) wave dominated, (2) tide dominated, and (3) one balanced between tides and waves.

Cape Development

The importance of wave energy and tidal currents to the morphology of coastal landforms has been used to explain the development and morphology of cape-like features (Davies 1964). Over the past three decades, the genesis and morphology of cape-like features has led to a great deal of study and scientific speculation. Most early models attributed their formation to accretional processes through the action of waves and currents, eddies from large ocean currents, and geologic controls. Hoyt and Henry (1971) observed that the popular models of cape formation based on modern processes (e.g., wave refraction, sediment transport, and deposition) did not explain the association of capes with major rivers and their similarity to ancestral capes in the region. Hoyt and Henry (1971) rejected the premise that waves and currents transport sufficient sediment out of adjacent embayments because it is not consistent with wave refraction and the concentration of wave energy on promontories. Their analysis of the Carolina capes suggests that erosion is the predominant process shaping the modern coastal configuration. The Carolina capes are most likely the result of the retreat and reworking of ancestral Pleistocene river deltas during Holocene sea-level rise. Moslow and Heron (1981) supported the

theory that the modern coastal configuration of these well-known capes can be attributed to the erosion and gradual landward retreat of formally larger antecedent capes.

Finkelstein (1983) proposed a model for the development of sedimentary capes on straight shorelines through development of a large recurved spit at a littoral barrier. Using Fishing Point, Virginia, at the southern end of Assateague Island as an example, Finkelstein (1983) demonstrated that cape-like features might occur where an inlet traps a large volume of littoral sediment. In the case of “cape” Assateague, a large volume of longshore sediment ($115,000 \text{ m}^3/\text{yr.}$) is trapped updrift of Chincoteague Inlet. This cape has developed since 1859 and has a pronounced influence on the shoreline by sand starving the barrier islands to the south (Rice and Leatherman 1983). The impoundment of longshore sediment in the cape has starved the downdrift beaches to the extent that a 5 km offset is observed on the islands to the south, and erosion rates of $10+ \text{ m/yr.}$ are observed for tens of kilometers to the south. The net result is a highly concave shoreline extending 35 km south of the cape, which is termed an “arc of erosion” (Leatherman 1993).

Dolan et al. (1979) hypothesized that capes will develop along the Virginia barrier island chain. In their analysis, the authors postulated that the pronounced growth of the bulbous updrift ends of drumstick barrier islands were an indication of cape development. Unfortunately, they did not recognize the cyclic change of accretion and erosion of the bulbous ends of drumstick barrier islands that operate on timescales of decades to centuries. Hence, their prediction of cape development was flawed. Leatherman et al. (1982) conducted a geomorphic analysis of long-term (i.e., $>100 \text{ yr.}$) shoreline change maps and geomorphic data to explain the coastal configuration and varied patterns of retreat of the Virginia barrier islands. They showed how sediment supply, controlled by the net southerly littoral drift, and the influence of inlets played a critical role in individual island adjustments. Antecedent Pleistocene topography beneath the barrier islands influences inlet position and stability. Differential subsidence and littoral sediment starvation have triggered a more rapid landward migration in the northern group of islands. Finally, wave refraction over the inner continental shelf has induced wave focusing on the southern two groups of barrier islands, causing more changes in orientation than the northern group. The presumed capes are in reality drumstick barrier islands behaving as described by Hayes (1979), and the recorded changes in the coastal configuration of the Virginia barrier islands suggest that a semi-linear shoreline configuration (e.g., opposite of capes) can be anticipated in the future (Leatherman et al. 1982).

There are important changes in cape morphology over shorter temporal time frames that explain their mode of behavior. Bernstein (2002) examined the morphological evolution of Capes Hatteras, Lookout, and Fear, North Carolina,

which typify a cusped foreland coastline. The evolution and morphology of the subaerial cape were examined through a field-intensive study using real-time kinematic global positioning system (RTK-GPS). Their study suggests that variability in shoreline position and morphology increases with distance from the landward end of the cape to the seaward tip; and the seaward tip of the subaerial cape point responds uniquely to changes in the nearshore wind and wave energy. This indicates that this segment of the cape tip plays a key role in sediment exchange between the subaerial cape and cape-associated shoal. Bernstein's (2002) study suggests that given the unique behavior of the seaward portion and transitional area of these capes, a previously undescribed sequence of morphologic events plays a dominant role in the transfer of sand from the tip of the subaerial cape point offshore to the adjacent shoals.

Tidal Inlets

Tidal inlets represent the most important feature along a sandy coastline in relationship to littoral processes. The preeminent role of inlets in the cyclic nature of barrier island morphodynamics was introduced by Hayes (1975, 1979). Dean and Walton (1975) indicated that tidal inlets represented the largest sediment sink of beach sand along the coast and are believed to be responsible for much of the beach erosion in Florida. Hayes (1975) pointed out that microtidal inlets are more dynamic than their mesotidal counterparts. Mesotidal inlets are more stable and are strongly associated with Pleistocene drainage channels. Halsey's (1979) research along the Delmarva Peninsula strongly suggested that inlet channels are correlated with relict thalwegs (the deepest part of a stream channel) of ancient streams. Furthermore, the higher interfluvies become the loci or "nexus" for barrier island formation.

The critical role played by inlets, natural and jettied, in controlling the morphology and orientation of shorelines is well documented. There is growing research dedicated to determining the spatial extent of the downdrift effect of inlets (FitzGerald and Hayes 1980; Leatherman et al. 1987; Galgano 2007; Seminack and McBride 2016; Styles et al. 2016; Nienhuis and Ashton 2016). The critical role played by inlets is their cumulative influence on the regional orientation and composition of coastal reaches.

Tidal inlets are dynamic features and can dramatically affect the morphology of adjacent barrier shorelines (Hayes 1975). The processes of inlet migration can bring about erosion and deposition to flanking barrier islands. For example, Nienhuis and Ashton (2016) applied a mass balance framework of sediment fluxes around inlets that includes along-shore sediment bypassing and flood-tidal delta deposition. Their model demonstrates that tidal inlets can be net sink of up to 150% of the littoral sediment and demonstrates how,

with flood-tidal deltas acting as a littoral sediment sink, migrating tidal inlets can drive erosion of the downdrift barrier beach. This arc of erosion is possibly the most dramatic influence of tidal inlets on adjacent barrier islands. Finkelstein (1983) illustrated how Chincoteague Inlet served as a littoral barrier and caused an arc of erosion extending 35 km downdrift. Wave refraction associated with large ebb-tidal deltas has been shown to cause reversals in the longshore transport system, resulting in onshore or alongshore sediment transport and ultimately unique changes in beach planform (Hayes 1979).

The degree to which inlets modify adjacent shorelines is a function of their tidal range. In most instances, tidal range correlates well with the size of tidal inlets. South Carolina offers an excellent example of the role of tidal inlets on coastal orientation and morphology. FitzGerald et al. (1978) illustrated how the number of inlets along the South Carolina coast is correlated to tidal range. Along the northern segment of the South Carolina coastline, the mean tidal range is 1.5 m. In this area, tidal inlets comprise less than 2% of the total shoreline. In contrast, the southern segment of this coast has a tidal range of 2.2–2.5 m. Accordingly, the amount of tidal inlet shoreline increases to 25%. Since the processes related to ebb-tidal deltas can directly control the morphology of adjacent barrier islands, tidal inlets will have a greater influence on shoreline morphology along barrier island coastlines with increased tidal ranges (FitzGerald et al. 1978). Nummendal et al. (1977) noted that the erosional and accretional nature of South Carolina's barrier islands is intimately connected with the change of tidal inlets. Increasing tidal ranges south of Cape Romain dictate larger and more numerous tidal inlets and therefore increased seaward transport of littoral sediments. FitzGerald et al. (1978) proposed that the three primary types of shoreline change in South Carolina were associated with three types of tidal inlets.

Brown (1977) recognized three distinct geomorphic zones along the South Carolina coast. The northern segment, or the arcuate strand, is relatively stable through time and is characterized by few inlets. A cusped delta occupies the central portion of the coastline. In the south where there are numerous inlets, barrier islands with highly variable shoreline change rates predominate (Brown 1977). Anders et al. (1990) showed in their analysis of shoreline movements in South Carolina that the greatest variability of shoreline change rates was in the vicinity of coastal inlets. These findings were supported by Hubbard et al. (1977) who pointed out that the arcuate strand beaches of northern South Carolina are stable everywhere except in the vicinity of tidal inlets. Only where there are 15–20 km between inlets does one escape their influence on the shoreline. These observations were supported on a more quantitative basis by Galgano (1998), who demonstrated that tidal inlets control shoreline change on 70% of the beaches on the mid-Atlantic coastline.

The pervasive influence of inlets on barrier island morphology was further demonstrated by Seminack and McBride (2016). They conducted a historical analysis of Assateague Island, which has experienced many breaching events throughout its history. They used historical documents, aerial imagery, and LIDAR data, of former tidal inlets of Assateague Island to produce an updated inlet chronology of the mixed-energy, wave-dominated barrier island. They identified eleven former tidal inlets, and breach zones were identified from north to south. Their analysis suggests that, in total, 34% of Assateague Island is estimated to contain tidal inlet fill (Seminack and McBride 2016).

Leatherman (1984) demonstrated the dramatic influence of jettied inlets on the long-term evolution of a barrier island. In a natural system, it is generally accepted that the long-term evolution of a barrier island will follow a general model of landward migration through time. In this scenario, the entire system (e.g., the barrier, bay, and mainland shoreline) will migrate as a geomorphic unit and retain the same general configuration. Leatherman (1984) provided an example where the system has been altered to the extent that large-scale evolution of a coastal unit is fundamentally changed. Ocean City Inlet was stabilized by jetties in 1934/1935. These jetties had a very rapid and pronounced influence on the morphology of northern Assateague Island. The jetties interrupted a longshore sediment flow of $140,000 \text{ m}^3/\text{yr}$. Further, the ebb-tidal delta captured an estimated $6,000,000 \text{ m}^3$ of sand, thus completely depriving the downdrift beaches of sand and creating an arc of erosion that extends some 10 km downdrift. Northern Assateague Island, which eroded at an historical rate of 0.6 m/yr. prior to inlet stabilization, reached rates averaging 11.5 m/yr. (Leatherman 1984).

Leatherman (1984) indicated that the implications of this dramatic increase in erosion will have important consequences on the long-term evolution of northern Assateague Island, Maryland. The northern segment of the island will not migrate landward as an entity as some predict in accordance with accepted models of shoreline change. If sea level alone were the driving influence behind this migration, this assumption would be essentially correct, but sea-level rise represents only a small percentage of the erosion potential and the bay is shrinking in width over time. Leatherman (1984) theorized that there would be a loss of a section of the barrier island with the opening of the mainland to ocean waves in the foreseeable future, similar to what occurred at Nauset Beach, Chatham, Massachusetts (Leatherman 1984).

Leatherman et al. (1987) modeled the erosion of northern Assateague Island to quantitatively determine its future movement. The authors proposed that the arc of erosion will continue to migrate southward through time, and this pattern of shoreline evolution can be mathematically modeled. The northernmost 10 km of the island currently represents the arc of erosion. There is no terminal end to this arc in a temporal

sense. In other words, the arc of erosion will continue to migrate southward through time. Furthermore, the area of maximum erosion will migrate downdrift so that the erosion rate will represent a non-steady-state condition. Leatherman et al. (1987) determined that because of the island's low elevation and narrow width, a storm will cause inlet breaching by the year 2020 with loss of the northern portion of the island as the sand is welded onto the mainland (unless artificial beach nourishment is undertaken).

Galgano's (2007) analysis of stabilized microtidal inlets along the East Coast supports the assessment by Leatherman et al. (1987). Galgano (2007) determined that the spatial extent of the arc of erosion downdrift from a stabilized tidal inlet is site specific and highly correlated to the age of the inlet and the volume of net longshore transport. The arc of erosion has no downdrift termination in a temporal sense, and the length of the arc expands at a nonlinear rate while conserving the volume of the area of change over time. The arc of erosion exhibits an explicit temporal evolution or mode of behavior. Finally, the data suggest that there is a predisposition for inlet breaching at selected locations.

Barrier Migration

Tidal inlets have been ascribed an increasingly important role in barrier island migration. Most existing barrier islands are believed to have originated 6000–7000 years ago and have migrated to their modern positions in response to postglacial sea-level rise. A number of theories describing the mechanisms for barrier island migration have emerged. Perhaps, the most widely accepted theory is the concept of continuous, albeit intermittent, barrier island migration by shoreface retreat (Leatherman 1983). Research by Kraft (1971), Kraft et al. (1975), and Belknap and Kraft (1985) using extensive core samples and radiometric data demonstrated how barrier islands migrate landward and upward on the continental shelf during periods of sea-level rise. Their data, based on carbon dating of peat samples from coastal Delaware, suggests that our present-day barrier islands originated 7000 to 8000 years ago (Kraft 1971; Kraft et al. 1975; Belknap and Kraft 1985). Ramsey and Baxter (1996) analyzed 231 geologic samples from the offshore, coastal, and upland areas of Delaware and suggest that modern barrier islands may have originated at about 4–6000 years before present. Their findings are supported by Barnhardt (2009) who suggests that modern barrier islands along the eastern United States were formed during a slowdown in Holocene sea level rise that occurred 6000 years before present. The landward migration or “roll-over” of barrier islands in a transgressive environment is a central issue in the study of modes of shoreline change.

Early researchers (e.g., Godfrey and Godfrey 1973) held that overwash was the most important agent in landward

migration. Field data supported what was intuitively obvious – that wave-driven overwash sediments were transported across the island, causing bayside accretion. Hence, landward migration was accomplished by a steady progression of beach erosion and bayside accretion. McGowan and Scott (1975) supported this hypothesis suggesting that overwash events transported considerable volumes of sand to the landward side of the island. Therefore, in a condition of dynamic equilibrium, the barrier island profile does not change its size or shape, but instead maintains its volume and migrates landward with equal magnitudes of seaside erosion and bayside accretion.

Later, work by Armon (1979) in Canada and Leatherman (1979) in Maryland countered that inlets played the principal role in barrier island migration. Using extensive on-site surveys of overwash fans on Assateague Island, Leatherman (1979) showed that overwash processes were most important for building the vertical elevation of the barrier island. But, overwash processes do not transport sufficient volumes of sediment to the bayside to maintain barrier island width. Barrier island width is maintained by inlet processes through the creation of large flood tidal deltas. Relict flood deltas serve as the platform onto which overwash and aeolian sediments accumulate.

Leatherman and Zaremba (1986) researched the long-term evolution of Nauset Spit, Cape Cod, Massachusetts, to determine the relative influence of inlet and overwash processes on barrier island migration. The lateral growth of the spit is largely controlled by inlet dynamics as inlets migrate down-drift (south), whereas overwash during major storm events caused substantial widening of this very narrow barrier spit in local areas. Overwash is the primary means of transporting sediment across barriers backed by glacial headlands or extensive salt marshes because the establishment of an inlet channel is precluded. Overwash in these circumstances tends to be the most effective barrier migration process through interaction with aeolian transport and dune-building processes in the subaerial zone. This vertical accretion permits the barrier island to maintain its elevation during landward retreat.

Time Frames for Barrier Migration

While the mechanisms and geometry for barrier island migration and modes of shoreline change are intuitively well understood, the time frames decidedly are not. Much speculation exists in the literature, and some envision that barrier island migration occurs extremely rapidly on the order of 50–100 years (Dean 1999). The implied assumption in this argument is that the barrier will conserve its mass and move as a unit. Others imply that in some situations, barrier islands are overtaken by sea level and drown in place (e.g., Rampino and Sanders 1980). This barrier island drowning model has never

been observed or supported by data (Panageotou and Leatherman 1986). Recent studies (Schwab et al. 2000) substantiate the findings of Panageotou and Leatherman (1986) that the 18 m isobath off Fire Island, New York, is not a drowned barrier island.

Time frames for shoreline movement are a subject of much debate, and the literature is replete with examples of shoreline movements at all timescales. Moody (1964) showed how a single, large storm event could displace the shoreline by a hundred meters in a matter of a few days, but a continuum exists between short-term storm events to long-term evolution (e.g., order of centuries). In addition, a very high degree of post-storm recovery has been observed, even after great storms (Morton et al. 1994; Douglas et al. 1998).

The prevailing theme in geomorphic literature is that barrier islands are continuously migrating landward in response to sea-level rise. In doing so, they maintain their mass through equal amounts of seaside erosion and bayside deposition (from overwash events). However, quantitative studies at Fire Island, New York, and Hatteras Island, North Carolina, indicated that both sides of the islands are eroding rather than migrating (Leatherman 1987). Therefore, beach erosion does not necessarily signify barrier island migration. Migration denotes a change in the barrier island's centroid; this is not accomplished by erosion alone. Barrier island migration is a time-averaged phenomenon. This migration is not a continuous process, but instead is caused by episodic and site-specific, storm-generated events over longer time frames. Recent studies of salt marsh loss in the Virginia barrier islands by Deaton et al. (2016) suggest that the primary barrier-backbarrier coupling is the loss of marsh and tidal flat area because of barrier island migration; and thus, in this region, the barrier islands appear to be “slimming” down as a precursor to migration.

Leatherman (1987) presented geomorphic and shoreline change map data that indicated that many barrier islands are thinning by bayside and oceanside erosion to some critical barrier width. Once the barrier island has achieved this critical width, migration by inlet and overwash processes will become effective in barrier island translocation. McBride et al. (1995) observed this slimming-down process in their study of long-term barrier island behavior in the Gulf of Mexico. They concluded that when shore position is monitored over time, shoreline change could be quantified and classified into a number of geomorphic response types as a function of scale. These studies indicate the need for the longest possible period of record of shoreline position data in order to make the correct geomorphic interpretation.

McBride et al. (1995) conducted a quantitative analysis of historical shoreline position over a spatial scale of 10–100 km. Their analysis provides a scientific basis for documenting process–response relationships that shape regional coastal morphodynamics. Even though “megascale”

Modes and Patterns of Shoreline Change, Table 2 Megascale geomorphic response types (After McBride et al. 1995)

Classification	Response type
Simple	Lateral movement
	Advance
	Dynamic equilibrium
	Retreat
Complex	In-place narrowing
	Landward rollover
	Breakup
	Rotational instability

shoreline change studies are typically undersampled spatially, this type of data is essential for formulating realistic research and management strategies regarding form-process relationships. The authors identified eight geomorphic response types to classify barrier coasts on a “megascala,” that is, coastal reaches of 10–100 km at time frames of decades to centuries (Table 2). The authors inferred that sea-level rise is one of the major factors that controls the occurrence of these geomorphic response types along the barrier chains in Georgia and Louisiana. In regions where sea-level rise occurs at lower rates, sediment supply appears to be the dominant factor.

Antecedent Geology

Contemporary literature offers a growing number of works that assign increasing importance to the role of antecedent geology controlling coastal configurations and modes of shoreline change. Many early researchers understood that there were important geologic controls influencing coastal configurations. Price (1953) was perhaps the first to establish the important influence of antecedent topography on coastal landforms, but this research was hindered by the absence of quantitative and stratigraphic data.

Kraft (1971) conducted studies of Holocene sediments in coastal Delaware, illustrating that Holocene sediments are deposited over a complex Pleistocene unconformity. Large, modern depositional features, such as spits, baymouth barriers, and tidal deltas in Delaware, are in large measure related to the presence and relative relief of the subsurface Pleistocene topography. Kraft (1971) showed how relict drainage patterns influenced the location of modern tidal inlets and shoreface sediment types. Further, Pleistocene headlands served as focal points for barrier island formation. In similar research, Halsey (1979) suggested that inlet channels along the Delmarva Peninsula were strongly correlated to relict stream thalwegs, and interflaves have become the loci for barrier island formation.

Kraft et al. (1975) presented a detailed study of the geologic structure of the Delaware coast and the influence of

Pleistocene topography on modern barrier island change. Relict drainage systems, Pleistocene headlands (subsurface and subaerial), and relict barrier formations combined to create a complex system of “coastal anomalies” in what is an otherwise straight coastal configuration. These “coastal anomalies,” or bulges and reentrants along the coast, were quantified by detailed shoreline change mapping of the Delaware coast by Galgano (1989). This mapping indicated the spatial diversity of modes of shoreline change within a coastal compartment.

Davis and Kuhn (1985) conducted research along the western coast of Florida and determined that antecedent geology played a vital but very different role in controlling shoreline change. This segment of the Florida coast is atypical for a microtidal coast in that it is composed of numerous drumstick barrier islands with stable inlet channels and large ebb shoals. Core samples and seismic profiling revealed that pre-Pleistocene geologic formations were responsible for this somewhat uncharacteristic situation. The data revealed that a subtle but well-defined limestone ridge of Miocene age with a relief of approximately 1 m underlies the modern barrier islands. Further, shoreward of the barriers, the irregular limestone surface slopes gradually toward the Gulf of Mexico. Davis and Kuhn (1985) indicate that the limestone platform is instrumental in controlling the location of the barrier islands and attenuating wave energy.

Oertel et al. (1992) interpreted the function of antecedent topography in barrier lagoon development. Microtidal lagoons are relatively wide and open, with few marshes. Conversely, mesotidal lagoons are narrow with numerous marshes. The authors theorize that pre-Holocene topography is the key determining influence for this phenomenon. Oertel et al. (1992) suggested that lagoon floors along tide-dominated coasts reflect the antecedent topography, which is dominated by well-defined fluvial channels and interflaves. The interflaves produce shallow areas in the lagoon, which become colonized by marshes. In the microtidal setting, the floors of barrier lagoons are initially smooth after formation.

The major theme in much of this research is that passive margin coastlines, such as the US Atlantic coast, are significantly influenced by the geologic framework. Some of these barriers have been shown to be perched barriers, resting on a pre-Holocene structure that controls modern beach dynamics and morphology (Kraft et al. 1975). Along other segments of the Atlantic coast, bathymetric features on the inner shelf modify waves and currents, ultimately affecting patterns of erosion and deposition. Riggs et al. (1995) presented an appraisal of the influence on geologic structure on the North Carolina coast. Their research suggested that some shoreline features in North Carolina are controlled by pre-Holocene stratigraphic framework and beaches are perched on pre-Holocene sediments.

Statistical Modeling

One of the fundamental objectives of coastal geomorphology is to determine shoreline trends. To this end, coastal scientists have applied different statistical techniques to predict future shoreline positions. In the past several years, this practice has assumed increased significance because many coastal states are using historical shoreline change data to project shoreline position for application in land use policies, primarily in establishing building setback lines and insurance zones (Heinz Center 2000).

The selection of an appropriate statistical model for shoreline change is a matter of some contention because it is critical to the final result. This problem is amplified because shoreline position data are typically limited and spaced irregularly through time. The principal issue is which shoreline positions represent the actual trend (i.e., not post-storm or wintertime shoreline data), what is the minimum acceptable period of record, how far into the future can the trend be usefully predicted, and which technique will best extrapolate the actual trend? The fundamental consequence of statistical modeling, regardless of the relative complexity, is that it will lose its physical and practical meaning if the geomorphic setting is not considered (Leatherman 1993).

Fenster et al. (1993) proposed a new method, the minimum description length (MDL), for predicting shoreline change rates and forecasting future shoreline positions. This method selects the model (e.g., linear regression or polynomial) on the basis of the trend of the most recent data. It is predicated on the appearance of trend reversals within the data record and depends on the persistence of the data. The authors, however, neglected to disclose the quality of the prediction beyond a few years, nor did they adequately test the model against a spectrum of real data. Crowell et al. (1997) used the MDL to predict trends using a temporally rich set of sea-level rise data, replete with interannual variations, to assess the ability of the MDL to predict future trends against the established record of sea-level rise. That research demonstrated that the MDL is not a useful predictor of long-term trends. Crowell et al. (1997) concluded that in spite of interannual variability in the data, the most reliable long-term forecast of shoreline position will be made from linear regression using the longest possible time series. Finally, Galgano and Douglas (2000) illustrated the necessity to use more than 80 years of data to determine a shoreline change trend. At time frames less than 80 years, simple calculations of change rate can actually produce the wrong sign (i.e., accretion versus erosion) and will likely produce a two- or three-sigma error in the trend. Furthermore, Galgano and Douglas (2000) demonstrated that the incorporation of post-storm and winter shoreline position data violate the assumptions of the linear regression model when mixed with summer shoreline position data.

Conclusions

Beach erosion and changes in coastal configurations are complex physical processes encompassing a number of natural and human-induced factors. These alterations drive the evolution of coastal configurations that generally conform to observable patterns; these modes of shoreline change are defined as a discrete pattern of shoreline movement identified through recognition of a unique change in shape, a rate of movement, or a cycle of change. Natural conditions that drive these changes include such variables as sea-level rise, tidal variations, wave energy, sediment supply, antecedent geology, seasonal variations in wave energy, and the episodic influence of storms. Humans influence the spatial and temporal variability of shoreline change by building structures (e.g., groins and jetties), dredging, damming rivers, and beach nourishment projects. Spontaneous variations in natural processes and the effects of human activity on the coast drive spatial and temporal variations in shoreline change and the evolution of coastal landforms. Therefore, we should not anticipate that shoreline change rates and coastal configurations will remain uniform through time; instead, we should realize that accelerations and decelerations in rates of erosion/accretion and trend reversals could occur. Reliance on statistical methods of shoreline change analysis can often yield misleading results, whereas defining modes of change based on geomorphic principles and hence understanding the reasons for shoreline change provides a practical management tool.

Cross-References

- [Barrier Island Landforms](#)
- [Beach Erosion](#)
- [Beach Processes](#)
- [Coastline Changes](#)
- [Coasts, Coastlines, Shores, and Shorelines](#)
- [Erosion: Historical Analysis and Forecasting](#)
- [Littoral Cells](#)
- [Microtidal Coasts](#)
- [Shoreline and Coastal Terrain Mapping](#)
- [Tidal Inlets](#)
- [Tidal Prism](#)
- [Tide-Dominated Coasts](#)
- [Wave-Dominated Coasts](#)
- [Wave-Tide-Dominated Coasts](#)

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Monitoring Coastal Ecology

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Introduction: Human Action

Early human coastal development was probably in harmony with coastal processes, modifying the habitats rather than destroying them. As human populations have increased and land-use intensified, these habitats have come under greater pressure. Intensive agriculture, afforestation, and infrastructure development, including the mass tourism boom in many areas, have resulted in the destruction of natural areas especially along the shores of the Mediterranean (Doody 1995). The impact on the environment increased the perception that coastal areas are vulnerable. This has led to the remaining areas, especially those dominated by “natural” landscapes, to be considered particularly precious. National and international legislation seeks to prevent further damage and destruction in these “fragile” environments.

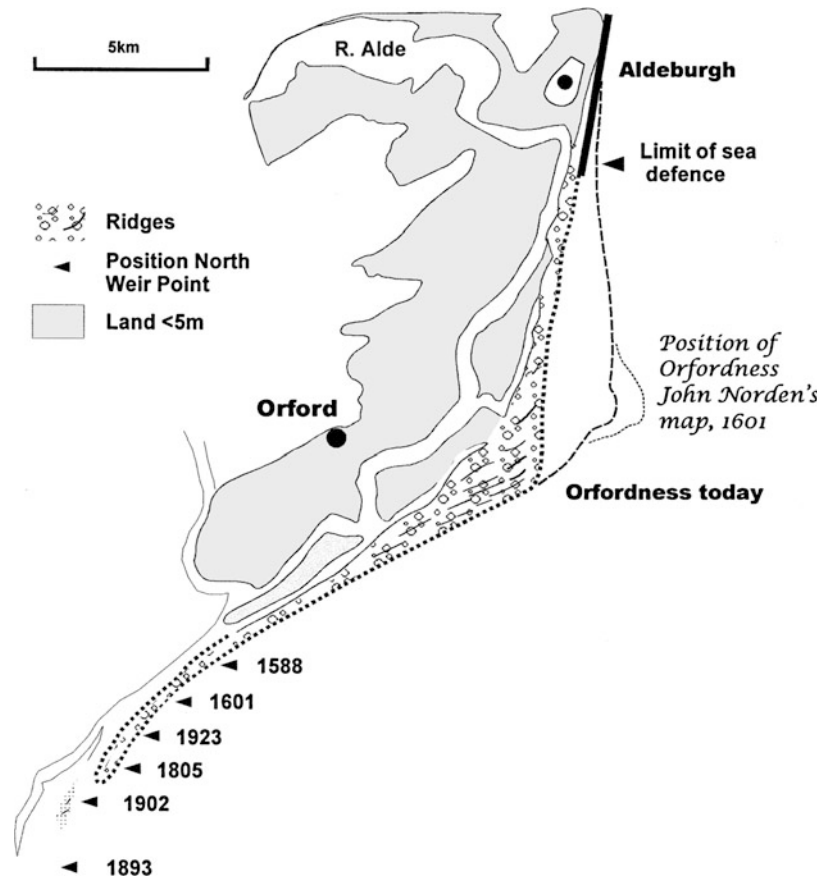
Faced with mounting pressure on coastal and marine resources, there is increasing concern about the ability of the coast to sustain the many uses to which it is put, including those involving socio-economic development. This has been brought into sharp focus with the recognition that global warming is a reality and that one of its consequences is a rising sea level. If sustainability of use is to be achieved then understanding coastal processes and human influence on them is a prerequisite for management, policy, and planning. The way in which this understanding is achieved lies in the survey of landforms, habitats and species, and monitoring or surveillance of change in relation to natural process and the impact of human activities.

An Historical Legacy

Knowledge of the status and distribution of landscapes, habitats, and species provide essential baselines against which to measure change. Some of the earlier historical maps are important in this context. While they do not provide detailed

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Fig. 1 Change along the Suffolk coast between 1530 and 1963 derived from maps in Anon, 1977. (Adapted from Doody 2001.)



information about habitats or species distribution, they can give great insights into long-term change. Early maps of the Suffolk Coast in eastern England around the time of Henry VIII, for example, show the configuration of the coast in the vicinity of Orford, then an important port (Anon 1977). When compared to more recent maps a picture of change can be built up, which gives an indication of the scale and time scale over which it occurs (Fig. 1).

This historical perspective is important to our understanding of the coast and coastal systems and the way in which they react to the impact of human actions over time scales longer than that of a single generation. In the United Kingdom, for example, there are a number of early texts which look at coastal issues from the point of view of the engineer and discuss these in terms of “Coast erosion and protection” (Wheeler 1903; Matthews 1934) and “Shoreline problems” (Carey and Oliver 1918), or in relation to the creation of new land from the sea (Du-Plat-Taylor 1931). Wheeler (1903) provides a series of geographical descriptions of the coast of England and short essays on northern France, Belgium, and Holland. A more systematic survey of the coastline of England and Wales was undertaken at the request of the Ministry of Town and Country Planning in 1943 (Steers 1946) which was later extended to Scotland (Steers 1973). These studies give a description of the geology, geomorphology, and geography of the coast and

helped form the basis for planning policy on the coastline of Great Britain in later years.

Though useful, this historical approach can only give a broad indication of change. It does not provide the more detailed information required for the management of individual sites and resources or unravel the many and often complex changes, which characterize the coast. For this targeted surveys are needed, together with a detailed research, surveillance and monitoring. These aspects are dealt with next.

Definitions

Survey

Surveying involves the collection of both quantitative and qualitative information about a feature or features using a standardized methodology, but where there is no pre-formed view on the likely findings. Thus, other than identifying what will be surveyed (a species, plant community, habitat, or human activity) the primary aim is to obtain information on the nature, scale, and location of the chosen subject.

Surveillance

Repeat surveys provide a means of identifying change over time, for example, in the status of habitats and species or the

rate of exploitation of a natural resource. As with the original survey there is no predetermined notion as to what change might be expected. The purpose of the new survey is to establish the nature of any change from the previous “norm.”

Monitoring

This implies the need to assess and understand the outcome of a particular course of action, such as the management of a habitat or species or the extent to which resource depletion is sustainable. This may relate to compliance with a standard or deviation from an acceptable state.

Survey

In order to assess the significance of a particular feature it is essential to know where it is, how widespread and how much of it exists. Survey is the first basic tool in assessing the value of a particular coastal feature and as an aid to its conservation and management.

Habitat Survey

Systematic surveys of landscapes, habitats, and species are a first stage in understanding the coast especially when considering its conservation. For example wide-scale habitat surveys are used to describe the nature conservation resource within a country as an aid to the selection of sites for statutory protection. In the mid-late 1980s in the United Kingdom, the nature conservation agencies (notably the Nature Conservancy Council) commissioned a series of coastal habitat surveys of Great Britain covering salt marshes (Burd 1988), sand dunes (Dargie 1993, 1995; Radley 1994) and shingle structures (Sneddon and Randall 1993). Attempts have been made to bring information together at a wider European scale but these are hampered by the variety of definitions and survey methodologies used and can only give a broad indication of the nature and scale of the resource (Dijkema 1984; Doody 1991). Site-specific surveys are often much more detailed and are used as a basis for developing management strategies such as those for the conservation of the flora and vegetation of the Wadden Sea islands, in the southern North Sea (Dijkema and Wolff 1983).

Species Recording

Establishing the location and population numbers of birds, (wintering and breeding) is also an essential prerequisite for the selection of sites to protect rare and specialized species. It is important to know where the majority of a population of a restricted species resides during critical periods of its life cycle. For example, most of the world population of the gannet *Morus bassanus* nest on the coast of Great Britain and Ireland. In the case of this species it is even more important for the protection of the species, to know that 90,000

breeding pairs (approximately 5% of the total population) occupy one small site, the Bass Rock in south east Scotland. Similar considerations apply to other vulnerable and specialist species such as marine turtles which need to come ashore to nest on sandy beaches. Knowing where these are, and the number of breeding individuals is an essential prerequisite for any conservation program. The results of habitats and species surveys can therefore provide:

- Summarized data from local to regional and national levels;
- Detailed maps showing distribution of vegetation, habitats, or breeding species;
- The basis for selection of important conservation areas;
- An assessment of the degree of conservation protection.

This information alone, tells us little about change, whether it is in terms of distribution, life-cycle movements, or population status. For this more frequent surveys are needed.

Identifying and Understanding Change: Species Studies

Managing the coast, whether for rare plants, specialist animals, important habitats, or human activities requires information on change and the causes of the change. If these show detrimental effects, either to the resources themselves, or knock-on effects to other interests, then it is essential to know what these are. By way of illustration, three elements can be discerned when monitoring individual species:

1. A species is rare or of special interest and managed to protect the population – are we being successful?
2. The species is invasive and we need to assess if our control measures are effective.
3. The species provides an indicator of change in the environment (Keddy 1991).

Each of these is considered by reference to three examples:

1. Bird migration studies.
2. The expansion of *Spartina anglica* into many salt marshes throughout the temperate regions of the world.
3. Population studies of sea birds.

Bird Migration

Determining whether a resident animal species is stable, declining, or increasing is relatively easy since recording breeding success or overall population numbers may be all that is required. However, unravelling the status of a migrant species is much more difficult. In this context, a systematic



Monitoring Coastal Ecology, Fig. 2 Migration patterns of the knot *Calidris canutus islandica* (see also British Trust for Ornithology, Migration Atlas Project). (Adapted from Doody 2001.)

approach has been adopted for monitoring bird populations involving the use of bird-ringing techniques. The original purpose of this was to unravel the mysteries of bird migration. However, knowing the pattern of movement is only part of the story. These studies also provide a vital tool for helping to determine the life history and population dynamics of many bird species. The information needed includes the distances traveled, the stopping off (refuelling) points, and the routes over which migration takes place, as well as studies of the bird's breeding success and physical condition. Armed with this information it is possible to begin developing conservation strategies for species which may breed in the Arctic, winter in estuaries in the southern North Sea or even as far away as South Africa (Fig. 2).

The importance of these studies is reflected in North America by the fact that the US Fish and Wildlife Service maintains a Division devoted to migratory bird management. Organizations undertaking such survey work, include the International Shorebird Survey which has been gathering standardized information on the numbers of shorebirds congregating at migratory stopover sites in the spring and fall since 1974. Standardizing and coordinating national studies using agreed practices for data collection and computerization are also recognized as part of the process. In North America, a "Migration Monitoring Network" was established in 1998 to coordinate activities including establishing standards and guidelines for the operation of monitoring programs. In Europe, in order to understand the significance of different parts of the species life cycle and

the influence of human and other factors upon them, a European Union for Bird Ringing coordinates continent-wide research projects. This work is also used to monitor bird populations against international conventions and is a prerequisite for effective protective measurements for the many declining bird species, especially in their more vulnerable locations.

In the United Kingdom, the WeBS (Wetland Bird Survey) Low Tide Counts are part of a joint scheme coordinated by the British Trust for Ornithology. In this case, counts are made on about 20 selected estuaries each winter (November–February) to determine the distribution of birds during low tide and to identify important feeding areas that may not be recognized during Core Counts which, on estuaries, are mostly made at high tide. In this context, British estuaries are known to be important wintering areas for several waders (including the knot) that breed in Arctic Canada and Greenland, and staging areas for those that breed in Siberia and winter in Africa. The pattern of bird ringing recoveries, combined with the knowledge of breeding success and observed changes in wintering numbers provide the basis for assessing the status of the species. Evidence from bird counts in estuaries in the southern North Sea suggest that knot numbers have recovered from losses in the 1980s, though only to about 70% of its numbers in the 1970s. The reasons why the numbers have not returned to their previous levels are not clear. Loss of intertidal habitat through land-claim and deterioration in the quality of habitat at its wintering sites are implicated, in the absence of major change in breeding success. This has important implications for policy in relation to land claim around the estuaries of the southern North Sea, where large wintering populations of this species occur (Doody 2001).

Spartina Invasion

The invasion of non-native species and their effect on a wide variety of habitats and species is a major conservation (and in places economic) concern. The precise pattern of change is often impossible to predict. For example, no one could have anticipated the impact of a chance meeting between two species originating on opposite sides of the Atlantic on salt marshes throughout the world. An American plant, *Spartina alterniflora* was brought to Southampton, in England in ships ballast in the early 1800s. Hybridization between this species and the British *Spartina maritima* prior to 1870 produced a sterile hybrid *Spartina* × *townsendii*. Further change resulted in a natural fertile amphidiploid *S. anglica* being formed (Hubbard and Stebbings 1967). At first the plant was heralded as an aid to land claim though latterly, as the new species spread in just a few decades to many estuaries in England and Wales, it began to cause concern to conservationists and others (Gray and Benham 1990).

Monitoring this change took place through a variety of ad hoc approaches including field survey (Hubbard and Stebbings 1967; Burd 1988) and more localized research studies, Langstone Harbour (Haynes 1986) and Poole Harbour (Gray and Pearson 1986). The overall picture, which emerges is of a highly invasive species, which has had a profound effect on salt marshes both in England and throughout the world. These studies were not coordinated, yet together they provide a fascinating account of the scale and impact of change. This example serves to emphasize that there is often no one approach to monitoring and that especially over the medium to long-term a variety of methodologies, if properly interpreted, can provide an understanding of what is happening. From this an assessment of policy and management responses is possible, which in the case of *Spartina* has resulted in major efforts to control its spread both in England and Wales, and in the United States (Wecker 1998).

Seabirds as Indicators of Change?

Once the status of a species has been determined, it is possible to relate any change, whether it is declining or expanding, to environmental or other human factors. These changes can be used not only as a means of assessing the population itself, but also as an indication of the state of other environmental conditions. For example, in the coastal environment the status of breeding seabirds might be used as an indication of the health of fish stocks. However, before such a causal link can be established, between a decline in a seabird population and a reduction in fish stocks other factors must be eliminated. These include:

- suitability of the nesting site (increases in grazing pressure may make burrows used by petrels and puffins more susceptible to collapse);
- incidence of predation (increases in predatory birds due to increased fishing effort and greater quantities of discarded offal causing death to young and/or reduced availability of food as parents are forced to disgorge before reaching their young);
- impact of offshore oil pollution and death of adult birds;
- changes in weather patterns.

Any or all of these could have an adverse effect on population size, which may hide the long-term implications of over-fishing. However, once these factors are recognized, and analyzed, change in animal populations can provide an important indicator of wider environmental damage. One of the special strengths of the use of bird population information, lies in the standardized methodology used for collecting data and the relatively long time frame over which the monitoring has been carried out.

Indicators for Integrated Coastal Zone Management

The above discussion shows how change in individual populations provides information, which is not only relevant to the conservation or control of the species being recorded, but can also be an indicator of the state of the wider environment. The examples given above involve quite simple interactions, though even these require detailed investigation to establish why change is occurring. When looking at the coast more widely, especially when socio-economic issues are included, assessing cause and effect, and hence policy implications can become highly complex and a real challenge. Making sense of the plethora of sometimes contradictory information, requires a systematic approach to its use. Although approaches differ in detail the basic sequence is the same and involves identifying the pressures which influence the state of the environment and the impact on human health and ecosystems. Society responds with various policy measures, such as regulations, information and taxes, or management action, which can be directed at any part of the system. The way in which this process develops and the factors important to the way in which it operates is illustrated next.

Taking Stock

Taking stock of information on the environment is the first stage in any strategy to identify key issues for policy formulation and management action. In the United Kingdom in 1984 the need for an assessment of the status of the North Sea and its resources was recognized. As part of this assessment “a comprehensive description of the coastal margin of the North Sea and Celtic Seas, their habitats, species and human activities” was initiated. The resulting documents (Doody et al. 1993; Barne et al. 1995–98) and associated electronic publications provide a description of the whole of the coastal and marine environment of the United Kingdom and Ireland, the human use to which it is put and policies and organizations responsible for management. The subjects covered were determined by the wide range of stakeholders involved in the process, who also helped to identify the key information requirements and issues on the basis of their own knowledge and understanding. The results gave an insight into what has happened and a baseline to assess the need for action and monitor the effectiveness of policy decisions and management.

State of the Coast Reporting

The scale and range of information available to facilitate the understanding of the coast and its policy and management requirements at national level also is large and complex. When more than one nation state is involved the situation becomes even more complex. Different approaches, methodologies, and languages across national boundaries increase the barriers to the collection and collation of information. However, despite

this State of the Environment reporting is used to address issues in geographical areas, including several, which cross national boundaries. A comprehensive review of the state of the Arctic has been prepared (Nilsson 1997). In Europe, in addition to the European Union's general State of the Environment Report the "Dobris Assessments," there are other reports covering the regional seas, for example, the "Mediterranean Sea: Environmental State and Pressures" (Izzo 1998). Few of these reports deal specifically with the coast most are either general State of the Coast reports or, as in the last case, are largely concerned with marine issues. However, as a methodology all have the characteristics of being comprehensive in their approach and including a variety of key stakeholders in their preparation.

Key Issues

Stock taking and/or State of the Coast Reports provide a means of identifying key issues. These are important for the identification of indicators used in predicting change and the efficacy of policy implementation and management action. The key issues also help to identify key pressures on the system; another important element in reporting on the state of the system. A list of the issues most relevant to the coastal zone is given above, derived from various sources (Table 1).

Understanding the Coast: Research

Research is vital to understanding the way in which coastal systems develop, especially in helping to unravel cause and effect when considering the impact of policy and management action. This is particularly important in such a dynamic environment where coastal processes are often key in determining the nature of the coast and its ability to sustain human use.

Research

There are extensive research programs throughout the world designed to provide the necessary understanding of the impact on and limits to the use of the coast and its resources. Two approaches will be used to illustrate some of the techniques employed to identify the nature and scale of change and the underlying processes which "drive" the "natural" coast, namely *Lacoast* (Land cover changes in Coastal zones) which uses satellite images to assess land-cover changes in coastal zones and LOIS (the Land Ocean Interaction Study) which is designed to gain an understanding of, and an ability to predict, the nature of environmental change in the coastal zone around the United Kingdom.

Lacoast. The aim of the Lacoast project is to assess quantitative changes of land cover/land use in European coastal zones during the last decades. It is being undertaken by the Agriculture and Regional Information Systems (ARIS) unit and the Space Applications Institute (SAI) at the European

Monitoring Coastal Ecology, Table 1 A list of key issues adapted from publications covering different levels of administration in the European Union

Key issues (not specifically coastal) derived at European level are:
Socio-economy
Transport
Agriculture
Energy
Tourism
Industry
Households
Air quality and ozone (stratospheric and tropospheric)
Climate change
Water stress
Eutrophication
Acidification
Noise
Biodiversity and landscapes
Exposure to chemicals
Health and environment
Land and soil degradation
Waste management (after Stanners and Bourdeau 1995)
In the North Sea key issues are
Heavy metal contamination
Artificial radionuclides
Oil
Shipping
Nutrients
Fisheries
Organic contaminants
Protection of habitats and species
More specific key issues identified for the coastline of England and Wales
Sea-level change and increased storminess
Quality of bathing waters and beaches
Loss of habitats and implications for biodiversity
Pollution by hazardous substances
Development pressures on the coast (after Environment Agency 1999)

Union's Joint Research Centre. The project covers the entire coastal zones of 10 European Member States and includes a 10 km wide strip of land bounded by the coastline. This project provides quantitative estimates of land cover and land use change in the coastal zone, focusing in particular, on those caused by human activities. Where changes are observed, an exercise to identify and interpret the factors responsible for change is undertaken. The results so far have provided data relevant to environmental status and policy formulation at national and European scales helping to:

- identify major change at national levels;
- identify hot spots of change that when combined with other information can show trends in development pressures; and

- show the nature of human activity and hence the key issues requiring a change of policy and/or management action.

LOIS. The Land–Ocean Interaction Study, a 6-year multi million pound project (1992–98) funded by the United Kingdom’s Natural Environment Research Council, involved more than 360 scientists from 11 institutes and 27 universities. It collected a vast amount of data and is a major research investment aimed at improving the understanding of the way in which coastal systems interact. As its title implies, it was designed primarily to help elucidate the relationships that exist between the exchange, movement, and storage of materials at the land–ocean boundary. Some 300 papers have been submitted to academic journals, possibly half the number that will finally be published, and a series of CD-ROMs produced (Natural Environment Research Council 1998). Thus, as a research enterprise it is deemed to have been very successful.

A review of the main findings of the study suggests that there are some important broad conclusions to be drawn from the work – conclusions that have a more general value to the process of integrated management. For example, new insights to nutrient budgets show that 90% of the fluxes occur during only 5% of the time, that is, “pulses” are very important in delivering nutrient loads to the sea. This finding is important to any monitoring program designed to measure changes in nutrient levels since it would be easy to miss these important events. The work has also helped establish the validity of technologies, such as remote sensing, in identifying vegetation patterns, thus providing the possibility for a more cost-effective methods of survey and monitoring. A better understanding of sediment budgets and their relationship to erosion in an estuarine system has also been established. This leads the way to interpreting land–sea-level change and the implications for erosion and flooding.

Conclusion

Determining the status of the coastline its landscapes, habitats, and species concentrations and human interactions is a product of a wide variety of methodologies. These range from anecdotal observations to sophisticated, targeted research programs using modern technologies. Two phases can be discerned in the process of developing an understanding of the impact of human use on the ability of the environment to continue to sustain that use. The first involves a “stock take” of existing information which can be summarized as including the following:

1. *Geological, geomorphology, climate and sea level* variables, which provide the driving forces for the development of coastal systems.
2. *Coastal habitats and species* information, which set a baseline for the selection of important conservation areas (e.g., the European Union Habitats Directive).
3. *Human uses and activities* including the location and scale of infrastructure and intensity of use.

This can be described a “passive” stage in the context of monitoring as it is concerned initially with the location and scale of the resource or activities which take place there, rather than attempting to identify change and the causes of change. Undertaking an audit of coastal information as described above will also help to identify gaps in information and hence the need for updated or new surveys, and to assess their scale and frequency. Defining key issues for policy formulation and establishing the key human uses requiring surveillance or monitoring, is the next important step. Armed with this analysis it should be possible to develop a more “active” interaction between the provision of information and decision making. This may be summarized under four headings:

1. *Monitoring* to help identify and assess impacts, determine action and give feedback on its effectiveness.
2. *Surveillance* to identify unforeseen change and where appropriate ensure compliance with agreed legislative or other control mechanisms.
3. *Prediction* of the possible outcome of policy decisions whether in reaction to an unforeseen incident (e.g., oil pollution) or “natural” event (e.g., sea-level rise).
4. *Assessing* the effectiveness of action (management) is the final stage if we are to learn from good and bad practice.

One key aspect of the coast is its unpredictability. Identifying change without knowing what that change might be (surveillance) is therefore potentially very difficult. However, retrospective comparisons of maps, published works, the use of satellites and research (e.g., into sediment fluxes) as has been shown above, all help understanding. Twice daily tidal cycles are predictable, storms and tidal surges are not. It is often these events which cause substantial change. Similarly, cliff erosion rarely takes place on a small and regular basis and none of these unpredictable events are easily monitored. However, detailed surveillance of individual habitats can be effective and help elucidate wider scale change.

Monitoring: Salt Marshes and Sea-Level Change

Salt marshes are important habitats and occur around the coast throughout the world. In addition to their nature conservation interest as habitats for specialist plants and animals, they also form part of the functioning system of coastal wetlands (in estuaries and tidal deltas). In some areas, salt marshes are

eroding and a number of factors have been implicated including pollution and sea-level rise. However, a question arises as to whether this erosion is part of the natural dynamic or a longer term, and ultimately more significant change, resulting from global sea level rise or other more local effects caused by human intervention. Clearly determining the appropriate policy response will depend on understanding the extent to which human action is responsible for the observed change.

Visual and anecdotal evidence from south east England suggested that the salt marsh resource had eroded at a considerable rate during the 1980s and early 1990s. This in turn appeared to be the reason why a number of sea-walls were being undermined with an increased risk of flooding to adjacent low-lying land (Fig. 3). In an area already experiencing a relative rise in sea level of 5.4 mm per year due to isostatic change and sea-level rise this suggested that threats to life and property were intensifying. Detailed field survey and comparison between sets of aerial photographs appeared to confirm anecdotal evidence and the salt marshes were assessed as eroding at an alarming rate (Burd 1991).

On the face of it sea-level rise appears to be the main cause of loss. However, even if it is the main agent, many other factors

come into play. Tides and tidal range, sediment availability, and the nature of the coast all influence the development of salt marsh. Local weather conditions including rainfall, discharge rates of rivers, and the state of the tide all contribute to the incidence and severity of flooding. Erosion rates depend on the strength of the feature being affected by the erosive force. Changes in sea level *per se* (eustasy) are the result of global forces associated with the atmospheric temperature whether due to human influences or not. In areas where the land level is stable or sinking (isostasy) there will be an inundation of the coast. Given all these factors it may be impossible to determine the precise reasons for the loss of salt marsh. Despite this, the fact remains that salt marshes are eroding and as a consequence sea-walls are being undermined and land threatened with flooding. Thus, the measurement of change in salt marsh provides a very powerful indication that an adverse change is taking place and one which needs to be addressed.

Satellites as a Means of Monitoring Coastal Change

Using satellites to look at the environment of the earth is an important part of the space program. Remote Sensing Satellites provide information on a variety of features including physical oceanography, polar science, and climate research. However, satellites are currently under-utilized as a source of surveillance and monitoring data for coastal zone management (Doody and Pamplin 1998). Part of the problem lies in the absence of any strong link between those concerned with the management and conservation of coastal areas and those interpreting satellite data or developing new missions. Examples of the former might be the “field ecologist” or conservation site manager, the latter the satellite “technocrat.”

The key point in relation to the use of satellites lies in the fact that more frequent observations can be made and that these can, despite the initial costs of the data acquisition, may be more cost-effective than carrying out detailed and time consuming field survey. Although not all coastal habitats lend themselves to this approach and while cliffs (Fig. 4) present special difficulties, flat expanses of tidal land (Fig. 5), sand dunes and shingle beaches, and structures do not.

Time scales for change are also an important factor. While satellites can cover large areas and record cumulative effects they are less good at measuring more rapid change. Until recently, a further impediment was the scale of resolution, which limited the level of detail that could be obtained. As the frequency of survey and resolution improves so will the ability of the imagery to provide more effective surveillance of the coastal zone. This will not negate the need for more traditional forms of survey and monitoring. Sloping landforms, detailed vegetation studies and measurements of river flows, sediment movement, tides, etc. will still require traditional approaches to



Monitoring Coastal Ecology, Fig. 3 Land below the 5 m contour potentially at risk from coastal flooding and over-topping by the sea in Great Britain. (Adapted from Doody 2001.)

Monitoring Coastal Ecology,
Fig. 4 Clifed landscapes require oblique photography to show features, the chalk cliffs of Beachy Head, southern England. (Original photograph JPD)



Monitoring Coastal Ecology,
Fig. 5 Tidal sandy flats are easily discerned from satellite images, Brancaster, North Norfolk, England. (Original photograph JPD)



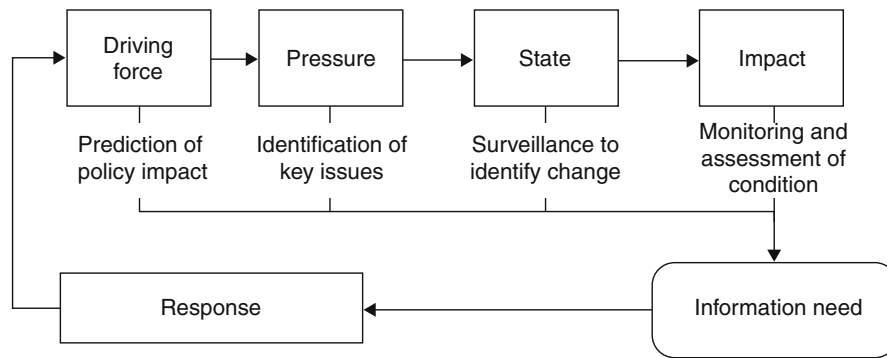
research, survey, and monitoring. The point is that satellites can and should be used more extensively for monitoring and surveillance of the coastal zone.

Indicators for Policy Response and Management Action

The importance of the coast to the economic and social fabric of society in many parts of the world, increases the need for more widely based monitoring, encompassing both

environmental and economic issues. Linking these into an assessment of the effectiveness of policy and action, add another dimension to the complexity of the monitoring system. Information on future trends on the state of the environment and prospects for socio-economic and sectoral aspects are crucial for determining progress against policy targets and to ascertain among other things:

- whether current policy measures are to be expected to deliver the required improvements taking into account trends in external factors;



Monitoring Coastal Ecology, Fig. 6 The European Environment Agencies modified United Nations' Pressure-State-Response model. This shows driving forces (e.g., industry and transport) producing pressures on the environment (e.g., polluting emissions), which then degrade

the State of the environment and have impacts on human health and ecosystems. Society responds with various policy measures, such as regulations, information and taxes, or management action which can be directed at any other part of the system

- whether additional policies might be considered as necessary to achieve the expected improvements;
- whether new policy needs are likely to emerge in uncovered areas.

At this stage, in the discussion monitoring has moved from being a concern of individual sectors or research academics, to being at the center of policy formulation and decision making. Developing indicators as an integral part of a strategy for policy formulation is a common approach (Fig. 6). Using a variety of techniques involving, existing information and new data derived from remotely sensed sources and traditional survey, monitoring and research, it is possible to establish an integrated approach which matches the needs of the coastal zone. For example, we should be able to measure salt marsh change, relate this to sea-level rise, and provide a means of assessing different options for sea defense according to economic and social criteria.

The use of remote sensing techniques will not provide information on bird migration routes or the population density and success rate of breeding birds. Nor will it replace survey and surveillance of narrow beaches or vertical cliffs, or detailed vegetation maps needed for site management. These forms of data acquisition will still be required, as will the view of the informed "specialist," or "local" knowledge. When linked to the more traditional forms of survey, monitoring and research, and other programs designed to synthesize and summarize existing data, an interactive information system which provides support for coastal management decisions can be devised. This will require a multi-disciplinary approach, if the demands for more integrated management of the coast and in the coastal zone are to be realized. The communication options afforded by the Internet also provide an opportunity to involve, and exchange information between the many relevant sectors operating on the coast. Achieving integration between the historical legacy of information, survey, surveillance, and

monitoring as well as research, and across the environmental and the socio-economic divide is the challenge for monitoring environmental and human wellbeing and hence the future of the coast.

Cross-References

- [Environmental Quality](#)
- [Europe, Coastal Ecology](#)
- [History, Coastal Ecology](#)
- [Monitoring Coastal Geomorphology](#)
- [Remote Sensing of Coastal Environments](#)
- [Remote Sensing: Wetlands Classification](#)
- [Salt Marsh](#)
- [Vegetated Coasts](#)

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Monitoring Coastal Geomorphology

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Coastal engineering and research, management of natural resources, beach and wetland restoration, navigation improvements, and military operations all share the need for copious amounts of data. These data typically are used to evaluate and monitor a specific reach of the coast. Ideally, a coastal monitoring program employs a multidiscipline approach in diagnosing the beach and nearshore zone. Many large programs such as the US Army Corps of Engineers' Shinnecock Inlet Study (Morang 1999; Militello and Kraus 2001; Pratt and Stauble 2001) or the Kings Bay Coastal and Estuarine Physical Monitoring and Evaluation Program (Kraus et al. 1994) collect data for a range of physical conditions such as wave climate, the morphology of the beach and nearshore surface, and accretion and erosion trends across and along the shoreline (high-tide line or coastline) under investigation. These collections allow scientists and engineers to understand the coastal processes and their variability in response to waves, currents, winds, and tides.

The high-energy, often hazardous coastal environment is one of the most difficult places on earth in which to collect data. Consider the enormous forces concentrated in the narrow coastal zone: in hours, beaches disappear; in days, new inlets are cut; in a generation, rock cliffs crumble. Even the most massive coastal works have often been buried in sand, swept away, or pounded into rubble. How are instruments expected to survive in this harsh environment? Despite the engineering challenges, scientists and engineers, who never feel that they have enough data, continue to develop innovative techniques and instruments to help them answer some of the elusive questions about the sea and the land it touches.

This entry summarizes the various types of *field* data that are typically collected for coastal research and engineering and briefly introduces some of the most common instruments. We exclude data developed in physical or numerical models, although clearly these are important tools also used by coastal

scientists. Biological monitoring is covered in a companion paper in this volume.

The quality of a field study depends on several factors:

1. Scientists must recognize the many problems, assumptions, and limitations in field data collection and make adjustments for them before attempting an interpretation. For example, a single cross-shore beach profile may not be representative of the beach topography during much of the year (e.g., possibly the survey was made after a storm when the beach was in an unusually eroded state).
2. The appropriate phenomena must be monitored to answer the engineering or scientific problem being investigated. For example, wave data are needed to choose appropriate stone size for a breakwater. But, an offshore ocean wave gauge may not provide useful wave statistics if the coast has complex topography with headlands and bays. A nearshore gauge is needed here.
3. Many of the techniques used to monitor processes or structures in the coastal zone are exceedingly complex. Inexperienced users should consult experts to prevent making important decisions based on poor data or data that do not truly answer their questions. Example 1: It is easy to purchase aerial photographs and interpret some sort of a shoreline (simply draw a line at the edge of the water). But what does this “shoreline” really mean, and can it be compared with shorelines identified from other dates that were based on different field techniques or criteria? Example 2: Offshore sand sources are being examined with geophysical techniques to determine if the sand is suitable for beach renourishment. Low-frequency acoustic systems may detect adequate quantities of material, but the

resolution is too low to reveal that the sand contains interbedded layers of coral fragments. Here, a combination of low- and high-frequency systems should have been used, combined with core samples. Example 3: A series of cross-shore beach profiles show that a significant amount of sand was lost from the subaerial beach after a storm battered a recently placed beach fill, leading an analyst to conclude that this sand is lost. However, the profiles were wading-depth only. If sled profiles that spanned the entire active zone had been collected, the analyst might have detected that much of the sand lost from the beach had accumulated in the offshore sand bars and that the total sand volume in the active zone was almost unchanged.

Classes of Monitoring Techniques

Coastal monitoring can be divided into three general classes of techniques, each of which use specific instruments and analysis methods (Table 1). Note that a comprehensive study might employ instruments and data from all three classes.

1. *Remote sensing methods*: instruments provide information about the land and the sea from a distance without being in physical contact (e.g., aerial photography, laser imaging).
2. *In situ instruments*: instruments are placed in the media being studied, such as current meters moored in the ocean.
3. *Sampling methods*: devices retrieve a sample of the material being examined (i.e., water, ice, sediment, biological material) so that the scientist can conduct more detailed examination in a laboratory.

Monitoring Coastal Geomorphology, Table 1 Classes of monitoring systems for coastal studies

Type	Characteristics	Examples	Advantages	Disadvantages
Remote sensing	Distant from the area or media being studied (without physical contact)	Air photographs Video monitoring Electromagnetic geophysics Acoustic geophysics	Broad coverage Cover large area in short time Suitable for hostile environments (military applications or harsh weather)	Interpretation is a model – must be verified with some field data or a priori knowledge. Resolution often too coarse for shoreline mapping or morphology studies Occasionally challenging or conflicting interpretation
In situ	Measures parameter from within the media	Current meters Wave gauges SBT (salinity bathythermographs) Strain gauges Tracking devices on turtles, whales	Often better resolution than remote systems Good temporal coverage Can sometimes be used to monitor changing phenomena not accessible remotely	Point source sampled over time – must interpolate spatially
Sample recovery	Retrieves a portion of the material	Plankton tows Coring devices Seafloor grab samplers Sediment traps	Allows detailed analysis in laboratory of the “real” material Re-analysis possible for different properties	Point data only – may not be representative spatially or temporally Preservation and archiving of samples is expensive

Remote Sensing and Geophysical Methods

Remote sensing methods include familiar and well-proven technologies such as aerial photography (first used in World War I), satellite imaging systems in the 1970s, and laser bathymetry in the 1990s. See Philipson (1997), Henderson and Lewis (1998), Lillesand and Kiefer (1999), and Rencz and Ryerson (1999) for manuals on various sensing technologies. All remote monitoring methods require measuring and recording some form of acoustic or electromagnetic energy and then relating the resulting data to specific earth parameters (Table 2).

The term “geophysics” is defined as the “study of the earth by quantitative physical methods” (Bates and Jackson 1984, p. 209). Geophysical methods are a form of remote sensing in that a researcher uses a tool to remotely image the seafloor or the strata below. The result is a depiction of the subsurface geology, a mathematical model based on varying acoustic impedances of air, water, sediment, and rock. The model, which must be interpreted, is based on numerous assumptions, and the user must always remember that the real earth may be very different than the model printed on paper or

displayed on his monitor. This warning notwithstanding, geophysical (particularly acoustic) methods, have proven to be extremely powerful tools in numerous coastal applications, including:

- Determining water depth (bathymetric or hydrographic surveys)
- Imaging the sea bottom to identify surficial sediments, measure bottom features such as ripples, and locate abandoned structures and hazardous debris
- Measuring the thickness of strata to locate suitable quantities of sand for beach renourishment
- Mapping gas pockets, rock outcrops, and other geological hazards
- Identifying coral, fish reefs, and other biologically sensitive areas
- Mapping cultural and archaeological resources such as sunken ships and pipelines.

Echo sounders or depth-sounders, side-scan sonar, and subbottom profilers are three classes of equipment commonly used to collect geophysical data in marine exploration. All three are acoustic systems that propagate sound pulses in the water and measure the lapsed time between the initiation of the pulse and the arrival of return signals reflected from target features on or beneath the seafloor. Other geophysical methods, such as magnetic, gravity, and electrical resistivity, laser line scan and electronic still cameras can be used in specialized engineering applications (Griffiths and King 1981; Reed 2001), but are not as common in reconnaissance coastal studies.

Until the twentieth century, measuring water depth consisted of the slow and laborious use of sounding poles and lead lines (Shalowitz 1962). The introduction of acoustic echo sounding after World War I revolutionized charting, and for the following 70 years, single-beam acoustic depth-sounders were used for most bathymetric surveys. In the United States, offshore waters are normally surveyed by the National Oceanic and Atmospheric Administration (NOAA) (Umbach 1976), while navigation channels, canals, and rivers are surveyed by the US Army Corps of Engineers (USACE 1994, Army Engineer Waterways Experiment Station (n.d.)). Most nations have standards that specify procedures to achieve required horizontal (positioning) and vertical (water depth) accuracies. The International Hydrographic Organization (IHO) has published universal standards for bathymetric surveys for the organization’s member states (IHO 1998). An exhaustive bibliography of hydrographic charting technology is listed in IHO (2001). Multi-beam echo sounders, becoming increasingly common now, are improvements on the traditional single-beam systems because they allow very detailed imaging of underwater structures and topography (Fig. 1) (Cowls 2000).

Monitoring Coastal Geomorphology, Table 2 Remote sensing systems

Energy type	Examples	Common applications
Electromagnetic	Radar, Synthetic Aperture Radar (SAR)	Ocean wave measurement Ground-penetrating radar Object identification (military)
	Laser	Terrestrial topography (LIDAR) Bathymetry (SHOALS)
	Multi-spectral	Satellite (LANDSAT, SPOT, AVHRR)
	Visual wavelengths	Aerial photography (features, terrain)
	Infrared	Aerial photography (plant life)
Acoustic	Low frequency (1 kHz)	Petroleum exploration, deep penetration Earth properties research, earthquake monitoring
	Medium frequency (0.5–7 kHz)	Seismic equipment (boomers, sparkers): near-surface strata, geohazards, sand resources, rock outcrops
	High frequency (12–450 kHz)	Bathymetry: sea bottom features, nautical charting Side-scan sonar: seafloor features, archaeology, shipwrecks, pipelines, breakwaters Multi-beam bathymetry: structures, breakwaters, detailed sea-bottom inspection



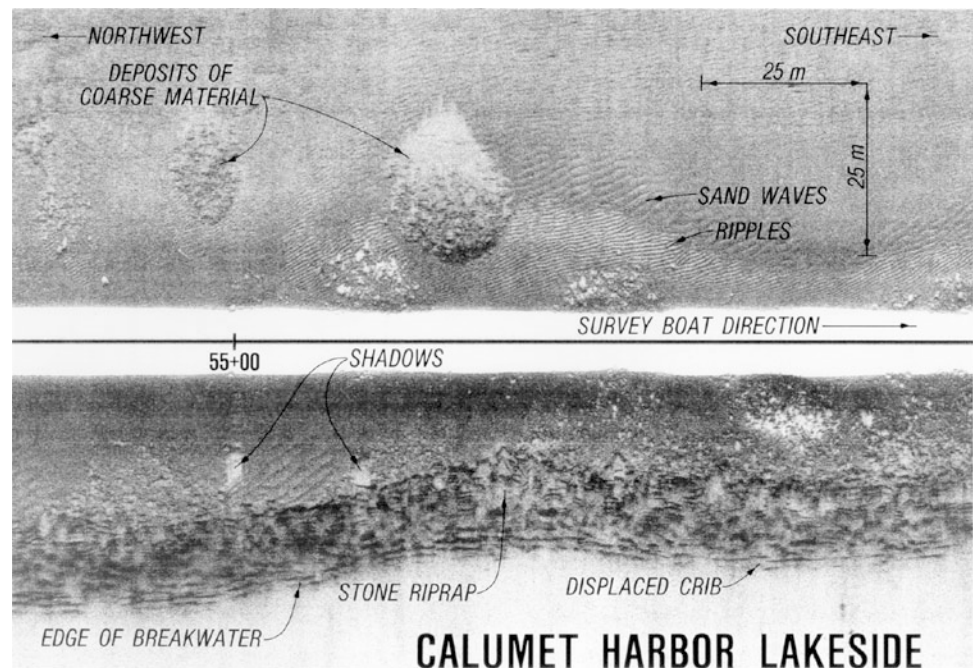
Monitoring Coastal Geomorphology, Fig. 1 Multi-beam acoustic data from Shinnecock Inlet, Long Island, New York, December 1998. Sand waves are evident in the inlet, and the highly irregular bottom near the west jetty indicates scour holes. North is to the top of the image. (Data collected and processed by State University of New York, Stony Brook)

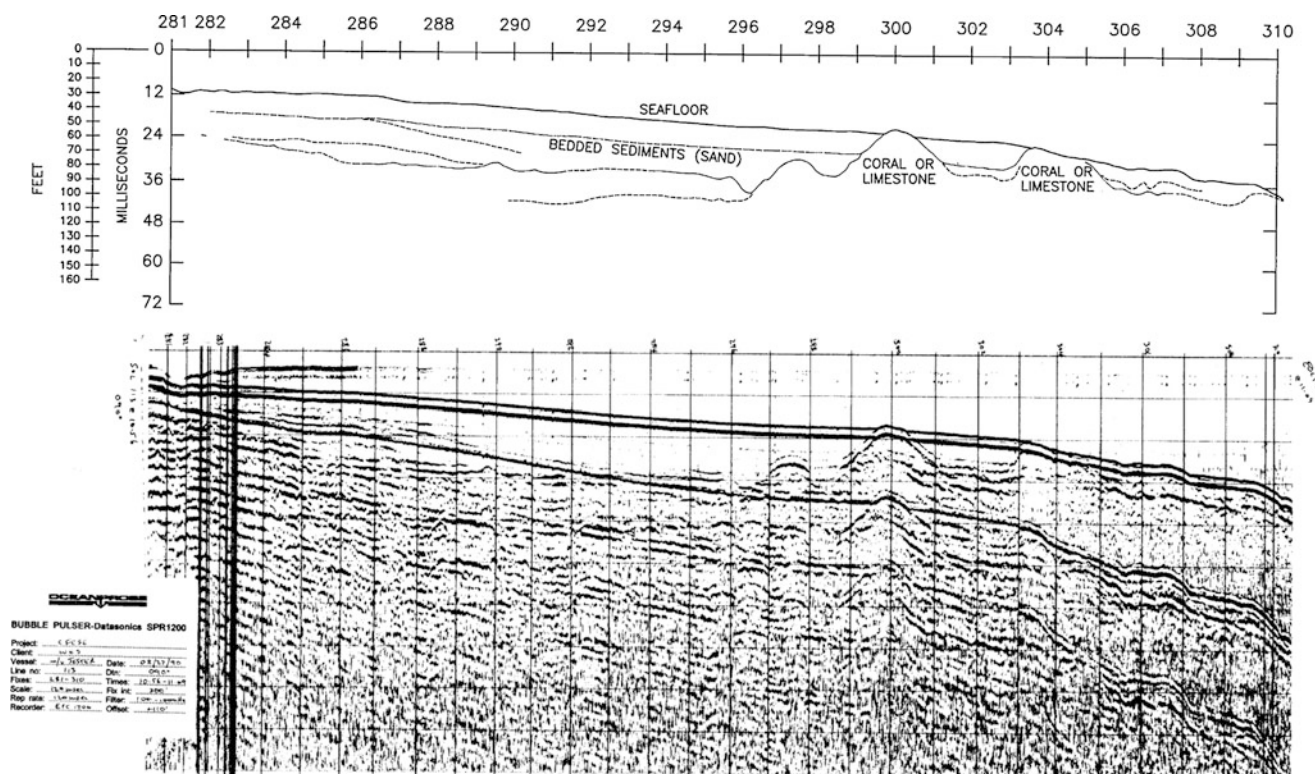
Side-scan sonar provides an image of the aerial distribution of sediment, surface bed forms, and large features such as shoals and channels. It can thus be helpful in mapping directions of sediment transport and areas of deposition or erosion. Belderson et al. (1972), Leenhardt (1974), Flemming (1976), Mazel (1985), and Fish and Carr (1990) provide additional details on the use and theory behind side-scan sonar. Skilled interpreters have long been able to use side-scan sonographs to visually identify seabed sediments, but this practice normally required some a priori knowledge of the survey area. New acoustic approaches have recently been developed that mathematically distinguish seafloor characteristics based on scattering parameters and waveform shapes (Whitehead and Cooper 2001). Side-scan sonar also is a powerful tool for offshore engineering and construction, where it is used to survey structures, inspect rock jetties and breakwater units, track pipelines and telephone cables, and identify hazards such as shipwrecks (Fig. 2). Up through the 1980s, most side-scan systems recorded on analog paper charts, but during the 1990s, modern computers have revolutionized the display of side-scan data, making possible real-time three-dimensional (3D) display of side-scan and multi-beam data (McAndrew 2001).

Subbottom profilers, as the name implies, are used to examine the stratigraphy below the seafloor. The principles of subbottom seismic profiling are fundamentally the same as those of acoustic depth-sounding, but subbottom acoustic transmitters and receivers employ lower frequency, higher power signals to penetrate the seafloor (Sheriff 1977; Sieck

Monitoring Coastal Geomorphology, Fig. 2

Annotated side-scan sonar record from Calumet harbor, southern Lake Michigan. The sonograph reveals a subtle displacement of some of the wood crib units and also shows how some construction material was deposited on the lake floor





Monitoring Coastal Geomorphology, Fig. 3 Analog subbottom profiler record from offshore Palm Beach County, Florida. Shore is to the left. The lump near Fix 300 is a coral or limestone outcrop. This is an

example of data valuable for locating offshore sand resources to use for beach renourishment

and Self 1977). “High-resolution” generally means that the surveys are intended for engineering purposes or for identifying strata and structures in the uppermost 50 or 60 m of the sediment column. Typical applications include reconnaissance geological surveys, foundation studies for offshore platforms, hazards surveys to locate buried debris and gas pockets, and surveys to identify mineral resources (sand for beach renourishment). Figure 3 is an example of subbottom profiling in a tropical area with coquina limestone reefs.

Ground-penetrating radar (GPR) uses electromagnetic energy to image subbottom sediments. Radiowave energy is transmitted through the sediment and reflects from materials as a function of variations in dielectric constants and electrical resistivity (Daniels 1989; Sellmann et al. 1992). The main limitation of GPR is that it must be used in fresh-water environments such as the Great Lakes or reservoirs, but FitzGerald et al. (1992) and van Heteren et al. (1994) have successfully used GPR to delineate structure and stratigraphy of beach ridges in New England.

A single geophysical method rarely provides enough information about subsurface conditions to be used without sediment samples or additional data from other geophysical methods. Each geophysical technique typically responds to different physical characteristics of earth materials, and correlation of data from several methods provides the most

meaningful results. Morang et al. (1997a) provide more details on geophysical systems and provide tables that list frequencies and resolution. *All geophysical methods rely heavily on experienced operators and analysts.* Inexperienced users should seek help both in contracting for surveys and in interpreting records.

Aerial photography is a traditional technology that continues to be very useful in monitoring coastal morphology, evaluating changes in wetlands and deltas, and mapping urban areas. Historical photographs, often dating to the 1930s, are invaluable in evaluating twentieth century changes to beaches and coasts. Figure 4 is an example of photographs taken 3 days after the Great New England Hurricane in 1938. These photographs document the opening of Shinnecock Inlet, Long Island, New York. Photogrammetry methods are covered in another entry in this volume.

Since the mid-1990s, Light Detection and Ranging (LIDAR) survey systems, based on lasers mounted in aircraft or helicopters, have been used around the world to obtain high-resolution topographic measurements. Most of these systems are for land use only, but one particular system, the Scanning Hydrographic Operational Airborne Lidar Survey (SHOALS), is now regularly used by the USACE to survey coastal waters and inlets. The system is based on the



Monitoring Coastal Geomorphology, Fig. 4 Vertical aerial photograph taken September 24, 1938 (3 days after the Great New England Hurricane crossed the south shore of Long Island, New York). This

image shows the new Shinnecock Inlet and numerous washover fans. Historical images like this are valuable in tracing the origin and behavior of coastal features (photographs from Beach Erosion Board Archives)

transmission and reflection of a pulsed coherent laser light from an aircraft or helicopter equipped with the SHOALS instrument pod and with data processing and navigation equipment (Lillycrop and Banic 1992; Estep et al. 1994; Irish et al. 2000; West et al. 2001). In operation, the SHOALS laser pulses 400 times per second and scans an arc across the aircraft flight path, producing a survey swath equal to about half of the aircraft altitude (usually 300–500 m). A strongly reflected light return is recorded from the water surface, followed closely by a weaker return from the seafloor. The difference in time of the returns corresponds to water depth. Data density can be adjusted by flying higher or lower, at different speeds, or by selecting different scan widths.

SHOALS has revolutionized hydrographic surveying in shallow water for several reasons:

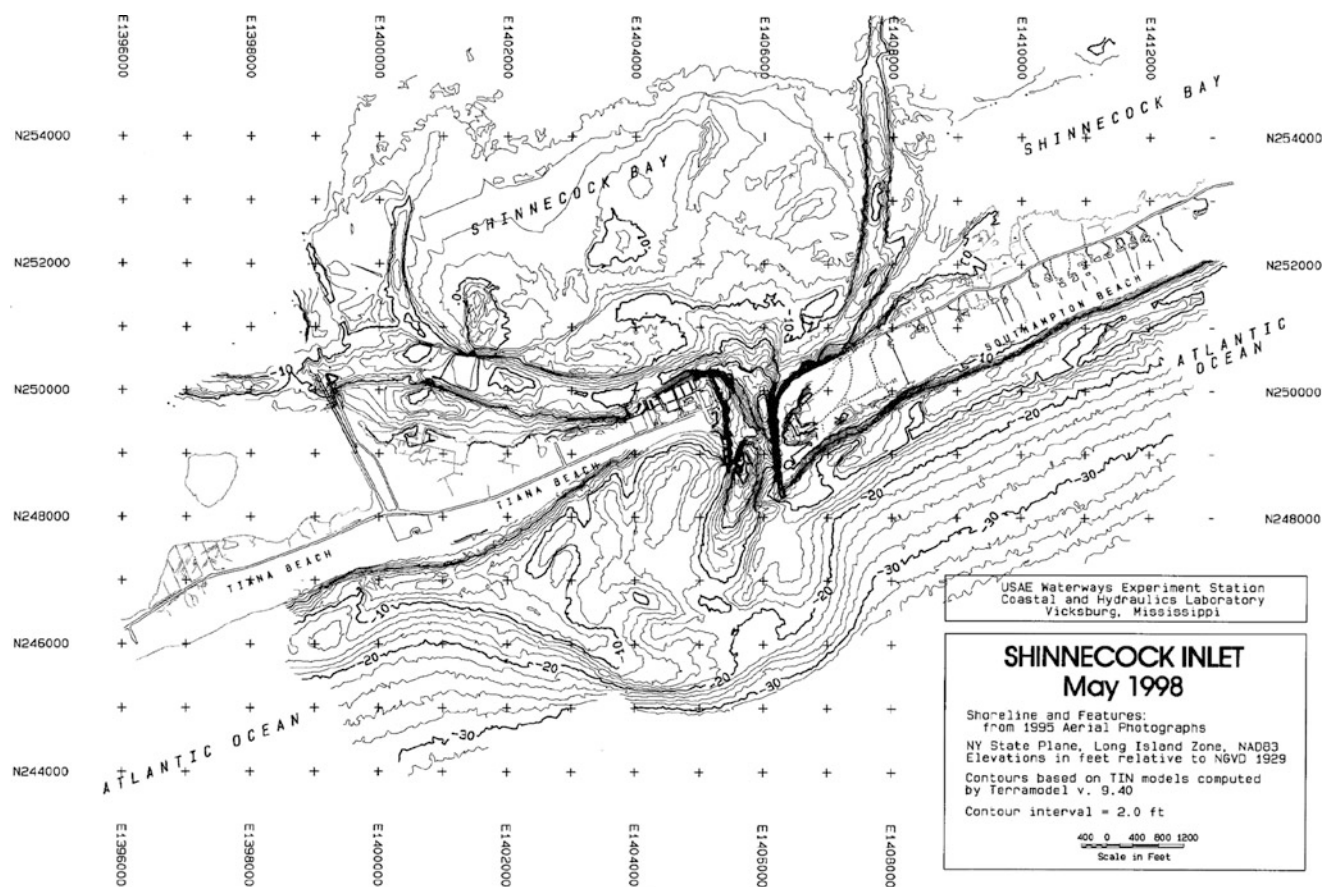
1. The most important advantage is that the system can survey up to 35 km²/h, thereby densely covering large stretches of the coast in a few days. This capability enables almost instantaneous data collection along shores subject to rapid changes.
2. The system can be mobilized quickly, allowing broad-area post-storm surveys or surveys of unexpected situations such as breaches across barriers.
3. SHOALS can survey directly from the nearshore zone across the beach, allowing efficient coverage of shoals, channels, or breaches that would normally be impossible or very difficult to survey using traditional methods, especially in winter. The system can now survey bluffs up to

10 m high, allowing a single platform to provide topographic and bathymetric data.

The main limitation of SHOALS is its critical dependence on water clarity. Maximum survey depth is over 30 m in clear water, like the Caribbean Sea, but the laser signal is drastically attenuated if the water is turbid due to suspended sediment, algae, or surf. Sometimes, a survey has to be delayed until a particular time of the year to be successful in an area that experiences river runoff or algae blooms. For example, the SHOALS survey at Shinnecock Inlet, on Long Island, New York, had to be carefully scheduled to avoid algal blooms (Fig. 5). Because of the immense amount of data that the system collects, data processing, archiving, and management are a challenge.

Airborne LIDAR Bathymetry (ALB) is a major advance in military operations because of its ability to rapidly survey unfamiliar littoral areas that may be subject to military operations (West et al. 2001). Conventional survey methods require the deployment of small boats equipped with single- or multi-beam acoustic equipment along with terrestrial surveyors to perform beach profiles. These traditional methods are slow, are highly susceptible to weather, and expose the surveyors to hostile fire.

A similar technology is Airborne Topographic Mapper (ATM), an aircraft-based system that provides highly detailed topographic data of beaches. The National Aeronautics and Space Administration (NASA) and a group of partners are in the process of mapping the entire coast of the continental United States in unprecedented detail.



Monitoring Coastal Geomorphology, Fig. 5 Example of contoured SHOALS hydrographic LIDAR data from Shinnecock Inlet, Long Island, New York, July 1989. Contouring based on 37,700 points, one-

tenth of the original data density. This shows how an ebb shoal has evolved since 1938 when the photographs in the previous figure were taken (from Morang 1999)

In Situ Technologies

The most common coastal applications of in situ systems are monitoring water flow and meteorological properties. Measuring waves and currents in the high-energy, hazardous nearshore zone is one of the more challenging endeavors of coastal engineering and research. Measurement programs must be thoroughly planned *before* any gauges are deployed to ensure that useful data are collected. A scientist must:

1. Determine what data units and analyzed products are needed to answer the critical engineering or scientific questions at the site.
2. Determine how long the gauges must be at the site (i.e., several years or just during the winter season).
3. Consider placing gauges in locations that are compatible with previous field programs.
4. Evaluate environmental constraints such as ice or trawler activity.
5. Be sure that enough funding is allocated for the analysis of the data.

Wave Gauges

Two general types of wave gauges are available: non-directional and directional. In general, directional gauges and gauge arrays are more expensive to build, deploy, and maintain than nondirectional gauges. Despite the greater cost and complexity, the former are preferred for most projects because the directional distribution of wave energy is an important parameter in many applications, such as sediment transport analysis and calculation of wave transformation. Wave gauges can be installed in buoys, placed directly on the sea or lake bottom, or mounted on existing structures, such as piers, jetties, or offshore platforms. Deployment of wave gauges is expensive and must be balanced against other project requirements. There are no hard-and-fast rules regarding how much wave or current data to collect – more is always better! Table 3 (from Morang et al. 1997b) provides broad guidelines for wave gauging.

A wave-gauging program requires several steps:

- Planning (determine goals, allocate budget).
- Instrument deployment, servicing, and recovery.

Monitoring Coastal Geomorphology, Table 3 Suggested wave gauge placement for coastal project monitoring

I. High-budget project (major harbor, highly populated area)
A. Recommended placement:
1. One (or more) wave gauge(s) close to shore near the most critical features being monitored (e.g., near an inlet). Although nearshore, gauges should be in intermediate or deepwater based on expected most common wave period. Depth can be calculated from formulas in the <i>Coastal Engineering Manual</i> (USACE 2002)
2. In addition, one wave gauge in deepwater if needed for establishing boundary conditions of models
B. Schedule:
1. Minimum: 1 year. Monitor winter/summer wave patterns (critical for Indian Ocean projects)
2. Optimum: 5 years or at least sufficient time to determine if there are noticeable changes in climatology over time. Try to include one El Niño season during coverage for North American projects
C. Notes:
1. Concurrent physical or numerical modeling: Placement of gauges may need to take into account modelers' requirements for input or model calibration. Placement in shallow water may be needed to test wave transformation models or wave hindcasts
2. Preexisting wave data may indicate that gauges should be placed in particular locations, or gauges may be placed in locations identical to the previous deployment to make the new data as compatible as possible with the older data. Long, continuous data sets are extremely valuable
3. Hazardous conditions: If there is a danger of gauges being damaged by anchors or fishing boats, the gauges must be protected, mounted on structures (if available), or deployed in a location that appears to be the least hazardous
II. Medium-budget project
A. Recommended placement:
1. One wave gauge close to shore near project site
2. Obtain data from nearest deepwater buoy (for US waters, NOAA, National Data Buoy Center (NDBC) buoys)
B. Schedule: Minimum 1-year deployment; longer if possible
C. Notes: same as IC above. Compatibility with existing data sets is very valuable.
III. Low budget, short-term project
A. Recommended placement: Gauge close to project site.
B. Schedule: If 1-year deployment is not possible, try to monitor the season when the highest waves are expected (usually winter, although this may not be true in areas where ice pack occurs or in tropical storm areas).
C. Notes: same as IC above. It is critical to use any and all data from the vicinity, anything to provide additional information on the wave climatology of the region

- Data recovery (either real-time or periodic for internal-recording gauges).
- Data reduction, quality control, and generation of statistics.
- Display of statistics and summary results.

Wave data analysis is a complex topic in itself. To a casual observer on a boat, the sea surface usually appears as a chaotic

jumble of waves of various heights and periods moving in many different directions. Most wave gauges measure and record a signal indicative of the changing elevation of the water surface. Unfortunately, when this signal is simply plotted against time, it provides little initial information about the characteristics of the individual waves that were present at the time the record was being made. Therefore, further processing is necessary to obtain wave statistics that can be used by coastal scientists or engineers to infer what wave forces have influenced their study area. Earle and Bishop (1984), Horikawa (1988), and Earle et al. (1995) provide background in the mathematics of wave analysis.

Another difficulty encountered in wave analysis is that different researchers are often inconsistent in their use of technical terms, and users are urged to be cautious of wave statistics from secondary sources and to be aware of how terms were defined and statistics calculated. For example, "significant wave height" (H_s) is defined as the average height of the highest one-third of the waves in a record. How long should this record be? Are the waves measured in the time domain by counting the wave upcrossings or downcrossings? Most automated processing programs *estimate* significant wave height by performing spectral analysis of a wave time series in the frequency domain and equating $H_s = H_{m0}$. The latter equivalency is usually considered valid in deep and intermediate water but may not be satisfactory in shallow water (Horikawa 1988). To prevent collecting the wrong data when specifying or planning a wave measurement program, researchers are urged to consult a standard list of wave parameters such as the one prepared by the IAHR Working Group on Wave Generation and Analysis (1989).

Current Meters

Currents can be measured by two general approaches. One of these, *Lagrangian*, follows the motion of an element of matter in its spatial and temporal evolution. The other, *Eulerian*, defines the motion of the water at a fixed point and determines its temporal evolution. Lagrangian current measuring devices are often used in sediment transport studies, in pollution monitoring, and for tracking ice drift. Eulerian, or fixed, current measurements are important for determining the variations in flow over time at a fixed location, such as in a tidal inlet or harbor entrance channel. Recently developed instruments combine aspects of both approaches.

Four general classes of current measuring technology have been tested or are in use (Appell and Curtin 1990):

- Lagrangian methods: drifters and dye.
- Spatially integrating methods: experimental (not used often).

- Point source and related technology: moored current meters.
- Acoustic Doppler Current Profilers (ADCP) and related technology (fixed point and boat-mounted).

The large number of instruments and methods that have been tested underscores that detection and analysis of fluid motion in the oceans is exceedingly complex. The difficulty arises from the large continuous scales of motion in the water. “There is no single velocity in the water, but many, which are characterized by their temporal and spatial spectra. Implicit then in the concept of a fluid ‘velocity’ is knowledge of the temporal and spatial averaging processes used in measuring it” (McCullough 1980, p. 106). In shallow water, particularly in the surf zone, additional difficulties are created by turbulence and air entrainment caused by breaking waves, by suspension of large concentrations of sediment, and by the physical violence of the environment. Trustworthy current measurement under these conditions becomes a daunting task.

Lagrangian methods: Dye, drogues, ship drift, bottles, temperature structures, oil slicks, radioactive materials, paper, wood chips, ice floes, trees, flora, fauna, and seabed drifters have all been used to study the motion of the oceans (McCullough 1980). A disadvantage of all drifters is that they are only quasi-Lagrangian sensors because, regardless of their design or mass, they cannot exactly follow the movement of the water. Nevertheless, they are particularly effective at revealing surface flow patterns if they are photographed or video recorded on a time-lapse basis. Simple drifter experiments can also be helpful in developing a sampling strategy for more sophisticated subsequent field investigations. Floats, bottom drifters, drogues, and dye are used especially in the littoral zone where fixed current meters are adversely affected by turbulence (e.g., see Ingle 1966). Resio and Hands (1994) examined the use of seabed drifters and commented on their value in conjunction with other instruments.

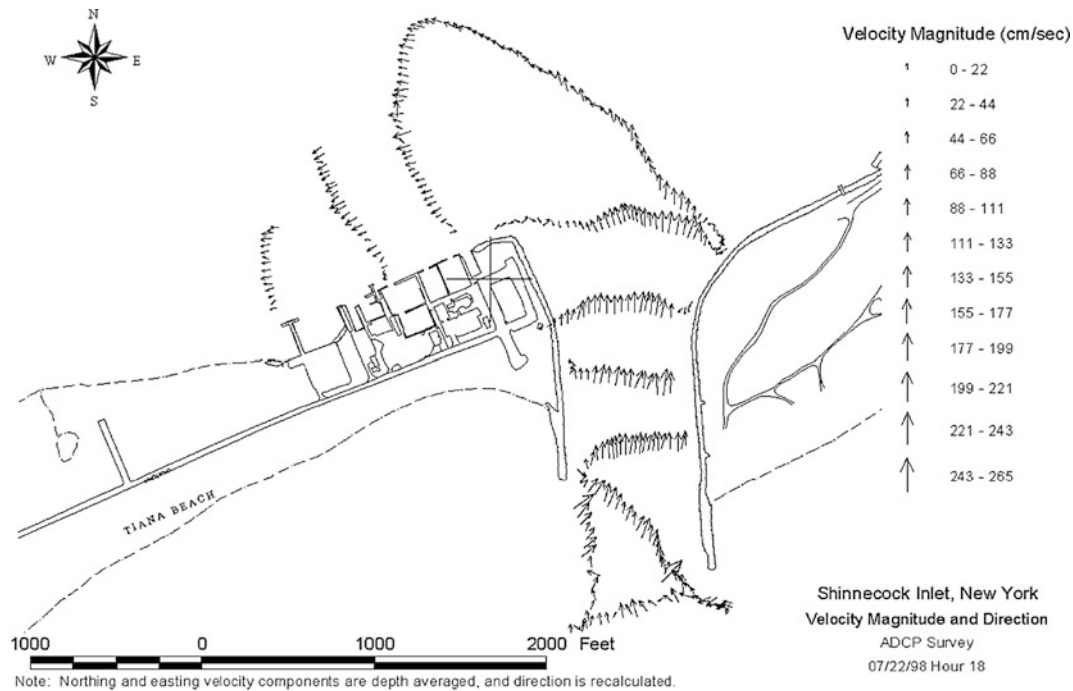
Point source (Eulerian) and related technology: In channels, bays, and offshore, direct measurements of the velocity and direction of current flow can be made by instruments deployed on the bottom or at various levels in the water column. Two general classes of current meters are available: mechanical (impeller-type) and electronic. Mechanical meters, based on rotors or propellers that turn in the water, have been largely phased out in favor of electronic instruments. These have the advantages of rapid response and self-contained design with no external moving parts to get fouled or corroded. They can be used in realtime systems and can be used to measure at least two velocity components, but only at a single depth. Vertical or horizontal arrays of meters can be deployed to measure the distribution of velocities vertically through the water column or across a specific area such as a channel.

Acoustic Doppler Current Profilers: These profilers (known as ADCPs) operate on the principle of Doppler shift in the backscattered acoustic energy caused by moving particles suspended in the water. Assuming that the particles have the same velocity as the ambient water, the Doppler shift is proportional to the velocity components of the water within the path of the instrument’s acoustic pulse. The backscattered acoustic signal is divided into parts corresponding to specific depth cells, often termed “bins.” The bins can be various sizes, depending upon the depth of water in which the instrument has been deployed, the frequency of the signal pulse, the time that each bin is sampled, and the acceptable accuracy of the estimated current velocity (Bos 1990). ADCPs have revolutionized current measurement in all water depths, but a great advantage of using ADCPs in shallow water is that they provide profiles of the velocities in the entire water column, allowing more comprehensive views of water motions than do strings of multiple point source meters. An ADCP can be moored at a fixed point to monitor the change in water velocities over time, or it can be mounted on a boat to measure flow patterns over an area, such as across an inlet (Fig. 6). ADCP data are inherently noisy, and signal processing and averaging are critical to successful performance.

Physical Sampling Methods

Despite the explosive development of remote sensing and geophysical technology during the last half of the twentieth century, there continues to be a need to collect actual samples of the sediment or rock underlying a study area. There are several reasons for this continued reliance on what is sometimes regarded as a primitive desire to touch the sediment. First, any remote sensing technique is still “remote.” It is an interpretation of what is down there and is hopefully an accurate picture, but this interpretation (or model) still needs field confirmation to provide ground-truthing. Hence, a seismic profiling study is often accompanied by a coring program. Second, many sediment characteristics can still be obtained only by analysis of the actual samples. For example, sieving is still the best way to determine the distribution of grain sizes in a sample. For geotechnical investigations, samples are needed to determine strength, organic content, compressibility, and permeability (USACE 1996). Surficial sediments provide information about the energy of the environment as well as the long-term processes and movement of materials, such as sediment transport pathways, sources, and sinks.

Three types of devices are available for retrieving sediments in the marine environment: grab samplers, dredges, and corers. For soft surface materials, a variety of grab-type samplers of different sizes and design have been developed (Bouma 1969; USACE 1996). Most consist of a set of



Monitoring Coastal Geomorphology, Fig. 6 Analysis of ADCP current-meter surveys in Shinnecock Inlet, Long Island, New York. The instrument was mounted on a boat that made transects across the inlet. The arrows show the direction and magnitude of the flow during

the flood tide. The tide enters the inlet from the Atlantic Ocean and diverges in the bay, with much of the flow moving to the west (from Pratt and Stauble 2001)

opposing, articulated scoop-shaped jaws that are lowered to the bottom in an open position and are then closed by various trip mechanisms to retrieve a sample. Some grab samplers are small enough to be deployed and retrieved by hand, but most require some type of lifting gear (Table 4). For hard seafloor areas, steel dredges can be dragged along the bottom from a boat, but dredges suffer from selective recovery (only loose boulders may be recovered), and positional information is poor.

For all sediment sampling, it is necessary to obtain a sufficient quantity of material to compute a statistically valid grain-size distribution, and if there is gravel in the sample, many liters of sample may be needed. For example, if the maximum particle size (10% of the sample) is 2.4 mm, only 0.1 kg of sample is required, but if the sample includes cobble of 64 mm, 50 kg are needed. Required quantities are specified in manuals prepared by the standards organizations of various countries (e.g., British Standards Institution 1990; American Society for Testing and Materials (ASTM) 2001).

Although surficial samples are helpful for assessing recent processes, they are typically of limited value in a stratigraphic study because grab devices usually recover less than 15 or 20 cm of material. To recover subbottom sediments, a type of coring system is needed. One of the simplest is the vibratory corer (vibracorer), commonly used by geologists to obtain samples in lacustrine, shallow marine, and coastal environments (Fuller and Meisberger 1982a, b; USACE 1996).

Vibracoring is described in a companion entry in this volume. The core may be up to 4 or 5 m long, which is adequate for borrow site investigations, near-surface cross sections, and other coastal studies.

Cores can be invaluable because they allow a direct, detailed examination of the layering and sequences of the subsurface sediment. The sequences provide information regarding the history of the depositional environment and the physical processes during the time of sedimentation. Depending upon the information required, the types of analysis that can be performed on the core include grain size, sedimentary structures, identification of shells and minerals, organic content, microfaunal identification (pollen counts), X-ray radiographs, radiometric dating, and engineering tests. If only information regarding recent processes is necessary, then a box corer, which samples up to 0.6 m depths, can provide sufficient sediment. Because of its greater width, a box corer can recover undisturbed sediment from immediately below the seafloor, allowing the examination of bedding microstructure and lamination. These structures are usually destroyed by traditional vibratory or rotary coring.

If it is necessary to obtain deep cores, or if cemented or very hard sediments are present, the only alternative is rotary coring. Truck- or skid-mounted drilling rigs can be conveniently used on beaches or on barges in lagoons and shallow water. Offshore, rotary drilling becomes more complex and expensive, usually requiring jack-up drilling barges or

Monitoring Coastal Geomorphology, Table 4 Subaqueous sediment sampling without drill rigs and casing

Device	Application	Description	Penetration depth	Comments
Grab samplers Petersen grab ^a	Large, relatively intact “grab” samples of sea-floor	Clam-shell type grab weighing about 450 kg with capacity about 0.4 ft ³ (0.1 m ³)	To about 10 cm	Effective in water depths to 600 m (or more with additional weight).
Corers (for subsurface sampling) Harpoon-type gravity corer	Cores 1.5–6-in., (3.8–15.2 cm) diameter in soft to firm sediments	Vaned weight connected to coring tube dropped directly from boat. Tube contains liners and core retainer	To about 10 m	Maximum water depth depends only on weight. Undisturbed (UD) sampling possible with short, large-diameter barrel
Free-fall gravity corer	(As above for harpoon type)	Device suspended on wire rope over vessel side at height above seafloor of about 5 m and then released	Soft sediment to about 5 m. Firm sediment to about 3 m	(As above for harpoon type)
Piston gravity corer (Ewing gravity corer)	2.5-in. (6.35 cm) sample in soft to firm sediments	Similar to free-fall corer except that coring tube contains a piston that remains stationary on the seafloor during sampling	Standard core barrel 10 ft. (3.0 m); additional 10-ft sections can be added	Can obtain high-quality UD samples
Piggott explosive coring tube	Cores of soft to hard bottom sediments	Similar to gravity corer. Drive weight serves as gun barrel and coring tube as projectile. When tube meets resistance of seafloor, weighted gun barrel slides over trigger mechanism to fire a cartridge. The exploding gas drives tube into sediments	Cores to 1 7/8 in. (4.75 cm) and to 10-ft lengths have been recovered in stiff to hard materials	Has been used successfully in 6000 m of water
Norwegian Geotechnical Institute gas-operated piston	Good-quality samples in soft clays	Similar to the Osterberg piston sampler except that the piston on the sampling tube is activated by gas pressure	About 10 m	
Vibracorer ^b	High-quality samples in soft to firm sediments. Dia. 3.0 in. (7.6 cm)	Apparatus is set on seafloor. Air pressure from the vessel activates an air-powered mechanical vibrator to cause penetration of the tube, which contains a plastic liner to retain the core	Length of 20 ft. (6 m) Rate of penetration varies with material strength. Samples a 20-ft core in soft sediment in 2 min	Maximum water depth about 60 m
Box corer	Large, intact slice of seafloor	Weighted box with closure of bottom for benthic biological sampling or geological microstructure	To about 0.3 m	Central part of sample is undisturbed

Source: Adapted from HUNT (1984) and other sources

^aSimilar grab devices include the Ponar[®], whose jaws close when the device strikes the seafloor, and the Ekman, whose jaws are activated by a messenger. A messenger is a metal plug that slides down the steel cable and strikes a spring-loaded jaw-control mechanism. Other types are the Van Veen and Smyth-Macintyre

^bVibracorer vary greatly in size. Many are custom-made at universities or consulting companies. Some are electrically or mechanically powered, and many use 3.0-in. (7.6 cm) thin-walled aluminum irrigation pipe as a combination core barrel and liner

four-point anchored drill ships. An experienced crew can drill and sample 100 m of the subsurface in about 24 h. Information on drilling and sampling practice is presented in Hunt (1984) and USACE (1996).

Mapping and Shoreline Change

Value of Comparing Historical with Modern Data

Collecting field data is just one phase of a coastal monitoring study. Analyzing, interpreting, and organizing the results are an equally important phase because this is how findings are

displayed and communicated to engineers, coastal managers, and policy-makers. Several levels of analysis are possible. The simplest typically is the measurement of linear changes of coastal features over time. Historical charts, modern maps, aerial photographs, and LIDAR or topographic data can reveal details on:

- Long- and short-term advance or retreat of the shore. Shoreline change data are critical for coastal managers tasked with establishing setback lines and guiding growth in the coastal zone, especially in low areas subject to flooding.

- The impact of storms, including barrier island breaches, overwash, and changes in inlets, vegetation, and dunes.
- Human impacts caused by coastal construction, dune destruction, or dredging.
- Compliance with permits, illegal filling, and dumping.
- Biological condition of wetlands, estuaries, and barrier islands.
- Susceptibility of urban areas to storm flooding and catastrophic events (e.g., hurricanes) by means of storm surge models.

If historical 3D (bathymetric and topographic) data are available, volumetric comparisons between the old and modern surveys can provide quantitative information on:

- Longshore sediment movement.
- Shoaling or siltation associated with tidal inlets, river mouths, estuaries, and harbors.
- Sediment changes on ebb and flood shoals and in inlet channels.
- Nearshore bathymetry changes over time.
- Migrations of channel thalwegs.

Coastal engineers often make volumetric comparisons to compute amounts of sediment trapped by structures, examine the growth of shoals in navigation channels, determine dredging requirements, compute dredging contract payment, and evaluate post-dredging channel conditions. Volumetric analyses are also used to monitor the performance of beach renourishment projects.

Definitions and Map Datums

Many coastal zone features and subdivisions are difficult to define because temporal variability or gradational changes between features obscure precise boundaries. In addition, nomenclature is not standardized, and ambiguity is especially evident in the terminology and zonation of shore and littoral areas. If an ambiguous term is not precisely defined (for instance, the ever-controversial “shoreline”), the intended boundaries may differ greatly. Therefore, one of the most critical issues when combining historical and recent geographical data is to establish what datums have been used. A datum is “a fixed or assumed point, line, or surface, in relation to which others are determined; any quantity or value that serves as a base or reference for other quantities or values” (Bates and Jackson 1984, p. 127). For coastal engineering and geologic studies, both horizontal (geographic location) and vertical (distance above sea level or other surface) datums must be established. Readers are referred to Umbach (1976), USACE (1994), Gorman et al. (1998), Moore (2000), and textbooks on geodetic surveying and photogrammetry for more background information on this complex topic.

Many possible datums can be used to monitor historical changes of the shoreline. The complexities around this term are discussed in a companion entry in this volume. Readers in the United States are also referred to the National Shoreline Data Standard, a draft standard prepared by NOAA, which defines shoreline as,

The line of contact between the land and a body of water. On Coast and Geodetic Survey nautical charts and surveys, the shoreline approximates the mean high-water line. In Coast Survey usage, the term is considered synonymous with coastline (Shalowitz 1962).

In many situations, the high water line (hwl) has been found to be the best indicator of the land–water interface, the coastline (Crowell et al. 1991, 1993). The hwl is easily recognizable in the field and can sometimes be identified in aerial photographs by a line of seaweed and debris (wave run-up line). The wet-dry line, also visible on air photos, is more difficult to interpret because it is a function of water table, which in turn is a function of tide level, recent rainfall, and other factors. The datum printed on historical US National Ocean Service (NOS) T-sheets (topographic) is listed as mean high water. Fortunately, the early NOS topographers approximated hwl during their survey procedures (Shalowitz 1962). Therefore, direct comparisons between historical T-sheets and modern aerial photograph interpretations are possible. However, even years ago, coastal scientists recognized that the hwl as mapped from aerial photographs was a dynamic feature and potentially a less reliable indicator of shoreline position than morphologic features on the shore that remain relatively constant from day to day. Some of these morphologic indicators are the berm crest, base of the dune, permanent vegetation line, crest of the overwash terrace, and shore protection structures, all of which are unaffected or only slightly altered by short-term changes in water levels (Morton 1997). Therefore, many recent field surveys using Global Positioning System (GPS) instruments mounted on all-terrain vehicles are defining the shoreline based on these morphologic features. Whether these shorelines can be directly compared with older air photo interpretations is problematic.

A discussion of water level datums and shoreline definition is beyond the scope of this entry. An introduction to this complicated topic is presented in NOS (1988). The classic *Shore and Sea Boundaries* (Shalowitz 1962, 1964; Reed 2000) are exhaustive references on technical and legal issues relating to datums in the United States. USACE (1994) provides details on establishing tidal datums for coastal hydrographic surveys. In the Great Lakes of North America, water levels fluctuate on yearly and on longer-term cycles due to meteorologic and hydrologic conditions over the continent. In the lakes, charts and water depths are referenced to the International Great Lakes Datum of 1985 (Coordinating Committee 1992).

Cross-Shore Profile Surveys

Topographic and nearshore profile surveys provide one of the most direct and accurate means to assess geologic and geomorphic changes on the shoreface to water depths of 10 or 15 m. Beach profiles conducted over time can document erosion and accretion in the coastal zone, shoreline changes, sand bar movement, and dune volume changes. They provide the basic data for evaluating what happens to sand placed in beach nourishment projects (Weggel 1995).

Permanent or semi-permanent benchmarks are required to reoccupy profile sites over successive months and years. These benchmarks should be located on a stable land feature at the landward end of the profile lines to minimize their likelihood of being damaged in storms. The locations of survey monuments must be carefully documented and referenced to other survey markers or to control points. Then, the profile data can be used to evaluate changes in sea level, changes in volume of sand on the beach, or other phenomena that require reference to established regional datums. The ability to accurately reestablish a survey monument is critical because it ensures that profile data collected over many years will be comparable. Locations that might experience dune erosion should be avoided, and care should also be taken to reduce the visibility of benchmarks to minimize vandalism. (Unfortunately, damage caused by vandals is an irritating and expensive problem at all coastal projects, even those far from urban areas.) With the increasing use of GPS receivers by surveyors, the rigorous need for duplicate survey monuments may be reduced. This is an evolving technology, and for now we still recommend that two monuments per survey line be established. US Army Corps of Engineers standards for survey monuments are specified in USACE (1990).

When planning a beach profiling study, both the frequency of the sampling and the overall duration of a project must be considered. There are no definitive guidelines for the timing of profile lines, but for most sites, summer and winter surveys are recommended. Resurveying profiles over a period of more than 1 year can be of substantial help in understanding seasonal changes. In addition, supplemental surveys can be made after big storms to determine their effects and measure the rate of recovery of the local beach. Table 5 outlines a suggested survey schedule for monitoring beach fill projects.

The longshore spacing of profiles must be carefully planned. Profiles should be at close enough intervals to show any significant changes in lateral continuity. Reviewing locations of historical shorelines in the study area is one way to establish the gross limits of the area that should be examined in detail, particularly along rapidly changing coasts. Often, a spacing of 1,000 ft. (300 m) is used in studies in the United States, but closer spacing may be needed near inlets, within project boundaries, or at the ends of spits or barriers that experience frequent changes,

Monitoring Coastal Geomorphology, Table 5 Example of profile survey scheme for monitoring beach fill (after Stauble 1994)

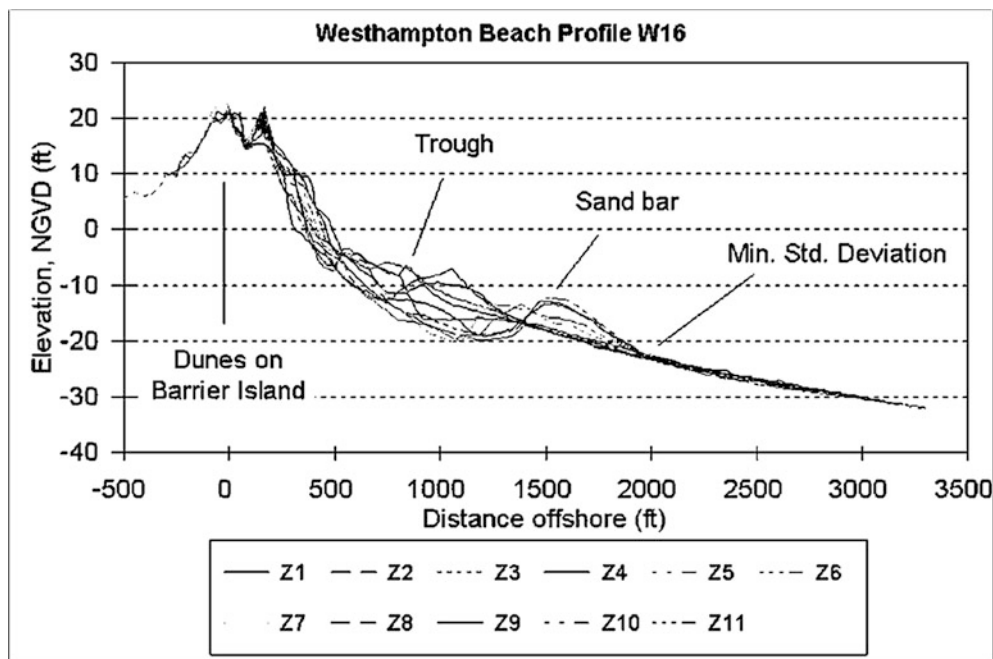
Time	Times/ year	Number of profiles
Pre-fill	2	Collect within fill area and at control locations in summer and winter months to characterize seasonal profile envelope (beach and nearshore to closure depth)
Post-fill	1	Collect all profiles immediately after fill placement at each site (beach and offshore) to document fill volume Collect control profiles immediately after project is completed
Year 1	4	Four quarterly survey trips collecting all beach and offshore profiles out to the depth of closure. Begin series during the quarter following the post-fill survey
Continue year 1 schedule to time of next renourishment (usually 4–6 years). If project is a single sand placement, taper surveys in subsequent years		
Year 2	2	6- and 12-month survey of all beach and offshore profiles
Year 3	2	6- and 12-month survey of all beach and offshore profiles
Year 4	1	12-month survey of beach and offshore profiles

Note: (1) If project is renourished, repeat survey schedule from post-fill immediately after each renourishment to document new fill quantity and behavior. (2) Monitoring fill after major storms is highly desirable to assess fill behavior and storm-protection ability. Include both profile and sediment sampling. Conduct less than 1 week after storm conditions abate to document the beach and offshore response

and fewer profiles may be needed in the more stable portions of barrierislands.

For proper coverage, profile lines should extend landward of the zone that can be inundated by storms, usually behind the frontal dunes. But, often it is not necessary to survey across an entire barrier island. For example, shore and dune deposits that are now inland from the modern shoreline may only be affected by marine or lacustrine processes during the most severe storms. Aerial photographs of these interior areas may be adequate to show morphologic changes. Seaward, the profiles should extend deep enough to include the portion of the shoreface where most sediment moves (i.e., to beyond closure – discussed in another entry in this volume). Areas subject to dune breaching and overwash require special efforts to establish a reusable benchmark system.

The land portion of a profile is typically surveyed with a stadia rod and a total station. The method is inexpensive and can extend offshore to about 1 m water depth (or deeper if the rodman is adventurous). The nearshore seafloor is often surveyed by a sled that is towed by boat out into the water from about 1.5 m to closure depth. This results in overlap between onshore rod surveys and sled surveys to assure that the two systems are recording the same elevations. With careful field technicians, vertical accuracy for sled surveys is in the range of 0.15 m (Clausner et al. 1986). If offshore surveys are conducted by boat-mounted echo sounder,



Monitoring Coastal Geomorphology, Fig. 7 Example of cross-shore profile surveys from Westhampton Beach, Long Island. Surveys were made at 6 month intervals and demonstrate significant movement

of the bar and trough. The location where the profiles converge, at the position of minimum standard deviation, is sometimes interpreted as the depth of closure

overlap with rod surveys is often not possible because most boats cannot survey in water shallower than about 2 m. Also, acoustic surveys are difficult in the surf zone because bubbles in the water attenuate the acoustic pulses. Jet skis have proven to be a useful way to obtain data in the difficult surf zone or in tidal inlets when a towed sled is not available or is not practical due to logistical conditions (Dugan et al. 1999). With advances in remote sensing technology and data processing methods, remote sensing may replace field profile surveys in some circumstances (Judge and Overton 2001).

Figure 7 shows an example of profile data collected at Westhampton Beach, Long Island, New York (facing southeast towards the Atlantic Ocean). In many areas around the world, beaches show a distinct pattern of summer growth and winter erosion. But along the south shore of Long Island, analysis of profiles has revealed that the seasonal effect is subtle (Morang et al. 1999). Storms occur during both the winter and the summer, and although the berm and dune crest often retreat and bars move offshore during each event, recovery typically is rapid as littoral material moves back onshore. For this study, the ability to detect subtle sand volume changes along the Long Island shore has underscored the value of a time series of topographic data spanning several years.

Conclusions

Inexpensive computing and data storage technology and innovative sensors have brought about a revolution in marine

data collection during the last two decades that shows no sign of ending. Advances have occurred in four areas: (1) sensors; (2) interactive or real-time data analysis of data; (3) data integration among different sensors and navigation systems; (4) data display and presentation, particularly with respect to Geographic Information System (GIS) organization and display. These new technologies have allowed marine scientists to examine phenomena in much greater detail than ever before and with much greater positional (navigation) accuracy. However, despite these advances, some areas of the coastal zone continue to be hazardous and difficult for any form of in situ data collection or monitoring. This is especially true of the breaker zone, where few instruments can survive during storms and where it is still difficult to obtain trustworthy bathymetry. As a result, some basic processes, such as the cross-shore distribution of longshore currents, are still little understood. Perversely, the more data we collect, the more we realize that we still know so little about marine processes in the coastal zone and the delicate interplay between hydrodynamics, geology, and biology.

Cross-References

- [Beach and Nearshore Instrumentation](#)
- [Beach Sediment Characteristics](#)
- [Coastal Warfare](#)
- [Global Positioning Systems \(GPS\)](#)

- Jet Probes
- Monitoring Coastal Ecology
- Photogrammetry
- Remote Sensing of Coastal Environments
- Remote Sensing: Wetlands Classification
- Shoreline and Coastal Terrain Mapping
- Synthetic Aperture Radar Systems
- Vibracore

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Morphodynamic Stability of Tidal Inlet-Bay Systems

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Definition

Sandy tidal inlets and associated back-barrier bays tend to show remarkably stable morphology in spite of sea level rise. This is brought about by morphodynamics which includes the process by which an inlet-bay system achieves a stable equilibrium configuration or form. In general, the relationship between process and form can be described empirically, by analytic modeling and by numerical modeling. The most important feature of morphodynamics is a dynamic feedback between process and form as the form-restoring mechanism.

Conditions under which inlet-bay systems achieve and sustain equilibrium configuration due to tide and wave forcing are briefly reviewed following the basic inlet stability model of Escoffier (1940). Several inlet-bay systems in the state of Florida have been selected as case studies.

Introduction

Tidal inlets flanking barrier islands are prominent coastal landforms along extensive sandy stretches of the Atlantic, Gulf of Mexico, and Pacific coasts of the USA. Information on how inlet-bay systems respond over short (days to years) and long (decades and longer) durations to astronomical tide, storm surge, water waves, wind, sea level changes, and biology is of considerable scientific and engineering significance (Otvos 2017a, b). Globally, many inlets and back-barrier bays show remarkably stable morphology in spite of the secular rise in sea level during the late Holocene epoch (e.g., last ~3000 years) (Davis 1994; Mörner 2017; Bird 2018). In that context, morphodynamics encompasses the process by which a loose-boundary landform, e.g., a sandy tidal channel, achieves a stable equilibrium configuration. If equilibrium is perturbed by bed scour or sediment deposition due to a storm or by engineering, the morphodynamic process can cause the channel to revert to its original configuration. If the perturbation is large, the channel may achieve a new configuration, or it may lose its integrity and perhaps close. We will briefly review conditions under which inlet-bay systems sustain stability. These systems are characterized by mobile beds of sand subject to tide and waves as the main sediment driving forces.

In general morphodynamics can be described in three ways – by empirical relationships, by simple analytic modeling, and by complex numerical modeling. In all cases the most important underpinning aspect of morphodynamics is a dynamic feedback between process and configuration or form as the equilibrium-restoring mechanism (Wright and Thom 1977). In that context a limitation of purely empirical relationships is that from them the role of feedback may not be apparent or whether it is absent altogether. Analytic models inherently include a simple feedback mechanism, and the functioning of numerical models is dependent on generally complex feedbacks, which also makes it feasible to address more complex morphologies.

Referring to the littoral state of Florida, we note that along its 1300 km sandy shoreline, there are over 70 inlets (Fig. 1) and related back-barrier bays. Although several inlets are artificial and some are only a few decades old, others have been in existence for centuries (Davis 1994). Nearly 80% of the state's 20 million residents live near its Atlantic and Gulf of Mexico coasts, and over a hundred million tourists visit its barrier island beaches each year (Haven 2015). Thus, it is critical that the basis of morphologic resilience of inlet-bays

be understood for an efficient management of the coastal zone. For illustrative case studies, we will consider several systems, many based on experience in Florida as a paradigm.

Coastal Morphology, Morphodynamics, and Modeling

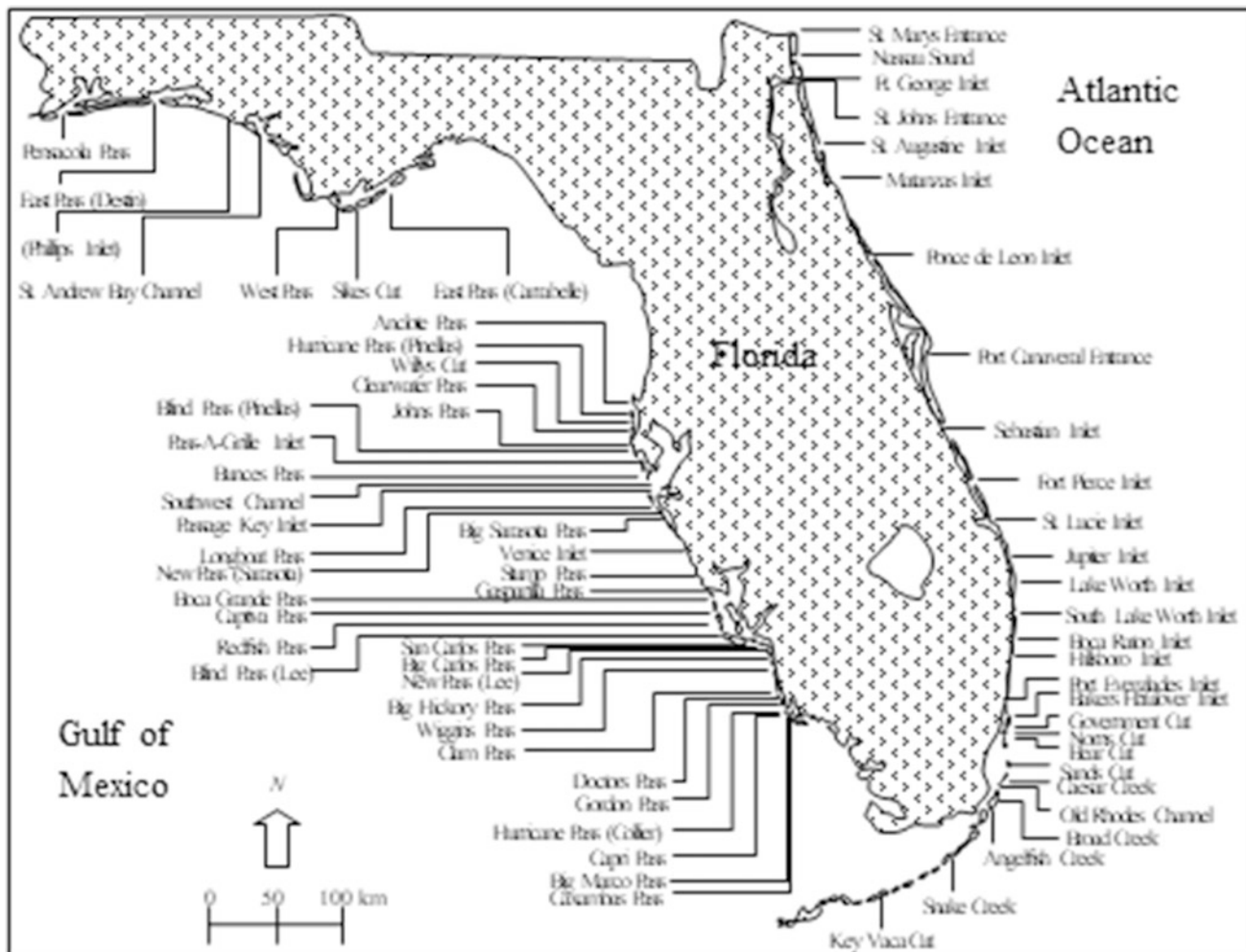
As a starting point the following definitions are relevant:

Coastal morphology is the study of coastal landforms, while *coastal morphodynamics* concerns the evolution of the landforms mainly due to tide, waves, and sea level change, although wind and, particularly, biological effects can have significant roles (Kakeh et al. 2016). *Coastal morphodynamic modeling* involves simulation of the evolution of coastal landforms.

The inception of coastal morphodynamics as a definable discipline began a few decades ago, initially relying on empirical relationships and analytic calculations. The need for numerical modeling arose because coastal processes occur over space and time scales spanning several orders of magnitude and over which landforms exhibit complex hierarchical forms. Consider, for instance, coastal features starting with a sand grain at the smallest scale and a barrier island at the other (Fig. 2). The length scale of a grain is $O(10^{-4}–10^{-3} \text{ m})$, and its rolling motion in turbulent flow has a period of $O(10^{-2}–10^0 \text{ s})$. In contrast, the length and evolutionary time scales for the barrier island and flanking inlets may be $O(10^3–10^6 \text{ m})$ and $O(10^8–10^{10} \text{ s})$, respectively.

The development of hierarchical forms exemplified in Fig. 2 is of scientific concern. From the engineering viewpoint, larger systems typically tend to be of greater concern. For instance, in economic terms, information on the morphology of a recreational beach is perhaps the most important design requirement for the nourishment of a popular but perhaps eroding beach, which may also impact neighboring inlets (Finkl and Jesse 2018). Unfortunately, even for large systems, morphodynamic modeling remains an imperfect tool because the mode of development of any form from a hierarchically lower one, e.g., small to large bedforms, is not fully understood. Accordingly, in practice, process-based modeling is generally limited to time scales of months to years, whereas decades and centuries are often dealt with empirically (de Vriend 1991a, b; van de Kreeke and Brouwer 2017; Dronkers 2016).

The earliest well-known analytic model for the assessment of tidal inlet stability and estimation of its equilibrium cross-sectional area is due to Escoffier (1940). This seminal development, as interpreted by, among others, van de Kreeke and Brouwer (2017), is the basis of inlet-bay analytic morphodynamic models summarized here. As we will note, the same basis is relevant to the relationship between back-barrier tidal flats and salt marshes and the formation of stable



Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 1 Florida's tidal inlets (Phillips Inlet (top left) undergoes periodic opening and closure)

morphologic cells in watercourses with intertwined shoals and channels.

A Typical Inlet-Bay System

As a typical inlet-bay system, consider Matanzas Inlet (Fig. 3a) south of Anastasia Island along the northeastern Atlantic coast of Florida (Fig. 1). The bay is defined by Matanzas River and the Atlantic Intracoastal Waterway (ICWW) (Fig. 3b). Important features include land areas above the local mean sea level (*msl*), sandy beaches, tidal channels, intertidal flats and shoals, and halophyte salt marshes. Some of the shoals define the characteristic ebb and flood deltas. In consonance with the seasonal directions of wave approach, the littoral sand drift is southward for about 7 summer months and northward during the 5 winter months. However, as the net annual drift is dominantly southward, the ebb delta is slightly skewed in that direction. To some extent for the

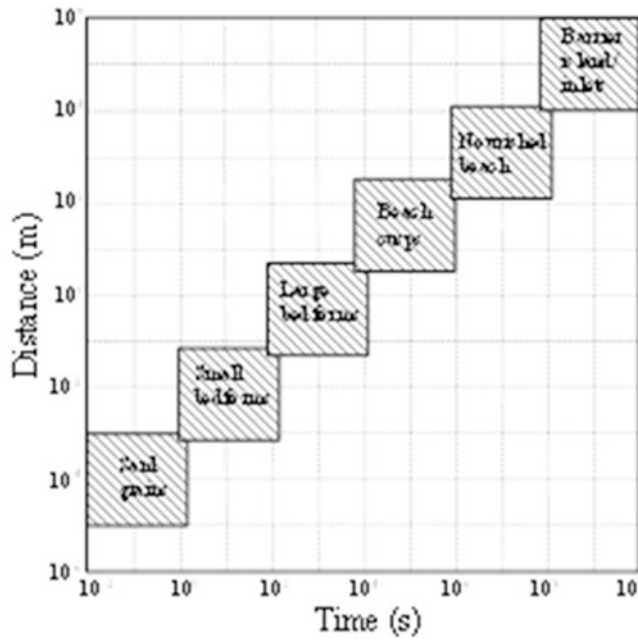
same reason, tidal flow occurs in preferred channels – flood flow enters from the north, and ebb flow exists southward.

During the passage of hurricane Matthew in 2016, the barrier island was breached about 1.35 km south of Matanzas Inlet. As a result a large slug of sand spilled into the south arm of Matanzas River and blocked it entirely. The breach (Matthew Cut) had to be closed by bulldozing the spill back to the island, as stability calculations suggested that it was large enough to remain open and adversely affect efficient functioning of Matanzas Inlet.

Conditions for a Stable Inlet

Equilibrium Cross Section

The stability of a sandy inlet is specified by the channel cross-sectional area. As the area varies along the length of the channel, the smallest area called the throat is taken as the characteristic value (O'Brien 1969). In the following we will



Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 2 Hierarchy of coastal landforms (see also de Vriend 1991a,b)

summarize an analytic method for determining the stable throat area from the governing equations of flow (van de Kreeke and Brouwer 2017; Escoffier 1940).

The throat area A_c (below *msl*) depends on the tide-maximum current velocity u_m . Referring to Fig. 4, let $\eta_0(t)$ be the (time-dependent) tidal water surface in the sea (or ocean) and $\eta_b(t)$ in the bay. For given A_c , u_m depends on the amplitude a_0 (or range $2a_0$) of $\eta_0(t)$, the tidal period T (or frequency $\sigma = 2\pi/T$), and the bay surface area A_b (of nominal linear dimension L_b). Let τ be the flow-induced shear stress at the channel bed. It is desired to determine A_c at equilibrium defined as the condition of zero net transport of sediment due to sediment erosion and deposition in the channel (per tidal cycle). The relevant shear stresses are the critical shear stress for erosion τ_{cr} and the shear stress τ_{eq} at sedimentary equilibrium (Julien 2010). For given sediment, diameter τ_{cr} is constant but τ and τ_{eq} characteristically vary with A_c .

Over the range of possible values of A_c at inlets with sand beds of about the same mean diameter (0.3–0.4 mm) in Florida, the aspect ratio h_c/w_c , where h_c is the depth and w_c is the width, is found to vary with A_c (Powell et al. 2006). Inasmuch as $w_c = A_c/h_c$, we may write

$$h_c = pA_c^q \quad (1)$$

where p and q are empirical coefficients. This morphologic relationship plays an essential role in the calculation of A_c at equilibrium. Data on Florida's Atlantic and Gulf Coast inlets along with values from moveable-bed laboratory models suggest $p = 0.182$ and $q = 0.419$ (Fig. 5).

For sustenance of morphology and function, the channel must be (1) morphodynamically stable and (2) in sedimentary equilibrium. For determination of the stable equilibrium area, τ and τ_{eq} have been plotted against A_c in Fig. 6, resulting in two curves.

Relative to the hydraulic curve $\tau = f'(A_c)$ (see subsequent discussion), where f' denotes a function, with an increase in A_c , τ rises at first because bed resistance to flow decreases as the channel becomes deeper and wider. At large values of A_c , τ begins to decrease as the head $\eta_0 - \eta_b$ driving the flow in the wide and deep channel decreases. When A_c becomes large, τ approaches zero as the channel becomes very deep and wide. Relative to the sedimentary curve $\tau_{eq} = f''(A_c)$ (subsequent discussion), in which f'' is another function, the effect of sediment transport on A_c is such that the curve decreases gradually. The stable equilibrium area A_{cV} occurs at the intersection V of the two curves, where $\tau = \tau_{eq}$. In general, intersection is possible only when the peak shear stress τ_{III} is greater than τ_{eqIII} . When τ_{III} is less than τ_{eqIII} , there can be no stable inlet.

If the initial value of A_c is not equal to A_{cV} , morphodynamic feedback can either change A_c to A_{cV} resulting in a stable channel or close an unstable channel, i. e. result in $A_c = 0$. Let $\Delta\tau$ and ΔA_c denote “small” positive or negative increments in τ and A_c , respectively. Starting with an arbitrary area A_{cIV} , a positive increment $+\Delta\tau$ will enlarge the channel by $+\Delta A_c$ due to scour due to increased bed shear stress. However, in the now deeper and wider channel, due to a decrease in the current velocity, τ will decrease by $-\Delta\tau$, which in turn will promote sediment deposition and revert the area to A_{cIV} . In a similar vein, a decrease $-\Delta\tau$ and deposition accompanied by an increase in the current velocity (and τ) followed by scour will revert the area to A_{cIV} . Thus A_{cIV} is a stable area; however, it is not necessarily the equilibrium value (A_{cV}).

It can be argued that beginning with any area A_{cII} greater than A_{cV} , as τ is greater than τ_{eq} , the area will enlarge by incremental scour to A_{cV} . In a somewhat similar sense, if the initial area is A_{cVI} , as τ is smaller than τ_{eq} , the area will decrease to A_{cV} by incremental deposition. In contrast, if A_c decreases by, say, storm-induced deposition (or a dredge-and-fill operation) to A_{cVII} that is less than A_{cV} , the channel will close by sequential cycles of incremental deposition and decrease in τ . Thus a channel can be stable only when A_c is greater than A_{cV} .

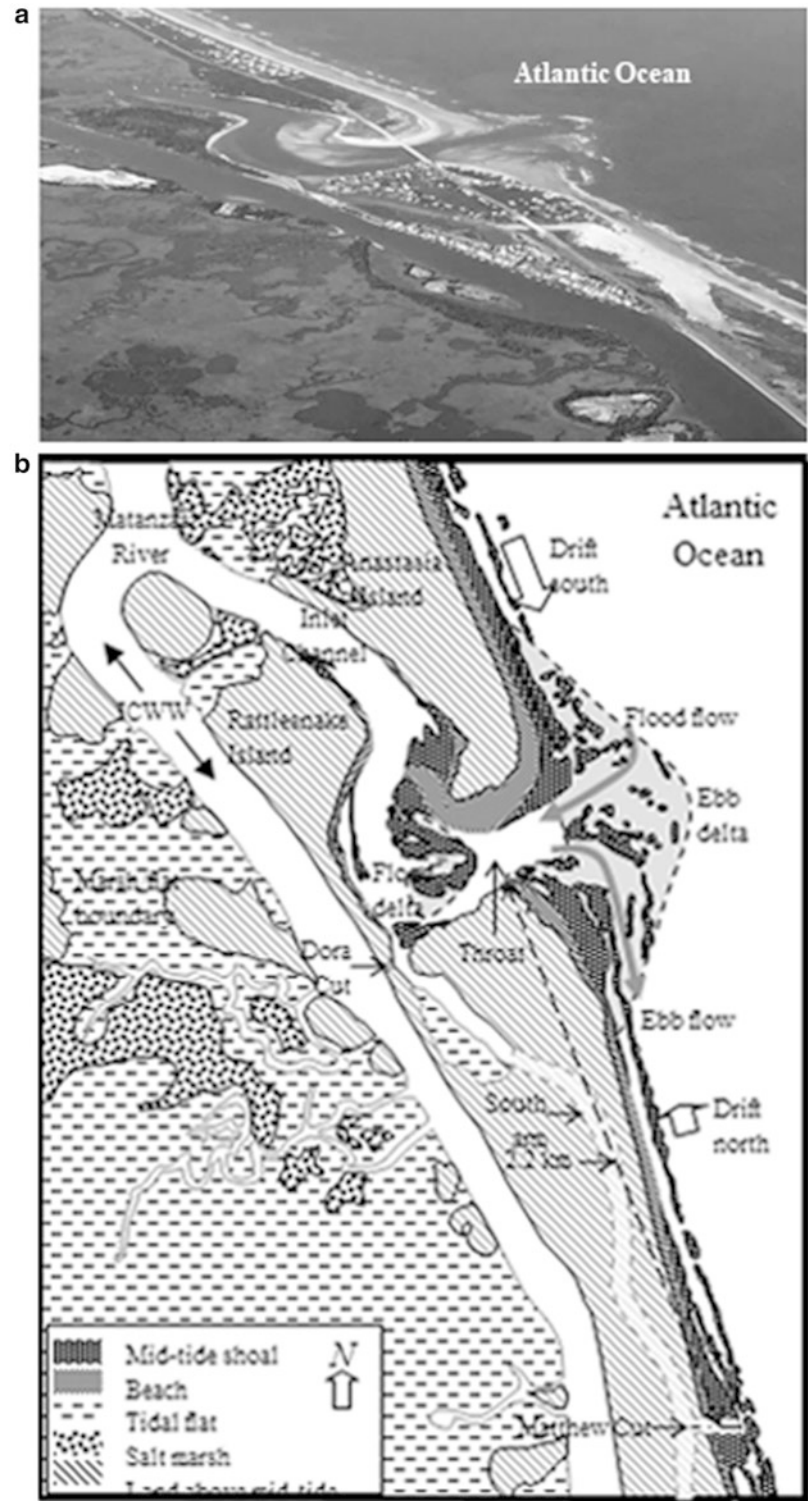
Given tide-maximum value u_m of the instantaneous current velocity $u(t)$, the respective bed shear stress τ in Fig. 6 is defined in hydraulics (Julien 2010) as

$$\tau = \frac{\rho_w g n^2}{h_c^{1/3}} u_m |u_m| \quad (2)$$

and similarly

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Fig. 3 (a) Matanzas Inlet and back-barrier bay in 2017. (Adapted from Google image by Capt. Rich Santos, flyfishjax.com). (b) Morphologic features at Matanzas Inlet and vicinity



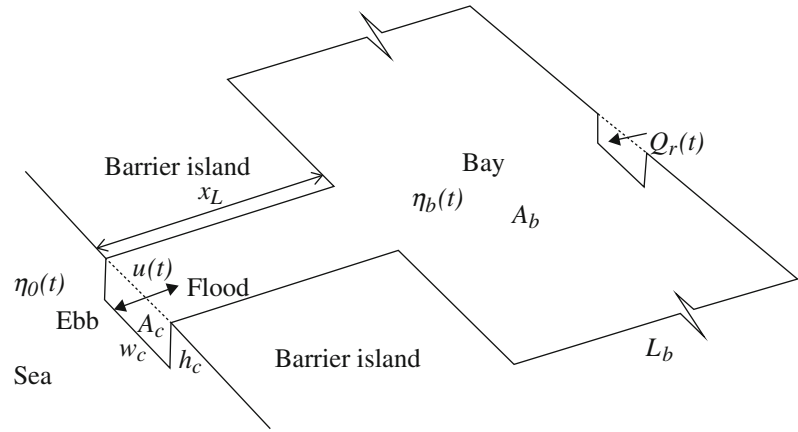
$$\tau_{eq} = \frac{\rho_w g n^2}{h_c^{1/3}} u_{eq} |u_{eq}| \quad (3)$$

where ρ_w is the water density, g is the acceleration due to gravity, n is the Manning's bed resistance coefficient, and u_{eq}

is the current velocity at equilibrium. These relationships make it equally convenient to consider the hydraulic and sedimentary curves in terms of current velocity in lieu of shear stress.

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Fig. 4 Inlet-bay system parameters



Hydraulic Relationship $\tau = f'(A_c)$

The hydraulic curve is derived from the respective depth-integrated equations of momentum and continuity of tidal flow in the channel (van de Kreeke and Brouwer 2017):

$$\frac{d^2\eta_b}{dt^2} + \frac{A_b F}{2x_L} \frac{d\eta_b}{dt} \left| \frac{d\eta_b}{dt} \right| = \frac{gA_c}{x_L A_b} (\eta_0 - \eta_b) \quad (4)$$

$$u = \frac{A_b}{A_c} \frac{d\eta_b}{dt} + \frac{Q_r}{A_c} \quad (5)$$

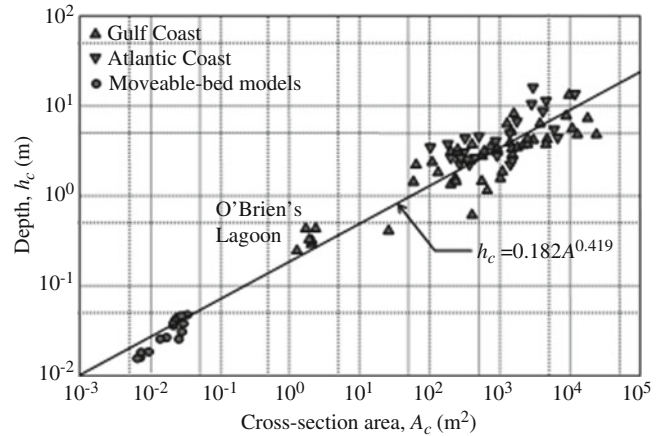
In the momentum balance Eq. (4), the first term represents the effect of inertia of water mass in the channel, the second accounts for energy dissipation due to bed resistance, and the third represents hydrostatic pressure force due to the instantaneous head $\eta_0(t) - \eta_b(t)$. In the continuity equation Eq. (5), $Q_r(t)$ is the river discharge entering the bay (and leaving the inlet). The ratios a_b/h_c and x_L/L_b are assumed to be small, in the bay currents are negligible, and as the bay is assumed to be deep (hence “frictionless”), at any instant η_b is the same everywhere in the bay.

To solve Eqs. (5) and (6) analytically, dissipation (the middle term in Eq. (4)) is linearized (van de Kreeke and Brouwer 2017). As a result, Eq. (2) reduces to a linear relationship between τ and u_m , and in turn τ in Fig. 6 may be readily replaced by u_m , which is obtained by solving Eq. (4) in the linearized form simultaneously with Eq. (5). The result is the hydraulic relationship:

$$u_m = \frac{\sqrt{2k_0 A_c^{(4q+3)/3}}}{k_4} \sqrt{\sqrt{\left(1 - \frac{k_1}{A_c}\right)^4 + \left(\frac{k_4}{A_c^{2(2q+3)/3}}\right)^2} - \left(1 - \frac{k_1}{A_c}\right)^2}$$

$$k_0 = a_0 A_B \sigma; k_1 = \frac{x_L A_B \sigma^2}{g}; k_2 = \frac{a_0 A_B}{2x_L}; k_3 = \frac{2gx_L n^2}{p^{4/3}}; k_4 = \frac{16}{3\pi} k_1 k_2 k_3 \quad (6)$$

Define the tidal prism P as the volume of water which enters the bay during flood flow each tidal cycle. In Eq. (6), it is assumed that, typically,



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Fig. 5 Variation of inlet channel depth with cross-sectional area. O'Brien's Lagoon no longer exists. (Data from Powell et al. 2006)

$$\frac{Q_r T}{2} \ll P \quad (7)$$

i.e., river discharge volume over one-half tidal cycle is negligible compared to the tidal prism.

Sedimentary Relationship $\tau_{eq} = f''(A_c)$

From data on five sandy inlets in California, LeConte (1905) observed that “nature requires 33 sq.ft. of mean-tide section for each and every million cubic feet of tidal waters passing in and out at spring tides.” A generalized form of this statement is

$$A_c = \Omega P^m \quad (8)$$

O'Brien (1969) showed that for 28 US inlets, the coefficient $\Omega = 4.69 \times 10^{-4}$, and the exponent $m = 0.85$ are best-fit values (with prism P measured in m^3 and A_c in m^2).

Equation (8), which is also applicable to mean tides, has been interpreted in terms of sediment transport (Hughes 2002;

Powell et al. 2006; Mehta et al. 2015). This is done for the condition of sedimentary equilibrium using tide-averaged volumetric sediment load Q_s to show that m has a “theoretical” value of $6/7$, or 0.86 , which is remarkably close to O’Brien’s (1969) 0.85 . Furthermore, it can be shown that, consistent with $m = 6/7$,

$$\Omega = \left(K \frac{C_k \sigma^6}{\pi^6 P^2 Q_s} \right)^{1/7} A_c^{4/21} \quad (9)$$

The coefficient $C_k = 0.86$ accounts for the effect of over-tide, which increases P above its value obtained from discharge assumed to be sinusoidal (Keulegan 1967). The sediment transport coefficient K is

$$K = \frac{20gn^6}{9v(s-1)^2} \quad (10)$$

where v is the kinematic viscosity of water and s is the specific weight of sand.

If we consider estuaries in equilibrium rather than merely their short inlet segments, $m = 6/7$ appears to be reasonable

as well (Fig. 7), which lends credence to the physical basis of Eq. (8). The mean value of Ω is 1.50×10^{-3} , and a noteworthy feature of the plot is that it includes the large Delaware estuary with a mouth (inlet) area of $2.97 \times 10^5 \text{ m}^2$ as well as the small Unnamed estuary, VA (4.65 m^2). By comparing $\Omega = 1.50 \times 10^{-3}$ with O’Brien’s 4.69×10^{-4} , we can infer that the inlet throat tends to be smaller than would be predicted by the estuary value of Ω . This is because it does not quite account, if at all, for the channel widening effect of waves in the inlet. Note that a fraction of this difference is likely to be due to the use of mean tides for the estuaries as opposed to spring tides for the inlets.

From the definition of the tidal prism, it can be shown that (Keulegan 1967)

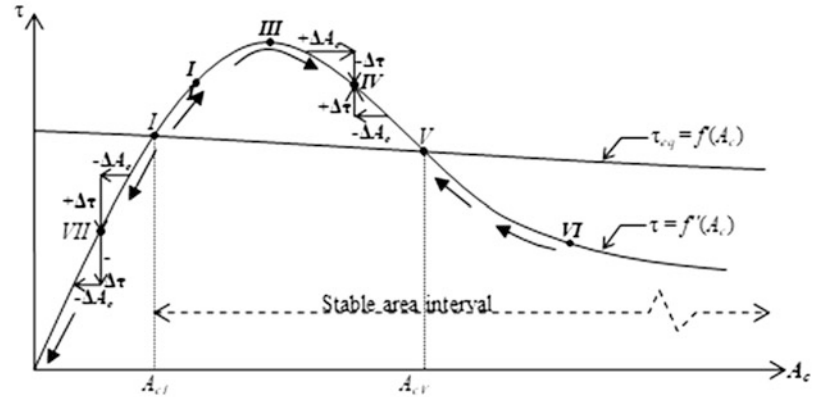
$$u_m = \frac{\pi P C_k \sigma}{2 A_c} \quad (11)$$

Combining with Eq. (8),

$$u_m = \frac{\pi C_k \sigma}{2 \Omega^{1/m}} A_c^{(1-m)/m} \quad (12)$$

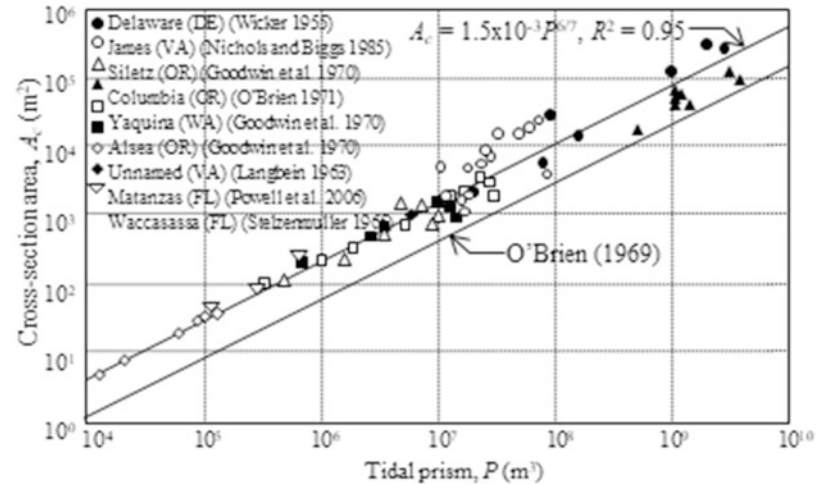
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Fig. 6 Diagram showing the essential concept of inlet stability and equilibrium



Morphodynamic Stability of Tidal Inlet-Bay Systems,

Fig. 7 Relationship between cross-sectional area and prism along US estuaries, and O’Brien’s (1969) relationship for inlets



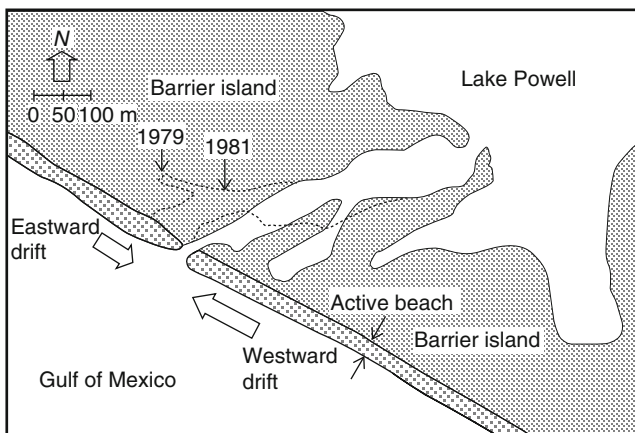
which is the desired relationship for sedimentary equilibrium. Finally, the equilibrium area, e.g., A_{cVI} in Fig. 6, is obtained solving Eqs. (6) and (12) simultaneously.

Effect of Tidal Range

Let us consider Phillips “Inlet,” an intermittently open channel connecting Lake Powell to the Gulf of Mexico (Figs. 1 and 8). The lake has an effective area of $1.20 \times 10^6 \text{ m}^2$, and the tide is mixed (semidiurnal + diurnal) with a mean period $T = 6.71 \times 10^4 \text{ s}$. During dry winter months, at spring tide (range 0.2 m) the channel is often open, but at neap tide (0.1 m) it tends to close. Thus the inlet goes through a spring-neap shift between stable and unstable states. Figure 9 shows a closed channel in 1979 and open in 1981. In the open channel, the current is strong enough to cut through the wave-active beach, along which littoral drift attempts to close the channel. At times of heavy rainfall, drainage from the lake enlarges the channel and increases its stability.



Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 8 Phillips Inlet and Lake Powell in 2017. (Adapted from Google image by Artesanohotels.com)



Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 9 Phillips Inlet closed in 1979 and open in 1981. (Based on Mehta 1985)

For very shallow inlet-bay systems some of the assumptions inherent in the hydraulic relationship are not fully met and the results tend to be somewhat qualitative. Yet the application of Eq. (6) can be insightful. Characteristic parameters for Phillips Inlet are $Q_s = 1.5 \times 10^{-5} \text{ m}^3 \text{ s}^{-1}$, $x_L = 385 \text{ m}$, $s = 2.65$, $g = 9.81 \text{ m s}^{-2}$, $\nu = 10^{-6} \text{ m}^2 \text{ s}^{-1}$, $p = 0.182$, and $q = 0.419$. Figure 10 shows the sedimentary curve of Eq. (12) and the spring/neap hydraulic curves without “river” discharge Eq. (6). During spring tide ($a_0 = 0.10 \text{ m}$), the equilibrium area is 38 m^2 . However, when the tide decreases to 0.05 m at neap, the two curves do not intersect; hence the channel is unable to remain open.

Effect of River Discharge

Even a small (fractional) increase in the tide-maximum ebb current velocity u_m by “river” discharge velocity $u_r = Q_r/A_c$ may enhance inlet stability. For that purpose a solution for u_m without discarding Q_r in Eq. (5) has been provided by Escoffier and Walton (1979) (see also van de Kreeke and Brouwer 2017). The result reverts to Eq. (6) for $Q_r = 0$. The investigators have employed parameterization relating the depth h_c to area A_c that is seemingly different from Eq. (1); however, it can be readily shown to be identical.

As an approximation of the Escoffier and Walton (1979) approach, when $u_r < u_m$, u_r may be simply added to u_m and the sum $u_{mf} = u_m + u_r$ plotted against A_c . Based on Eqs. (6) and (12), at Phillips Inlet the equilibrium area of 38 m^2 (Fig. 10) without Q_r corresponds to prism $P = 2.85 \times 10^5 \text{ m}^3$. Taking $Q_r = 0.7 \text{ m}^3 \text{ s}^{-1}$ (Mehta 1985), we obtain $2Q_r/PT = 0.0266$, which is only about 3% and has been ignored. However, with Q_r the hydraulic curve (Fig. 10) at neap tide yields an equilibrium area of 17 m^2 . In other words a stable channel can be maintained at a smaller cross section than in the absence of river discharge. Observe also that with continued decrease in area along the horizontal axis, the ebb velocity u_{mf} (expectedly) increases.

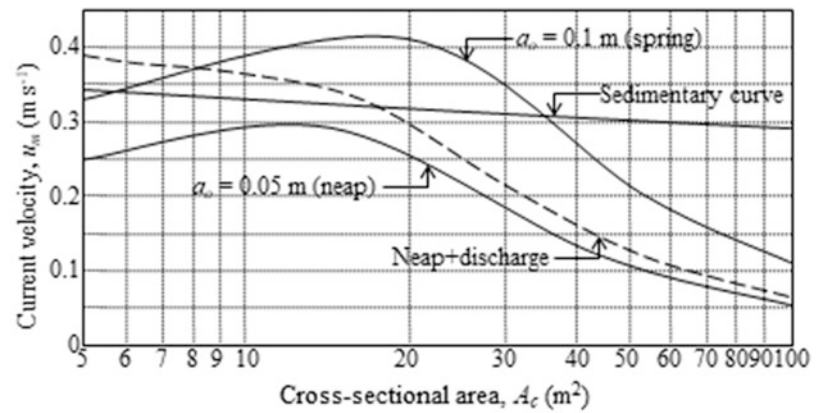
Effect of Sediment Load

In Eq. (9) Ω inversely varies with sediment load Q_s . This indicates that if the load were to increase or decrease naturally or by engineering, the resulting disequilibrium would change A_c . Thus, for instance, a decrease in Q_s will increase Ω , and, for given prism P , A_c will increase in accordance with Eq. (8).

Let Q_s be equal to $\alpha_s Q_{s0}$, where Q_{s0} is the sediment load before reduction and α_s is the fraction by which it is reduced. Consider Matanzas Inlet (Figs. 3a, b), where in 1964 the impact of Hurricane Dora led to a breach (Dora Cut) in Rattlesnake Island. Inasmuch as for the next 12 years until 1976, the inlet did not experience significant storm impacts, we may assume that the channel had achieved sedimentary equilibrium. For 1976 the relevant parametric values are $x_L = 3000 \text{ m}$; A_c (at equilibrium) = 820 m^2 ; $A_b = 3.5 \times 10^6 \text{ m}^2$; $a_0 = 0.75 \text{ m}$; $T = 44,712$; $Q_{s0} = 1.8 \times 10^{-4} \text{ m}^3 \text{ s}^{-1}$;

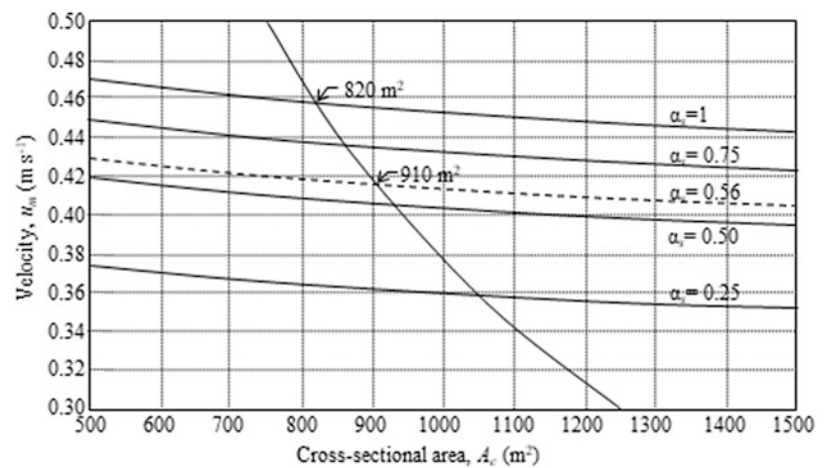
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Fig. 10 Stability diagram for Phillips Inlet without “river” discharge



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Fig. 11 Effect of sediment load reduction on area at Matanzas Inlet (Mehta et al. 2015)



$s = 2.65$; $v = 10^{-6} \text{ m}^2 \text{ s}^{-1}$; $p = 0.182$; and $q = 0.419$. The tidal prism $P = 6.2 \times 10^6 \text{ m}^3$ from Eq. (8) is close to the measured value of $5.2 \times 10^6 \text{ m}^3$ (Powell et al. 2006).

In 1977 Dora Cut was sealed by dredge filling with sand, and both ends of the fill were buttressed by rock revetments (Hayter and Mehta 1979). This change immediately decreased the sediment load in the inlet due to a reduction in u_m resulting from an effective decrease in the bay area. In Fig. 11 the hydraulic curve and the sedimentary curve for selected arbitrary values of $\alpha_s = 1, 0.75, 0.50$, and 0.25 intersect at increasing areas of 820, 864, 928, and 1047 m^2 , respectively, as decreasing load permits increasingly wide channel at equilibrium. At Matanzas, the time to reach new inlet equilibrium consistent with the reduced load took 19 years (1995), when the throat area widened to 910 m^2 . This increase corresponds to volumetric discharge $Q_s = 1 \times 10^{-4} \text{ m}^3 \text{ s}^{-1}$, i.e., $\alpha_s = 0.56$.

Effect of Second Inlet

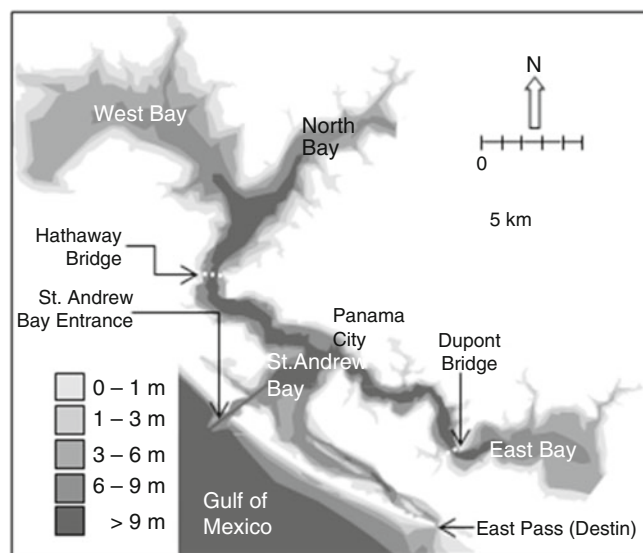
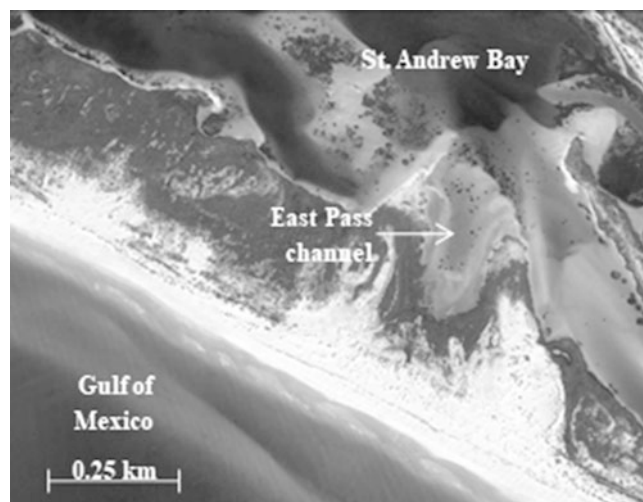
As in the case of the opening of Mathew Cut south of Matanzas Inlet (Fig. 3b), concern may arise when the newly formed second inlet alters the hydrodynamics of the first inlet. In numerous cases, such an occurrence has led to a

contraction and even closure of the first channel, while the new channel has widened. Such a process may occur over a period of months to decades depending on the tidal energy, wave energy, and availability of sediment. At Boca Ciega Bay on the Gulf of Mexico Coast, regularly dredged John's Pass continued to enlarge as Blind Pass contracted (Table 1, from Jain et al. 2004). Similarly, at St. Andrew Bay the dredged main entrance widened, while East Pass (Destin) (Fig. 12) contracted and finally closed in 1998. An effort was made to keep it open by dredging a small channel during 2001–2002. However, it remains closed (Fig. 13).

van de Kreeke and Brouwer (2017) extended the inlet-bay stability analysis to N -inlets connected to the same bay. Within the framework of assumptions for solving the problem analytically, it was found that theoretically only one of the inlets (or none) can be stable. Thus, in the case of two inlets ($N = 2$), either the old or the new inlet would be expected to close, although theory does not indicate which inlet will close, or how long it will take for this to occur. In reality the outcome depends on the prevailing morphodynamic conditions, driven to varying extent by the magnitude and frequency of storms as well as human intervention, all extraneous to the analytic model.

Morphodynamic Stability of Tidal Inlet-Bay Systems, Table 1 Inlet cross sections in Boca Ciega Bay and St. Andrew Bay

Year	Flow area (m ²)		Year	Flow area (m ²)	
	John's Pass	Blind Pass		St. Andrew Bay Entrance	East Pass
1873	4.74×10^2	5.38×10^2	1934	1.84×10^3	3.40×10^3
1883	4.32×10^2	4.96×10^2	1946	3.53×10^3	2.15×10^3
1926	5.31×10^2	2.09×10^2	1983	3.94×10^3	1.39×10^3
1941	6.36×10^2	2.25×10^2	1998	4.22×10^3	0
1952	8.49×10^2	1.57×10^2	2001 ^a	6.37×10^3	0
1974	8.83×10^2	4.11×10^2	2001 ^b	5.21×10^3	3.00×10^2
1998	9.50×10^2	2.30×10^2	2002	6.33×10^3	2.55×10^2

^aIn September; when East Pass remained closed^bIn December, when East Pass was reopened**Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 12** St. Andrew Entrance and East Pass near Destin**Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 13** Closed East Pass channel in 2017 (Google image)

The interpretation of N -inlet stability calculations is somewhat more complicated than that for a single inlet. Their application to St. Andrew Bay Entrance and East Pass are given elsewhere (Jain 2002; Jain et al. 2004). The bay area (74 km²) is effectively confined to waters between Hathaway and Dupont Bridges (Fig. 12); areas beyond these limits do not contribute significantly to the hydrodynamics of the two inlets.

In the phase plane diagram of Fig. 14a, the point (A_s , A_e) represents the flow cross-sectional areas of St. Andrew Bay Entrance (A_s) and East Pass (A_e) in 2001. Neither inlet can be considered to be in equilibrium, as the point lies outside the stable inlet domains of both inlets (van de Kreeke and Brouwer 2017). The history of the inlets bears out this finding – as mentioned, the seeming stability of St. Andrew Bay Entrance has in fact been sustained by regular dredging of the navigation channel since its opening in 1934.

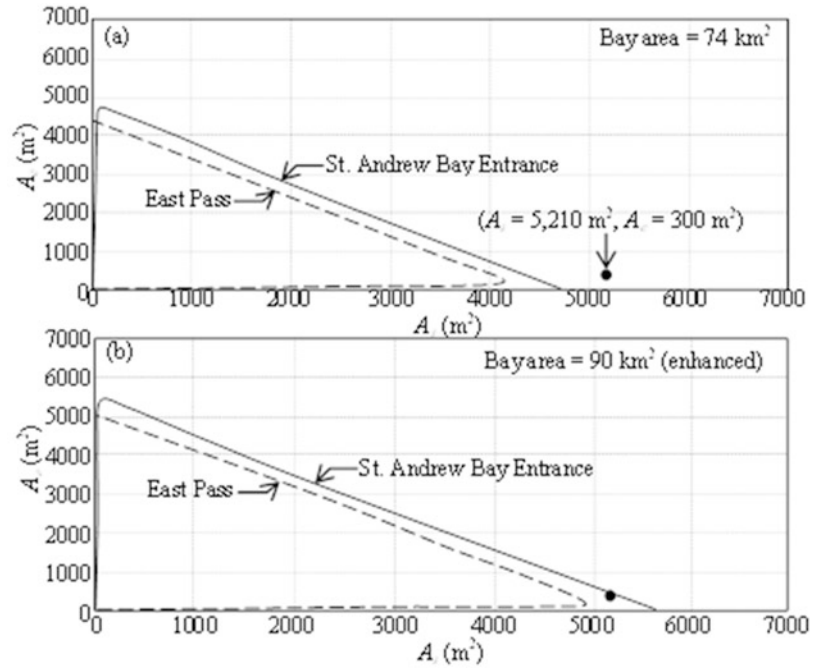
It is instructive to examine the potential effect of increasing the bay surface area by dredge deepening the channels under the two bridges. Deepening either channel would increase the bay tidal prism and enhance inlet stability. For instance, deepening the Dupont Bridge channel would increase the effective bay area from 74 to approximately 90 km² by adding the area of East Bay. The result in Fig. 14b implies that this would stabilize St. Andrew Bay Entrance but not East Pass (assuming that their cross sections do not change immediately due to bay enlargement).

Equilibrium Adaptation Time

Thus far we have considered equilibrium conditions at inlets. As mentioned, depending on the size of the system and hydrodynamics/sediment transport, inlets whose equilibrium is disturbed by storm impact or engineering, the time for adaptation to a new equilibrium can vary widely (O'Connor et al. 1991; Mehta et al. 2015; van de Kreeke and Brouwer 2017). For illustration let us consider a rough estimation of this time due to sediment load reduction mentioned earlier. Letting N_y denote years, the rate of widening of area A_c due to bed erosion at the onset ($N_y = 0$) of sediment load reduction can be stated as

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Fig. 14 Stability phase planes for St. Andrew Bay Entrance and East Pass: (a) existing condition with a bay area of 74 km²; (b) enhanced bay area of 90 km² (Jain et al. 2004; Jain 2002)



$$\left(\frac{dA_c}{dN_y}\right)_{N_y=0} = M(q_s - \alpha_s q_s) = Mq_s(1 - \alpha_s) \quad (13)$$

where M is the erosion rate constant and q_s is the volumetric sediment load per unit width of channel, i.e., Q_s divided by channel width (Julien 2010). Widening, initially at a rapid rate, gradually slows until A_c approaches a new equilibrium value A_{cf} . This approach can be expressed as

$$A_c = (A_{ef} - A_{ei})(1 - e^{-K_t N_y}) + A_{ei} \quad (14)$$

where subscript i denotes equilibrium before load reduction. The time constant K_t is obtained from Eqs. (13) and (14) as

$$K_t = \frac{Mq_s(1 - \alpha_s)}{A_{ef} - A_{ei}} \quad (15)$$

If after N_{yf} years the area reaches a value $(1 - \varepsilon) A_{cf}$, which is close to A_{ef} with $\varepsilon = 0.05$, a conveniently small fraction, we obtain

$$\frac{A_c}{A_{ci}} = \left(\frac{A_{cf}}{A_{ci}} - 1\right) \left[1 - e^{-\left\{\frac{1}{N_{yf}} \ln \left[\left(\frac{A_{cf}}{A_{ci}} - 1\right) / \left(\frac{A_{cf}}{A_{ci}} \varepsilon\right)\right]\right\} N_y}\right] + 1 \quad (16)$$

Finally,

$$N_f = \frac{A_{ci}(\hat{A}_{cf} - 1)}{Mq_s(1 - \alpha_s)} \ln \left(\frac{\hat{A}_{cf} - 1}{\hat{A}_{cf}} \varepsilon \right) \quad (17)$$

where $\hat{A}_{cf} = A_{cf}/A_{ci}$ is chosen for concise representation.

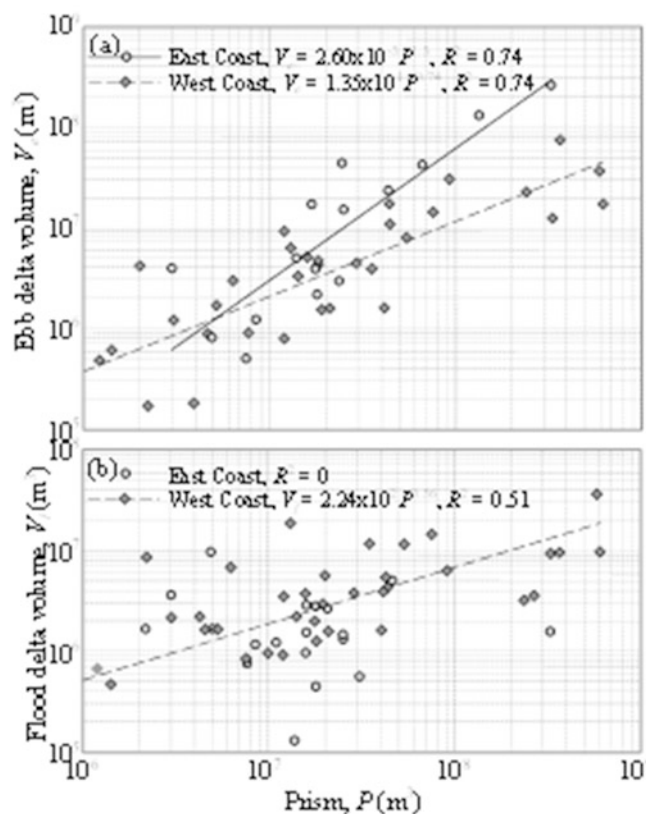
At Matanzas the parameters are $A_{ci} = 820 \text{ m}^2$, $h_{ci} = 2.8 \text{ m}$, $A_{cf} = 910 \text{ m}^2$, $\alpha_s = 0.56$, and $q_s = Q_s h_{ci}/A_{ci} = 6.14 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$. Accordingly, selecting $M = 1.24$ yields $N_f = 6 \times 10^8 \text{ s}$, which is equal to the measured 19 years (1976–1995).

Ebb and Flood Deltas

Empirical Relationships

Ebb and flood deltas at a new inlet eventually approach volumes which can be thought of as equilibrium values, even though sea level rise may sustain continued growth at a slow rate. In general, equilibrium implies that the volume would conform to a definable relationship with the tidal prism. As observed in Fig. 15a, ebb delta volumes V_e at Florida's Atlantic (East) and Gulf (West) coasts correlate positively with tidal prism P , as originally shown by Walton and Adams (1976). Along the East Coast, V_e increases rapidly with P relative to the West Coast. This is likely to be due to differences in the tidal range and to a degree due to wave energy flux (Table 2) between the two coasts, with the outcome that the coastal climate is more energetic along the East Coast.

Flood delta volumes at East Coast inlets are not well related to prism (Fig. 15b), to some extent due to limited accumulation of sand in the confined embayments, but to a greater degree due to the significant dredging activity at most inlets to maintain deeper than natural channels for navigation (Powell et al. 2006). As a result most flood deltas are not in equilibrium. Along the West Coast, a weak correlation does seem to exist. There as well the small bays constrain growth, and dredging takes place regularly in some channels.



Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 15 Delta volume against tidal prism at Florida's East and West coast inlets: (a) ebb deltas; (b) flood deltas. (Data are from Powell et al. (2006))

Morphodynamic Stability of Tidal Inlet-Bay Systems, Table 2 Representative physical parameters at Florida's East and West Coasts

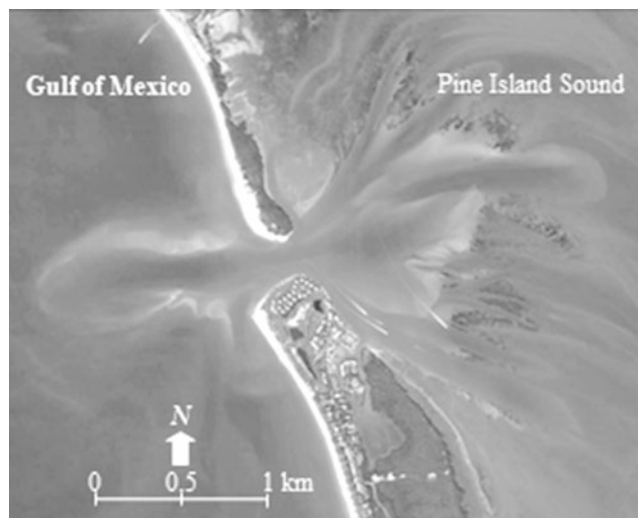
Parameter	East Coast	West Coast
Tidal range (m)	0.6–1.9	0.4–0.9
Wave energy flux ($\text{N m s}^{-1} \text{m}^{-1}$)	1.9×10^3 – 1.6×10^4	1.9×10^3 – 5.0×10^3
Grain diameter (mm)	0.2–0.7	0.1–0.5

An exception is the flood shoal at Redfish Pass in Pine Island Sound (Fig. 16), which shows an “explosive” growth pattern seemingly unencumbered by the bay morphology.

Powell et al. (2006) provide a phenomenological explanation of the dependence of ebb delta volume V_e on P . The general equation can be given as

$$V_e = a_e P^{b_e} \quad (18)$$

where a_e and b_e are empirical coefficients (Figs. 15a, b). The investigators consider the delta (Fig. 17a), characterized by a swash platform and a lobe face, to be a box of volume V_e . For this simple geometry, it is feasible to deduce Eq. (18) with exponent $b_e = 1.44$. This inference is approximately consistent with the power 1.30 for East Coast ebb deltas (Fig. 15a).

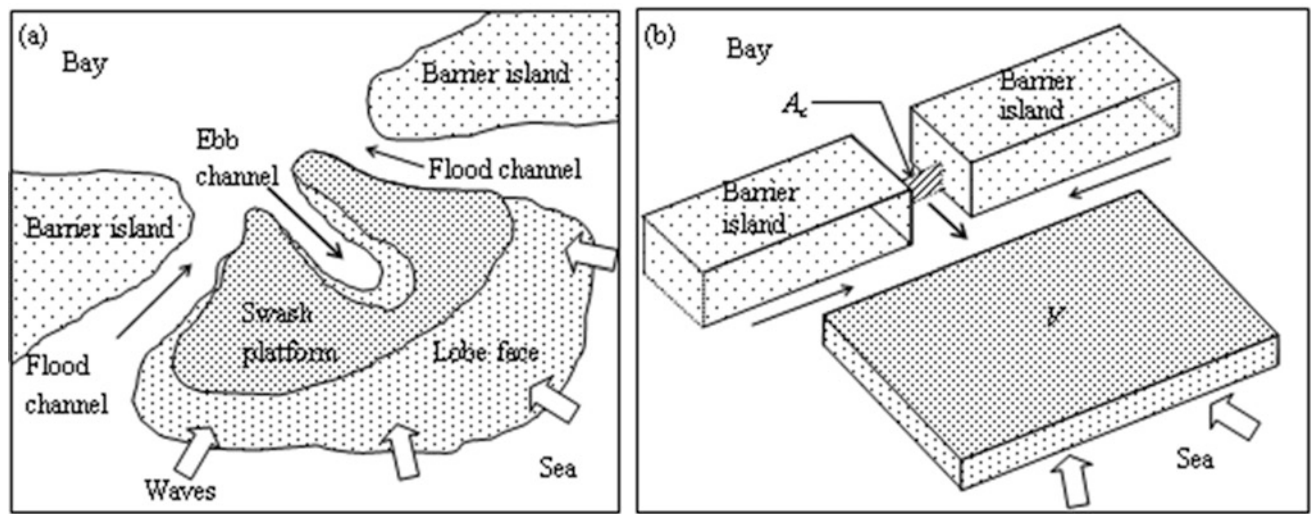


Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 16 Ebb and flood deltas at Redfish Pass on the West coast of Florida in 2017. (Adapted from Google image)

Based on data including several inlets along all three coasts of the USA, and selecting b_e to be 1.23 in all cases, Walton and Adams (1976) proposed the effect of waves on ebb delta volume in terms of the magnitude of a_e (Table 3), which indicates that V_e decreases with increasing wave energy, or more approximately with wave height. This is evident from Fig. 18 in which V_e for Jupiter Inlet, East Pass (Carabelle), and South Lake Worth Inlet (also known as Boynton Inlet) measured in summer are plotted against the maximum offshore wave height H_m during the preceding winter when the highest, volume-impacting, waves occurred.

In the microtidal environment, e.g., Florida, the response of ebb delta to waves appears to indicate that to a fair extent they govern the position and profile of the lobe face (Fig. 17a). The face is characteristically concave in shape, and downward from the edge of the swash platform appears to resemble a typically concave beach profile. Devine and Mehta (1999) demonstrated this similarity for the ebb delta at Jupiter Inlet and showed that the delta's response to waves can be “modeled” by assuming that the waves approach the lobe face radially. Figure 19a shows the idealized delta making use of the beach profile equation proposed by Dean and Dalrymple (2002). Parameters used to generate this morphology are based on the ebb delta at Jupiter Inlet. Modeling the lobe face as radially offshore-directed “beach” profiles makes it convenient to examine the effects of storm waves as well as surge on delta volume. This is illustrated in Fig. 19b, in which the curves have mainly qualitative significance, given the idealized morphology of the delta and the selected mode of lobe face response.

Although empirical modeling is typically concerned with “static” or equilibrium relationships, the above example

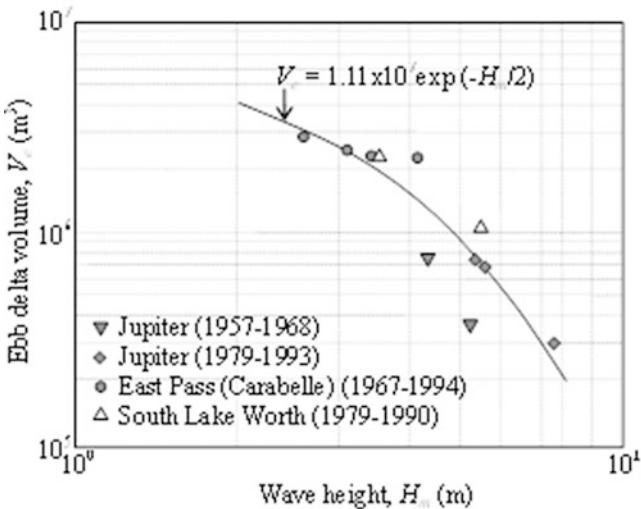


Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 17 (a) Ebb delta features; (b) idealized ebb delta. (Revised from Powell et al. 2006)

Morphodynamic Stability of Tidal Inlet-Bay Systems, Table 3 Wave exposure and a_e values^a

Exposure to waves	a_e
Mild	8.24×10^{-6}
Moderate	1.04×10^{-5}
High	1.45×10^{-5}

^aBased on Walton and Adams (1976)



Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 18 Variation of ebb delta volume at three inlets with annual maximum wave height. (Based on Devine and Mehta 1999)

shows that dynamic effects can be accounted for in an approximate way. Efforts have also been made to extend empirical analysis to some inlet-related transport phenomena. Kraus (2000) developed a semiempirical “reservoir model” involving a macroscale sediment budget to describe the way in

which littoral sand moves from one side of the inlet channel to the other over the ebb delta. Application of the model to Ocean City Inlet, MD, showed reasonable agreement with resulting delta growth over five decades.

Process-Based Modeling

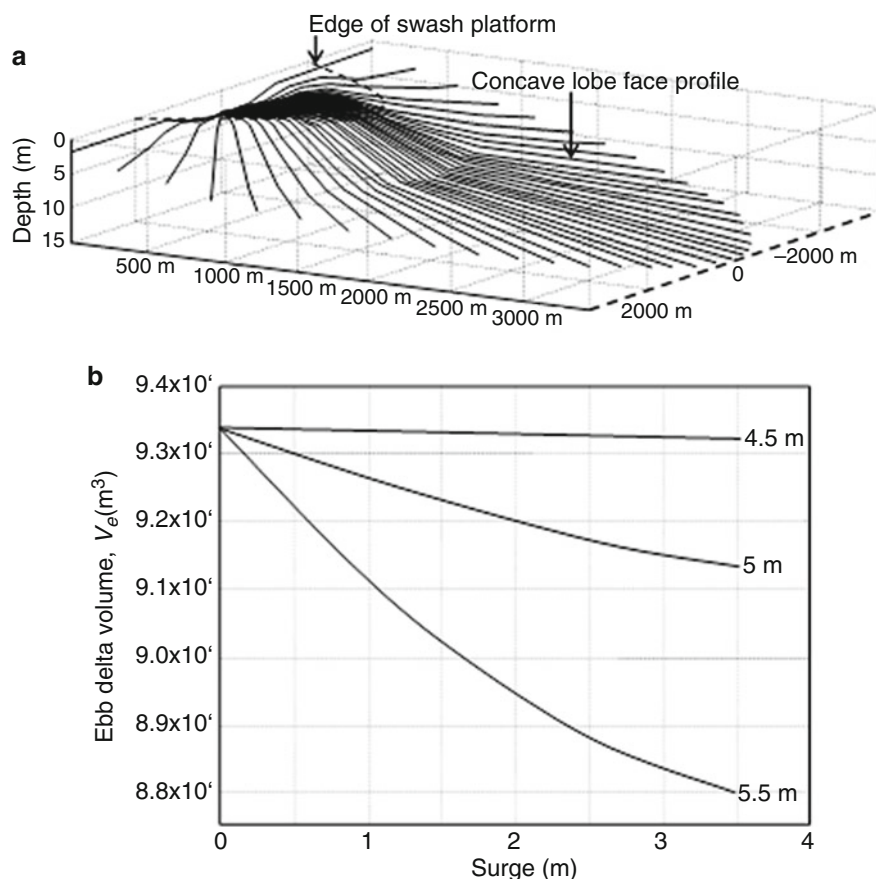
Modeling approach Inasmuch much as the role of physical processes is central to coastal morphodynamics, process-based modeling relies on the equations of flow momentum, continuity, and sediment mass balance in their finite-difference forms involving step changes in the space and time coordinates. In many cases, effects of wind, vegetation, and biology must be incorporated, although, on account of biodiversity, modeling is required to be significantly tailored to site specific applications.

Beginning with given initial morphology, process-based model computations are repeated every time-step (Fig. 20) for the desired duration, e.g., a decade. Accordingly, the morphology at the start of n^{th} ($n = 1, 2, 3, \dots$) time-step is subject to the effect of tide and waves followed by sediment transport for the duration of the time-step (e.g., 1 s to 1 h). The changed morphology (usually with respect to the bathymetry) at the end of the time-step is returned, or fed back, to the morphology at the beginning of the $n + 1$ th time-step and so on.

An important feature of feedback is the use of the so-called morphological factor $MF (\geq 1)$ to speed up the computation, inasmuch as, for example, for a 1-h time-step, the number of steps required to reach 10 years would be a high value of 3.15×10^8 . The selection of MF value requires judgment based on experience, as changing MF may bring about an undesired change in the final morphodynamic form, particularly due to nonlinear instabilities (Roelvink and Reniers 2010; van Maanen et al. 2011, 2013). To further reduce

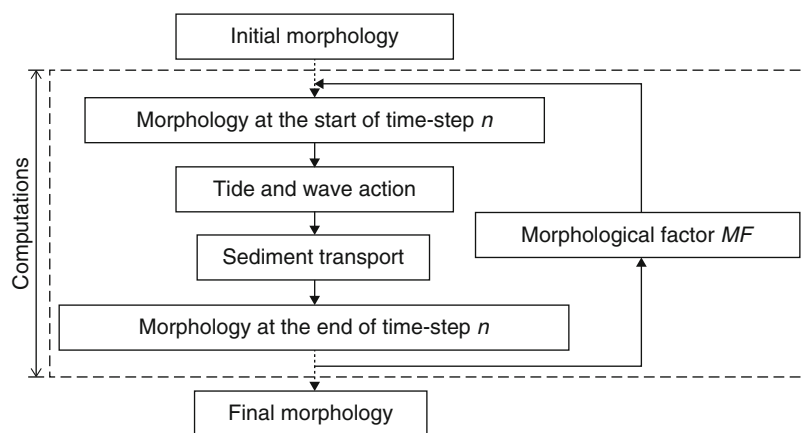
Morphodynamic Stability of Tidal Inlet-Bay Systems,

Fig. 19 (a) Idealized morphology of a microtidal ebb delta; (b) hypothetical ebb delta response to storm surge and wave height. (From Devine and Mehta 1999)



Morphodynamic Stability of Tidal Inlet-Bay Systems,

Fig. 20 Processed-based protocol for numerical modeling



computation time, simplification of calculations by ignoring the inertia term in the momentum equation is sometimes carried out. However, at the same time the approximation introduced by the use of MF in the first place must be recognized (e.g., D'Alpaos et al. 2005).

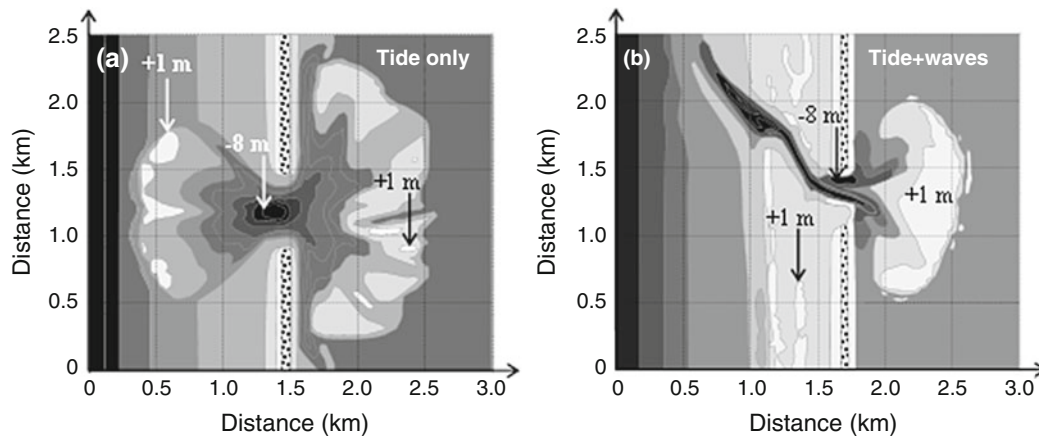
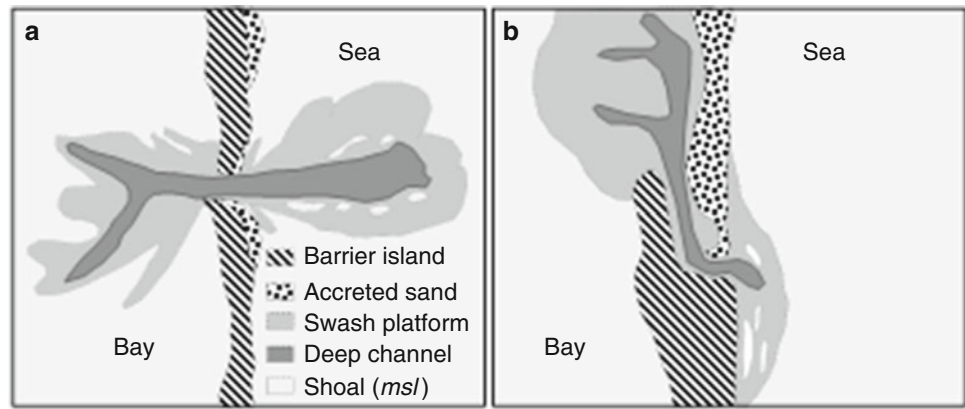
Flood delta simulation Based on ebb and flood deltas along the peninsular Gulf coast of Florida, Gibeaut and Davis (1989) identified generic morphologies of tide-dominated and wave-dominated deltas (Fig. 21). Subsequently, Tran et

al. (2012) examined the development of ebb and flood deltas at a newly opened (hypothetical) inlet in numerical experiments using the depth-averaged version of the hydrodynamic model Delft3D and the wave model SWAN (Lesser et al. 2004; Booij et al. 1999). The initial bathymetry was two-dimensional with shore-parallel depth contours on the seaward and bayward sides of the barrier island. The initial depth of channel cross section was 2 m, and the area was 850 m^2 .

Figure 22a shows the evolved ebb and flood deltas (to a height of about 1 m relative to msl) after 10,000 days

Morphodynamic Stability of Tidal Inlet-Bay Systems,

Fig. 21 Morphologies of microtidal inlet deltas: (a) tide-dominated; (b) wave-dominated. (Based on Gibeaut and Davis 1989)



Morphodynamic Stability of Tidal Inlet-Bay Systems,
Fig. 22 Numerical experiments on the development of ebb and flood deltas: (a) with tide only; (b) tide + waves. Descriptive rendition of

numerical results of Tran et al. (2012) reported in van de van de Kreeke and Brouwer (2017)

(27.4 years) of simulation. In that time the maximum depth increased to about 8 m. As the deltas were formed by eroded sand within the modeled system, sand mass was conserved. Figure 22b shows a simulation inclusive of the effect of waves run for 800 days (2.2 years) starting with a narrow channel (depth 2 m, area 250 m²). The waves were from the NE at an angle of 25° relative to the north. The development of the ebb delta was rapid due to waves and resulted in a southward skewed channel.

Offshore dispersal Morphodynamic modeling is commonly used in wider applications of engineering interest. Consider an analysis that is based on the flow momentum and sediment mass balances for predicting littoral sediment movement (Dean and Dalrymple 2002). Modeling bathymetric changes, along with a feedback for depth change every time-step, requires formulas for cross-shore as well as alongshore movements of particles due to waves and current. With respect to the elemental flow control volume in Fig. 23, consider the cross-shore coordinate x to be the dependent variable and

water depth h as the independent variable. The depth-mean mass balance is

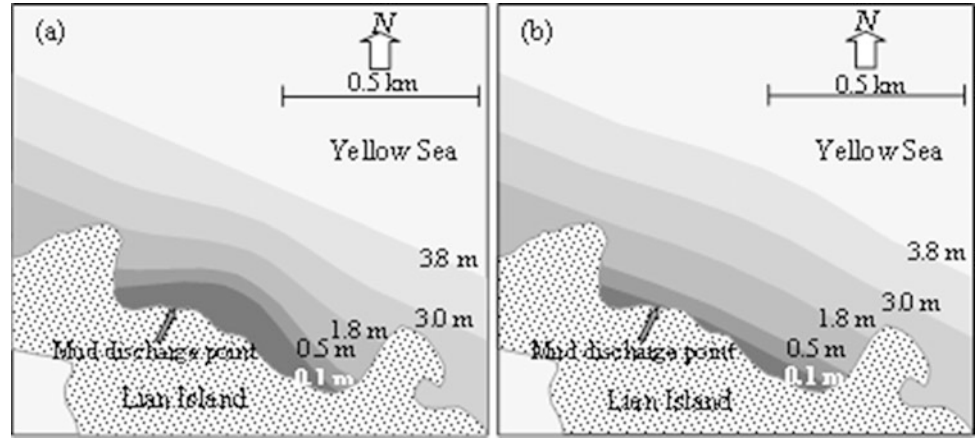
$$\frac{\partial x}{\partial t} = -\frac{1}{m_s \rho_D} \frac{\partial g_{sy}}{\partial y} - \frac{1}{\rho_D} \frac{\partial g_{sx}}{\partial h} \quad (19)$$

where g_{sx} and g_{sy} are the components of suspended sediment mass flux per unit length in the x - and y -direction, respectively, ρ_D is the dry density of the bottom deposit, and $m_e = \partial h / \partial x$ is the local bottom slope. This form of the equation permits direct tracking of the isobaths, i.e., depth contours, when they are displaced from their initial positions due to erosion or deposition. Thus, interpolation for identification of any isobath position, as would be essential when using a Cartesian (x, y) grid, is not required. The output at every time-step is a bottom contour “map” dependent on the cross-shore and alongshore sediment fluxes.

Consider an application related to the disposal and subsequent dispersal of dredged mud. In this case Eq. (19) is modeled together with formulas for the fluxes g_{sx} and g_{sy}

Morphodynamic Stability of Tidal Inlet-Bay Systems,

Fig. 25 Mud deposit simulations at Lian Island: (a) immediately after end of disposal; (b) 2 months later. (Revised from Rodriguez and Mehta 2001)



Morphodynamic Stability of Tidal Inlet-Bay Systems,

Fig. 26 Tidal flat and (eroding) salt marsh near Matanzas River in 2017 (Google image)



the bed elevation difference can be explained in terms of a stability analysis similar in principle to that of Escoffier (1940) for tidal inlets. Presently we are interested in water depth below *msl* over the flat. In analogy with inlets, the equilibrium depth is determined by the coupled solution of a hydrodynamic relationship and a sedimentary equilibrium relationship.

A simple way to estimate the bed shear stress τ_b is by combining the wave- and the current-induced shear stresses, such as by using the empirical method of Soulsby et al. (1993). Here, solely for illustrating the analysis, we will assume that τ_b can be estimated merely from the wave-induced peak velocity u_w (by considering the site to be wave-dominated). Thus

$$\tau_b = C_D \rho_w \frac{u_w^2}{2} \quad (20)$$

where C_D is the drag coefficient at the bed. From the Airy wave theory (e.g., Holthuijsen 2007), we obtain

$$u_w = \frac{H\sigma_w}{2\sinh kh} \quad (21)$$

where H is the wave height, σ_w is the wave frequency, h is the water depth, and k is the wave number. The latter is obtained from the wave dispersion relationship

$$k = \frac{\sigma_w}{\sqrt{\frac{g}{k} \tanh kh}} \quad (22)$$

As a wave approaches the salt marsh, its maximum height is limited by breaking. With κ as the wave-breaking index, we can make use of the well-known relationship

between the wave height and the water depth at the point of breaking

$$H = \kappa h \quad (23)$$

Thus

$$\tau_b = \frac{C_D \rho_w \kappa^2 \sigma_w^2 h^2}{8 \sinh^2 kh} \quad (24)$$

The required sedimentary equilibrium relationship is essentially the statement of the bed shear stress τ_{eq} at equilibrium, i.e.,

$$\tau_{eq} = C_e \quad (25)$$

in which C_e is determined by balancing erosion E and deposition D fluxes over the flat. These can be defined as

$$E = M_e (\tau_b - \tau_{cr}) \quad (26)$$

where M_e is the erosion flux constant, and

$$D = \rho_w w_s \phi \quad (27)$$

in which w_s is the mud particle/floc settling velocity and ϕ is the sediment volume fraction (Winterwerp and van Kesteren 2004). The condition for equilibrium is

$$E = D \text{ @ } \tau_b = \tau_{eq} \quad (28)$$

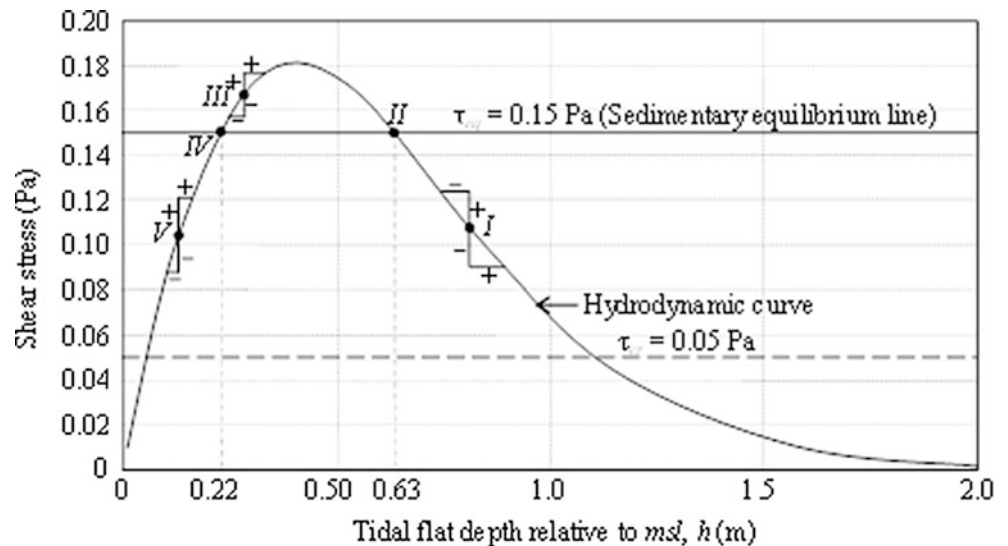
Thus

$$\tau_{eq} = C_e = \tau_{cr} + \frac{\rho_w w_s \phi}{M_e} \quad (29)$$

Consider characteristic parametric values: $\sigma_w = 1.57 \text{ rad s}^{-1}$ (wave period 4 s), $\phi = 1 \times 10^{-4}$, $w_s = 0.01 \text{ m s}^{-1}$, $\tau_{cr} = 0.05 \text{ Pa}$, $\rho_w = 1000 \text{ kg m}^{-3}$ (nominal), $C_D = 0.01$, $\kappa = 0.9$ (Holthuijsen 2007), and $M_e = 0.01$. For cohesive floes the settling velocity w_s in Eq. (27) is conventionally multiplied by the so-called probability of deposition $p_d (\leq 1) = 1 - (\tau_b/\tau_d)$, where τ_d is called the critical shear stress for deposition introduced by Krone (1962) based on laboratory flume experiments. For the present purpose, the simplest approach is to assume $p = 1$, which permits settling at all shear stresses τ_b , while erosion occurs when τ_b is greater than τ_{cr} . For erosion and deposition over sandy flats, equilibrium leads to the well-known bed load and suspended load formulas (Julien 2010). However, Eqs. (27), (28), and (29) remain as reasonable approximations.

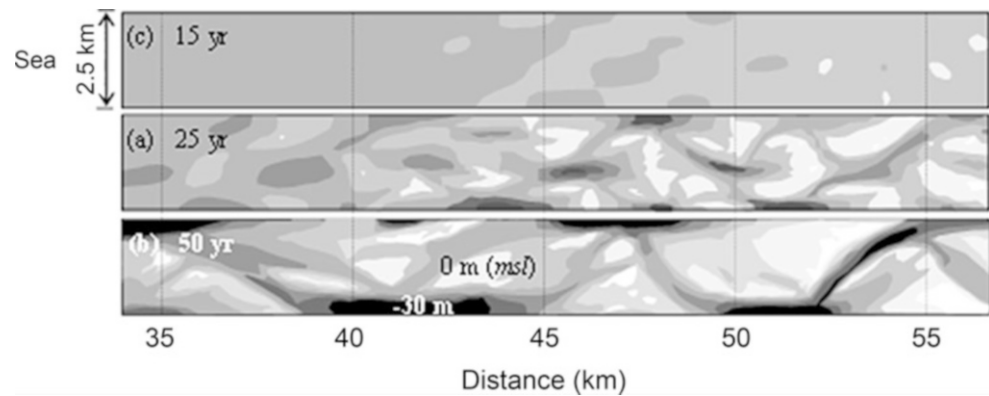
The resulting plot in Fig. 27 is analogous to Fig. 6 for inlets. It defines the threshold, a kind of “tipping point,” for development of a salt marsh table from the tidal flat due to naturally occurring positive and negative fluctuations in the bed shear stress. Mudflat depth h (below msl) at point I is a stable value because a slight increase (+) in τ_b would decrease (−) h and thereby nullify the effect of increase in τ_b . Similarly, a decrease (−) in τ_b would be countered by increasing (+) h . The actual depth of 0.63 m of the flat is determined by the intersection of the hydrodynamic curve and the sedimentary line. At point III an increase in τ_b would increase h due to erosion because $\tau_b > \tau_{eq}$. This in turn would further change h until III becomes coincident with II . Point III could also move to the left due to a decrease in τ_b , hence in h . At point IV , h would be 0.22 m and any further decrease in τ_b would cause h to “slide” to zero, i.e., the mudflat would sustain a salt marsh coincident in height with msl . Thus, for instance, mudflat at $h = 0.63 \text{ m}$ will be conducive to the formation of a salt marsh provided deposition of sediment,

Morphodynamic Stability of Tidal Inlet-Bay Systems,
Fig. 27 Diagram defining the threshold for sustenance of salt marsh at a tidal flat



Morphodynamic Stability of Tidal Inlet-Bay Systems,

Fig. 28 Simulated sand bed in hypothetical tidal canal after: (a) 15 year; (b) 25 year; (c) 50 year. (Based on Hibma et al. 2003)



say during a storm, raises the bed by 0.41 m to $h = 0.22$ m. If the amount is less, the deposit will erode back to the flat at $h = 0.63$ m.

Shoal-Channel Cells

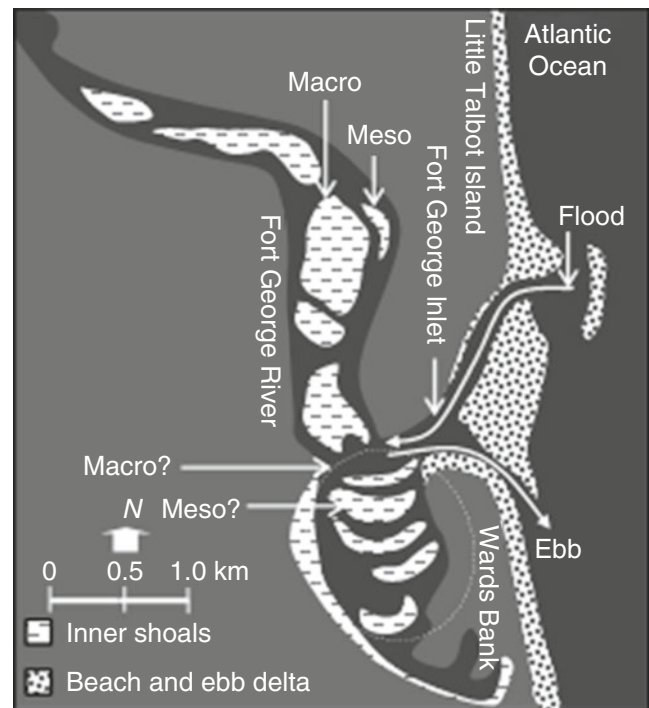
Shoal-and-Channel Development

The morphology of a combination of intertidal shoal and channels ubiquitous in estuaries tends to be surprisingly stable, although it may take years to develop into a mature form. To that end, starting with a 2.5 km wide and 80 km long hypothetical tidal canal with a 0.24 mm sand bed, Hibma et al. (2003) carried out numerical computations forced by a 1.75 m amplitude tide at the canal mouth (sea) to demonstrate the development of a realistic shoal-and-channel system at the end of 50-year simulation (Fig. 28). The initial depth was 15 m at the seaward end and 0 (msl) at the landward end. The three panels from the mid-reach of the canal indicate bed evolution at 15, 25, and 50 years. Each shoal (detached from the bank of the tidal channel) is flanked by a likely flood-dominated channel on one side and ebb-dominated one on the other. We will consider the stability of such a shoal-channel cell next.

Shoal-Channel Stability

Along the northeastern coast of Florida, nearly 2 m (i.e., nearly meso-)tide coupled with persistent northeasterly waves during the winter months, intertidal shoals, and meandering channels shows morphology that has persisted for decades. An example is Fort George River (Fig. 29), whose entrance at Fort George Inlet was configured by the construction of a long jetty (south of Wards Bank) at the mouth of the St. John's River (not shown) during the 1880s (Marino et al. 1990). Although the river and the system of shoals and channels have since migrated northward, shoal morphology has persisted.

For a likely explanation of such sustained shoal-and-channel systems, Winterwerp et al. (2001) developed a

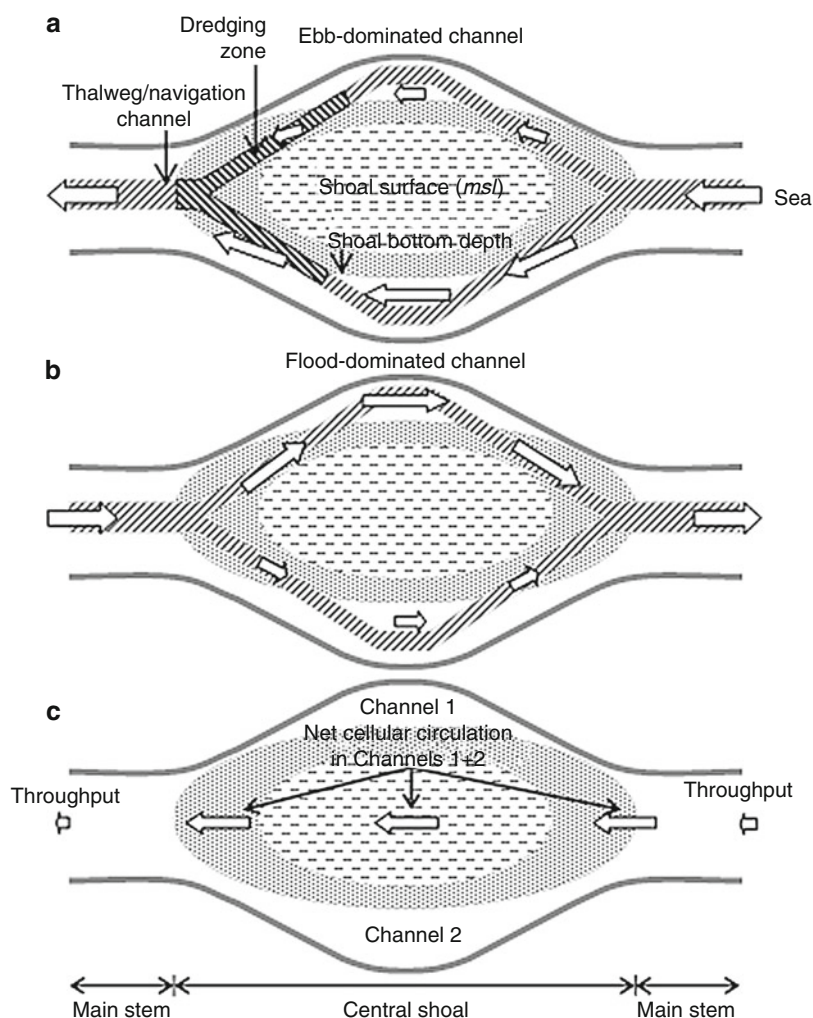


Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 29 Fort George River and entrance along the northeast coast of Florida

phenomenological model for the Western Scheldt estuary flowing through Belgium and the Netherlands. In the navigation channel of the estuary, heavy shoaling seems to occur at specific sites. To maintain navigable depth and at the same time comply with the legal requirement of no net change in the sediment inventory of the estuary as a whole, material dredged from the shoaled sites is disposed in waters elsewhere within the same system. Experience has shown that there is an upper limit of dredging, amounting to no more than about 10% of the gross sediment transport capacity, above which the shoal-channel system tends to destabilize. To explain this, the cell system in Fig. 30 was proposed as the basic morphologic unit whose stability is a matter of interest.

Morphodynamic Stability of Tidal Inlet-Bay Systems,

Fig. 30 Shoal-channel morphologic cell: (a) cell during tidal flood flow; (b) cell during tidal ebb flow, (c) net throughput in the main stem channel. (Based on Winterwerp et al. 2001)



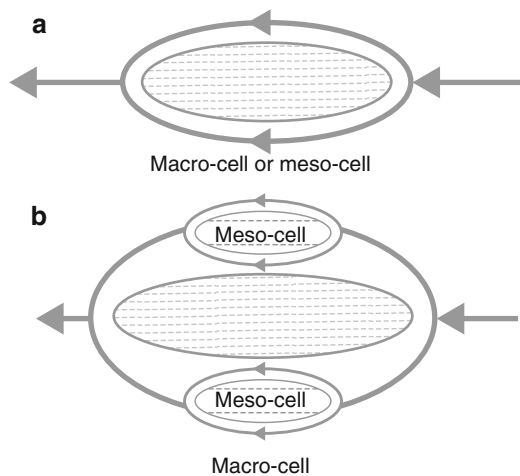
In Fig. 30a, the cell consists of a single channel (main stem) with a bifurcation leading to two sub-channels, characteristically a flood-dominated one and an ebb-dominated one, and a central shoal with surface at *msl*. At the base of the sloping periphery of the shoal, the depth is equal to that of the main stem. The navigation channel, which is coincident with the thalweg (deepest part of channel), experiences shoaling (one zone is shown), which requires dredging. The sediment transport vectors point in the direction of flood flow, which is stronger in the lower flood-dominated channel than in the upper ebb-dominated one. Figure 30b indicates the flow pattern during ebb, and Fig. 30c shows the long-term sediment throughput (discharge), directed upstream in this system. It leads to the formation of upstream shoals, as in Fort George River (Fig. 29).

In general, at a tide-dominated estuary such as the Western Scheldt, within the shoal-channel cell, the net (of flood and ebb) sediment discharge tends to be considerably higher than the throughput. As a result, the stability of the cell as a whole can be examined with respect to net circulation around the

shoal without considering throughput (Winterwerp et al. 2001). When excessive dredging is carried out, the cell is disturbed; hence the cell's morphodynamic stability is a matter of concern.

A cell may be classified as either large (macro) or small (micro) (Fig. 31a). A macro-cell consisting of meso-cells in a series is potentially unstable because, if for any reason, e.g., the impact of a significant river outflow or inflow due to storm surge, one meso-cell destabilizes, the entire macro-cell will eventually deteriorate. In contrast a macro-cell with meso-cells in parallel (Fig. 31b) would not necessarily break up the entire cell if one cell were to do so. By the same token, three meso-cells in parallel could retain overall stability even if two were to break up. In Fort George River (Fig. 29), the crescentic shoals eastward of Wards Bank possibly form a macro-cell with five meso-cells.

Jeuken and Wang (2010) discuss the basis of stability calculation for flood/ebb sub-channels (1 and 2 in Fig. 30c) relative to the outcome of dredged material disposal in one of the two channels. Numerical computations indicated that, in



Morphodynamic Stability of Tidal Inlet-Bay Systems, Fig. 31 (a) Macro- or meso-cell; (b) macro-cell with two meso-cells in parallel. (Based on Winterwerp et al. 2001)

qualitative analogy with the stability of two tidal inlets at a bay, if the sediment disposed is small compared to the gross sediment transport capacity, the decrease in channel depth is effectively countered by an increase in the current velocity which eliminates the excess deposit. However, if the 10% limit is exceeded, the channel tends to clog up and leads to degradation of the entire cell.

Summary

Sandy tidal inlets exist along extensive stretches of all three US coastlines. Many such inlets and the back-barrier bays show remarkably stable morphology in spite of the secular rise in sea level. Coastal morphodynamics includes the process by which a loose-boundary landform including the tidal inlet-bay system achieves a stable equilibrium configuration or form. In general, the relationship between process and form can be described by empirical formulas, by simple analytic models, and by process-based numerical modeling. The most important aspect of morphodynamics is a feedback between process and form as the equilibrium-restoring mechanism.

Conditions under which inlet-bay systems achieve equilibrium configuration due to tide and waves have been reviewed following the basic model of tidal inlet stability originally developed by Escoffier (1940) and subsequently interpreted by, among other, van de Kreeke and Brouwer (2017). This basis is also relevant to the relationship between tidal flats and salt marshes and the formation of stable morphologic cells in watercourses with intertwined shoals and channels. As case studies several inlet-bays in the state of Florida have been selected. Overall, in many engineering applications empirical as well as analytic modeling provides valuable answers regarding stable equilibrium morphology.

Although process-based modeling remains an imperfect tool, it is expected that with increasing speed of computation, the ability to model more complex systems over longer times than at present will continue to expand.

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Muddy Coasts

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Muddy coast is defined as a coastal depositional environment which exhibits muddy sediments as a major component of the sedimentary morphodynamic system. These morphodynamic deposits possess textural characteristics containing a high proportion of silt and clay, and exhibit identifiable sub-tidal, intertidal, and supra-tidal stratigraphy. Such deposits tend to form extensive low-gradient morphological surfaces, and are often manifest as broad intertidal flats, colloquially termed “mudflats.”

Terry R. Healy: deceased.

The visually obvious muddy coastal sedimentary deposits and geomorphic forms occur mainly within the intertidal zone of coastal fringe waters, and are clearly evident as surficial deposits forming tidal flat and drainage channel deposits. Muddy coastal depositional environments also occur in near-shore sub-tidal locations characterized by relatively low energy hydrodynamic conditions, such as in shallow estuaries and embayments. But direct marine processes influencing muddy deposition extend inland as far as storm surge tides are effective, and especially in low-gradient coasts, some muddy coast deposits occur on land areas above the normal high-tide level.

In principle, silt and clay particles comprising the suspended sediment load in coastal waters will deposit (1) if there is sufficient particle concentration such that enhanced deposition takes place, (2) if the suspended load particles are advected into quiet waters where the fine particles gradually sink below wave agitation and turbulence levels and deposit on the bottom, and (3) where salt wedge and “estuarine circulation” mixing processes at the head of estuaries induce flocculation of the fine particles.

Definition of “Mud”

In sedimentological terms “mud” is easily defined. Texturally “mud” is a detrital deposit composed of particle sizes smaller than 63 μm (i.e., 0.063 mm), and includes the textural classes of *silt*, 0.063–0.004 mm or 6ϕ – 10ϕ in the sedimentological (ϕ) classification, and *clay*, which is generally accepted as being composed of particle sizes smaller than 0.004 mm (4 μm) or smaller than 109 (Folk 1968; Leeder 1982).

However, muddy coastal deposits almost always contain a proportion of organic matter – typically ~3–5% – originating from the importation of fine suspended sediments into the coastal zone of the muddy sediment deposition. The organic matter includes both terrigenous (from fine vegetation detritus) and marine biogenic matter, as well as originating from in situ biogenic processes, such as mucus and feces from worms and other benthic organisms inhabiting the surficial muds. In many locations organic matter contributions include fine plant detritus originating from such diverse sources as mangrove stands, or adjacent “pastures” of sea grass, marsh grasses, and wetlands. Biogenic contribution to muddy sedimentary deposits may also include small shell fragments, minute sea urchin spines, and fragments of living and dead diatoms, foraminifera, ostracods, and coccoliths (Augustinus 2002; Wang et al. 2002a).

Additional mineral matter occurs in the mud deposits, including fine sands, terrigenous matrix fragments, and shelly gravel fragments. Composition of the mineral matter is influenced by local supply and may consist of quartz, feldspar and mica, and small proportions of heavy minerals.

Typically, mud deposited in the intertidal zone is soft, pliable, of high plasticity, thixotrophic, and contains a large volume of water within its physical structure. Practically, a person may sink up to their knees in wading through intertidal muddy deposits. This thixotrophic property suggests that in geotechnical terms mud behaves either as a plastic or a fluid substance indicating that mud deposits exhibit properties rather different from noncohesive clasts in the sedimentary environment.

In mud-rich shallow waters of muddy coastal areas, the Seafloor may be characterized by “fluid mud” – a state in which the particles occur partially as concentrated colloidal state, and the viscous substance possesses extremely high water content (Mehta and Hayter 1989; Mathew and Baba 1995).

Muddy coastal sediments also possess interesting geochemical characteristics, related to the chemical reactions associated with anoxic conditions and microorganism activity within the mud deposits. Below the oxic surface layers, usually only a few centimeters thick, the mud often looks blue-black and exhumes a pungent sulfurous smell.

Morphodynamic Types of Muddy Coast

For expansive open ocean tidal flats, for example, of China, muddy deposits are an integral component. The muddy deposits may range from several hundred meters to more than 10 km in extent (Wang et al. 2002b). But muddy coastal deposits may also exist in restricted lenses a few tens of meters wide. Wang et al. (2002a) identifies several morphodynamic categories of muddy coast:

Tidal flats: They are the typical geomorphic form of expansive muddy coasts and are characterized by intertidal surfaces with low slopes of order 0.5–1.0:1000. Distinct morphological and sedimentary zonation features can be identified (Wang et al. 2002b). Generally, a low relief salt marsh zone is developed within the supra-tidal zone, which may be surmounted by shelly chenier beaches or ridges, or artificial stopbanks (dikes) on populated flat lowland coasts. Below high-tide level, true “mudflats” are located in the upper part of the intertidal flat. Clay and silt layers are often developed in the central parts of the tidal flats, and the lower-to-sub-tidal sectors become more sandy-silty or silty-sand, with wave ripple bedform features developed on the surface. Dendritic meandering tidal creeks are common on broad muddy tidal flats, formed by the scouring of ebb-tidal waters as they drain off the large areas of intertidal flat upon the falling tide. Erosion scour features, both tidal current and wave induced, are often observed on the central sectors (Wang et al. 1990). For broad muddy tidal flats, spatially varying morpho-sedimentary zonation is evidently associated with different hydrodynamic forcing processes acting on the flat.

Inner deposits of barrier-enclosed lagoons: Within the landward side of barrier-enclosed lagoons, along the south-east coast of the United States, for example, a veneer of deposited mud is typically found. Such muddy deposits are typical of barrier lagoons, and may be associated variously with wetlands, oyster beds, and estuarine reentrant valley head deposits. Source of the muddy deposits is mainly from input from the surrounding land catchment, and the deposits are usually thin (0.5–2 m) but areally extensive in the available area, and overlies sands of the barrier lagoonal system.

Enclosed sheltered bay deposits: Muddy deposits of varying types may occur in sheltered embayments. Geomorphically, such embayments may be tectonically induced by block faulting, or form from drowned river valleys consequent upon the Holocene sea-level transgression. The deposits normally originate from the surrounding catchments and occur particularly in the tributary valley heads.

Estuarine drowned river valley deposits: Most funnel-shaped estuaries are drowned river valleys, which have significant freshwater input and within which there is strong mixing between the fresh and saline marine waters. The resulting “estuarine circulation” causes mud to be transported landwards on the bottom of the estuary, enhancing the tendency for deposits to become markedly more muddy toward the head of the estuary or bay. The freshwaters also carry suspended sediment loads of silt and clay, especially in river floods, and these tend to undergo flocculation and deposition upon mixing with the saline waters in the upper part of the estuary. The result is a concentration of muddy deposition at the heads of the estuaries.

Supra-tidal (storm surge) mud deposits: Lowland coasts with a broad shallow shelf, especially where they exist in the form of a large funnel, such as the German Friesland bight, the Bohai Bay in China, the Gulf of Thailand, or the Gulf of Bangladesh, may be subject to strong storm surges. Where there is abundant nearshore submarine muddy deposits, wave action during the storm surge event reagitates the silts and clays resulting in waters of high suspended load sediment concentration. These mud-laden waters can be swept onshore, mix with high suspended load river waters, and a mud layer deposited in the supra-tidal zone. Such supra-tidal deposits are found in China, and in Bangladesh as a result of storm surges and tropical cyclones. The material deposited tends to be mainly silt.

Chenier plains: Shelly chenier ridges can form plains as described by Schofield (1960) and Woodroffe et al. (1983) for the Firth of Thames in northern New Zealand. The shells comprise bivalves such as cockles, and during storm events become transported landwards, or otherwise move as wave-induced littoral drift along an active chenier beach face. The chenier ridges typically form shell spits overlying thick deposits of structureless mud, and the seaward-most chenier may comprise the active beach face surmounting a broad intertidal mudflat.

Swamp marsh and wetland deposits: This type of muddy coast deposit occurs in temperate coastal environments beyond the limit of mangroves. Good examples are found along the low terrain coasts which occur on a broadscale around the Dutch and German North Sea coasts, and in the Wash of England. Other well-known examples are found on the landward side of the extensive barrier enclosed lagoon system that is found around the southeast United States and the Gulf coast.

Mangrove forest and swamp deposits: Mangrove forests occur widely in the tropics and subtropics and are typically associated with muddy deposits. Their root mat assists in the trapping of fine suspended sediments. Mangrove stands also occur on the landward shores of coral platform reef and enclosed lagoons, and likewise may induce a veneer of muddy deposits to occur over the coral (Schaeffer-Novelli et al. 2002; Wolanski et al. 2002).

Mud veneer deposits on eroded shore platforms: In this type of muddy coast, an eroded shore platform has subsequently been layered with a veneer of muddy sediment. These types of muddy coast occur on a relatively small-scale and in sheltered harbor environments where there is a high source of muddy material either from adjacent catchments or from sublittoral deposits in the nearshore. Volumes of mud involved are relatively small. Examples occur in the harbors of northern New Zealand.

Ice deposited mud veneer: On Arctic shores, a muddy flat may be formed by ice rafts transporting mud and boulders, grounding and melting out in the thaw so that mud and boulders are jointly deposited on the tidal flats (Dionne 2002).

Sub-littoral mud deposits: Along many coasts in sheltered and semi-enclosed bays where there is an abundant source of fine sediment supply and not particularly strong tidal currents, mud deposits occur in the sublittoral zone. These deposits may be of the fluid mud type, or occur as offshore mud banks, as off the coast of Kerala, India (Mathew and Baba 1995). On occasions of strong wave agitation and shorewards wind-induced surface current, the agitated fine mud sediment particles may be eroded and suspended on the wave turbulence creating an estuarine “turbid fringe” and swept onshore and deposited as a mud layer either in the high tidal zone or the intertidal zone.

Beaches on Muddy Coasts

Although the dominant visual morphological feature of muddy coasts is often the broad intertidal flat, this is really just one component of a beach system. The extensive intertidal flat may also be part of a littoral drift system. In many estuarine beaches the following sequence can be observed: (1) a supra-tidal section with mainly storm surge silt and clay deposits active in storm surge events, and sometimes may

have wind-blown sand cappings; (2) a narrow, steep ($\sim 10^\circ$), active beach face comprised of sand and shelly gravel sediment; (3) a broad intertidal flat containing muddy sediment and often *Zostera* beds; and (4) a subtidal zone containing mainly silts. Of course many variations exist, for example, the intertidal flat may be mainly sand and biogenic gravel with relatively small amounts of mud, and the subtidal sector can contain coarser sediments. But the important point is that the entire system is an active beach and sediment may be transported by wave and current processes over (diabathic transport) and along (parabathic transport) its components. Thus, shells and even terrigenous gravels originating in the subtidal zone may be transported onshore to accrete on the upper beach face. As tidal waters submerge and drain the intertidal flats, agitation of the surface sediments by waves creates a *turbid fringe* of high suspended load sediment – a typical feature of muddy coasts.

Stratigraphy and Sedimentary Structures of Muddy Coastal Deposits

Wang et al. (2002a) identify that tidal flat muddy deposits are often stratified in fine laminae, the delicate bedding being caused by the interlamination of thin layers of very fine sand material in the more argillaceous basic substrate. The lamination of thin bedding is the result of varying hydrodynamic conditions, including the obvious flood and ebb oscillating tidal flows, but also reflecting neap and spring, storm, and seasonal changes. The bedding is seldom strictly parallel, but instead is generally streaky and lenticular. The alternating fine-sandy and clay layers wedge out rapidly in lateral extent and their thickness is not uniform. Such alternating bedding indicates frequent reworking of sediments under the influence of tidal currents of varying strength and direction, and the influence of meteorological factors, especially storm winds and associated waves.

Stratigraphically, modern coastal mud deposits tend to be thin relative to true deep-sea marine muds. This reflects the limited time – only some ~6500 years – since the Holocene sea level reached its approximate present level, as well as the shallow depositional environment, which is typically energetic on a day-to-day Timescale. In some areas the deposits, especially clays, appear to be structureless and massive, but in other deposits there are often rare lenses of sand and shell – the result of episodic storm processes (Park and Choi 2002).

In stratigraphic section, across broad tidal flats such as occur in China, shelly chenier deposits overlie the supratidal clay-rich muds. Typically, these muds laid down by storm surge processes, are clay-rich, massive, lensoid-shaped, and also contain relatively high organic matter, especially plant fragments. The supra-tidal clays overlie the silty muds of

the upper tidal flats, which overlie the alternating fine sand and clay lenses of the mid-tidal area. These in turn overlie the very fine sand and coarse silt of the low tidal zone flat (Wang et al. 2002b).

Sedimentary structures found in muddy coastal deposits include ripple bedforms on the sandier sediments. Within the larger tidal creeks with high current speeds and transporting sandy bedload, bedforms of megaripples, and current-cross-bedding is typical. Upon the mudflats, bioturbation structures produced by organisms inhabiting the tidal flats are also evident, while polygon cracks are to be found on the supra-tidal deposits. These sedimentary structures reflect a range of dynamic processes on the muddy coast tidal flats.

Geographical Spread of Muddy Coast

Muddy coasts are developed in a wide range of coastal climatic and oceanographic environments in the world, from Arctic oceanic coast to the tropics (Wang et al. 2002a). Their detailed geographic location and nature are presented in detail by Flemming (2002). Extensive well-formed, broad expansive intertidal muddy coasts are present around the Yellow River delta in the Bohai Sea of China, as well as the abandoned Yellow River delta in North Jiangsu Province along the Yellow Sea, and the southern embayments of the Changjiang River. Extensive sectors of muddy coast also occur along the east coast of the Americas, with the Amazon River mouth and the northern coasts of Guiana being of immediate note. Extensive muddy coast also occurs along the shores of Canada's Bay of Fundy, around the seasonal ice covered St. Laurence estuary, along the Atlantic east and southeast coasts of the United States, around the Mississippi delta and Vicinal coasts of the Gulf of Mexico, and inside several of the larger bays along the west coast of North America from California to Washington. In Europe, muddy coast deposits occur around the North Sea and Baltic coasts, such as Dollart, Lay and Jade Bays in Germany, the funnel-shaped mouths of large rivers (Elbe, Weser) in the east and north Friesian Bight, and in The Wash and other estuaries of east England. These exhibit many different examples and types of muddy coast. Muddy coast is widely evident in the tropical and monsoonal countries of Asia, including much of the coasts of India, Bangladesh, Thailand, Malaysia, Indonesia, and Vietnam. The northeast coasts of Australia, as well as the islands of the Pacific including New Zealand, New Guinea, and even the coral reef islands, contain areas of muddy coastal deposits. For the African continent, muddy coast occurs mainly within the tropical belt of both the east and west coasts.

It is evident that muddy coasts are characteristic of all continents (perhaps with the exception of Antarctica

which typically exhibits a steep rocky profile for its minor coastal sectors of ice free coastline). They occur in a wide variety of the global coastal environments, but can vary widely in size and morphogenetic environment. However, muddy coasts tend to be particularly conspicuous in tropical zones, especially in the Asia region, and the muddy coastal deposits of largest extent are associated with major continental river discharges.

Factors Leading to Formation of Muddy Coast

A number of factors facilitate and enhance muddy coast formation (Wang et al. 2002a).

Abundant Fine-Grained Sediment Supply

This is the most important factor in influencing the formation of muddy coastal deposits of various types. Fundamentally, if a major source of fine sediments continuously provides muddy sediments at a faster rate than the hydrodynamic conditions can remove them, then muddy coastal deposits will form. Conversely, if muddy source material is not available in sufficient quantities within a coastal environment then muddy deposits do not form. Sources of sediment for concentration to form muddy coast are many and varied, but include:

Adjacent rivers. Adjacent rivers with high concentrations of fine suspended sediment loads. The classic example is the Huangho River of China, which drains a huge catchment including highly erodible loess regolith, which provides the bulk of the fine sediments to the Bohai Bay. Other examples include the Amazon (draining a huge and partly steep catchment of highly weathered tropical soils), the Fly River of New Guinea, the Mekong River of Vietnam/Laos (likewise draining large and partly steep catchments of highly weathered tropical soils), and the Waipaoa River of New Zealand (draining a steepland catchment of highly erodible montmorillonitic clay-rich Tertiary rocks). In the latter cases, the mountainous steepland catchments are also subject to episodic, intense, orographically induced rainfalls, factors which enhance the high volume of sediment load delivered to the littoral zone.

Supply of sediment from offshore. Often sediment that forms or contributes to coastal muddy deposits may be of “secondary origin” in the sense, that it is reworked from offshore. In this sense, the reworked mud deposits are “palimpsest.” Examples include the onshore transfer of fine sediments from the ancient river deposits along the northern Jiangsu coast (Wang et al. 2002b), and the onshore transfer of eroded glacial till deposits from the North and Baltic seas.

Erosion of coastal sedimentary deposits and cliffs. In many enclosed and semienclosed seas, active erosion of

the cliffs provides a limited source of fine material for muddy coastal deposits. In the Kiel Bay of the Western Baltic the surrounding coast is formed of glacial till containing a high proportion of fine sediment (Healy et al. 1987). Rapid erosion of the boulder clay cliffs ($\sim 0.3\text{--}1.0\text{ m/a}$) provides considerable material for local muddy coastal deposits. Where there has been rapid cliff erosion and continuing supply of muddy sediment the shore platform may subsequently become veneered with mud. Likewise in sheltered steepland embayments containing an estuarine lagoon fronted by barrier spits (Healy et al. 1996) erosion of the cliffs produces mud deposits overlying the early Holocene marine sands.

Tidal Range

A macro tidal range ($>4\text{ m}$) is often regarded a necessary factor in explaining extensive broad muddy-coast deposits. However, a general dearth of sediment supply or high wave energy may restrict tidal flat development (Davis 1983). Certainly, the broad expansive intertidal flats and associated muddy deposits of the Jiangsu coast of China are very wide and have associated a macro tidal range. Other extensive muddy coastal deposits associated with large tidal ranges include the Bay of Fundy and The Wash of England (Amos 1995). But in other locations of muddy coasts the tidal ranges may be meso ($2\text{--}4\text{ m}$) or micro ($<2\text{ m}$), and still extensive muddy deposits of various types occur, for example, around the coast of Indonesia, India, the southeast coast of the United States off the mouth of the Hwang-ho (Yellow River) of China, the Fly River of New Guinea, and around northern New Zealand.

A fundamental question is: why should a large tidal range be associated with broad intertidal muddy coastal deposits? Evidently, the answer lies in a different morpho-depositional mechanism from sandy and mixed sand gravel beaches, combined with the thixotropic nature of the muddy deposits. While sandy beaches are characterized by noncohesive grains, and thus undergo regular morphodynamic change in response to varying wave conditions, they typically assume some modal morphodynamic beach state. However, beaches on muddy coasts tend not so obviously to change their morphology, although the beach is active over time. This is because: the individual muddy sediment clasts possess cohesive qualities and thus the surface layer of sediments is not so easily moved around by the waves and currents as are noncohesive sandy sediments. Rather, in energetic wave conditions the surface layers of mud particles are reagitated and taken into suspension within the water column. Moreover, the sediments in purely muddy beaches tend to be deposited in episodic storm surge events with a net onshore movement of suspended sediment providing the material to become deposited on top of the existing muddy flat sediments. Finally, the muddy bulk sediment possesses little strength, and is subject

to fluid and plastic flow, and this results in a low angle of natural repose within the intertidal zone.

Thus, “pure” muddy coast (i.e., without a landward surmounting narrow beach face of sand or shelly gravel or chenier beach) typically exhibits a muddy upper intertidal and supra-tidal zone of subdued relief, which undergoes relatively little morphodynamic change compared to a sandy beach. The essential point seems to be that the mode of formation of the intertidal muddy deposits is by net onshore sediment movement, and that the morphodynamics of deposition are such that an essentially flat topographic surface results. Providing there is abundant and continuing fine sediment supply then the extent or width of a muddy intertidal flat becomes a function of the tidal range (Wang et al. 2002b).

Conclusion

Muddy coast geomorphology and sedimentary depositional systems occur widely around the earths’ coastlines, but compared to sandy coasts have been relatively little studied. Texturally, “mud” is defined based upon particle size, and comprises mineral material of silt and clay sizes. Muddy coasts are defined as possessing distinctive muddy deposits, typically but not exclusively manifest as broad intertidal muddy flats. Additional geomorphic features found on muddy coasts include chenier ridges and beaches, veneer muddy deposits, and shallow offshore muddy deposits, while associated biologic features may occur such as mangrove stands and salt marshes. Muddy coasts occur widely in geographic distribution from the tropics to the Arctic. Their fundamental *raison d’existence* is abundant mud-sized sediment particles in the coastal environment, but the existence of broad intertidal mudflats is often associated with a large tidal range.

Cross-References

- Beach Sediment Characteristics
- Cheniers
- Coastal Sedimentary Facies
- Mangroves, Geomorphology
- Open-Coast Tidal Flat Deposits
- Salt Marsh
- Storm Surge
- Tidal Flats

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Natural Hazards

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Definition

Natural hazards are physical phenomena that expose the coastal zone to risk of property damage, loss of life, or environmental degradation. *Rapid-onset* hazards last over periods of minutes to several days. Examples include major cyclones, accompanied by high winds, waves, surges, and tsunamis – giant sea waves set off by earthquakes, volcanic eruptions, or submarine landslides. *Slow-onset* hazards develop incrementally over longer time periods. Examples include coastal erosion and sea-level rise.

The vulnerability of the world's coastlines to natural hazards varies considerably, because climate, tectonism, and other physical variables such as bathymetry, shelf width, landform, lithology, and coastal configuration change from place to place. For example, although tropical cyclones may form wherever sea surface temperatures exceed 27 °C, landfall occurs most commonly in particular regions such as southeast Asia, the Caribbean Sea, or southeastern United States (Fig. 1). Tsunamis are most prevalent around the Pacific Ocean, due to the seismicity and volcanicity associated with convergent-plate boundaries along the circum-Pacific "Ring of Fire."

Rapid coastal development and urbanization expose increasing numbers of people to natural hazards in a dynamic and unstable environment. Ten percent of the world's population and 13% of the urban population live on the 2% of the Earth's land area at or below 10 m (McGranahan 2007). In the United States, in 2010, 123.3 million people, or 39% of the

nation's population, lived in Coastal Shoreline Counties, defined as "counties directly adjacent to the open ocean, major estuaries, and the Great Lakes" (NOAA 2013a). Anticipated climate changes will greatly amplify risks to coastal populations. By the end of this century, rise in sea level by 2–5 times present rates could lead to more frequent flooding and inundation of low-lying regions, worsening beach erosion, as well as loss of ecologically productive coastal wetlands, and saltwater intrusion into coastal aquifers and estuaries (IPCC 2013, 2014).

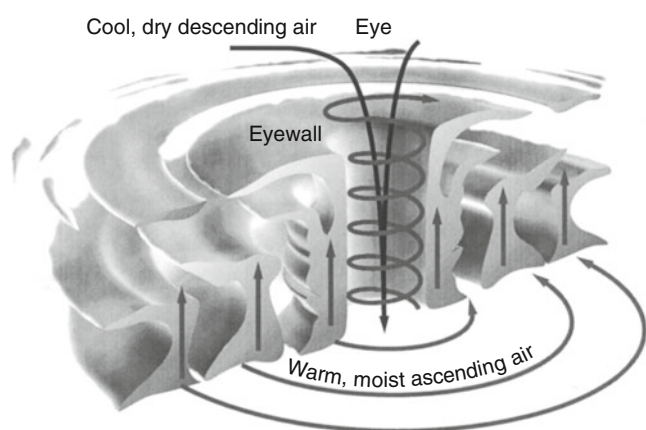
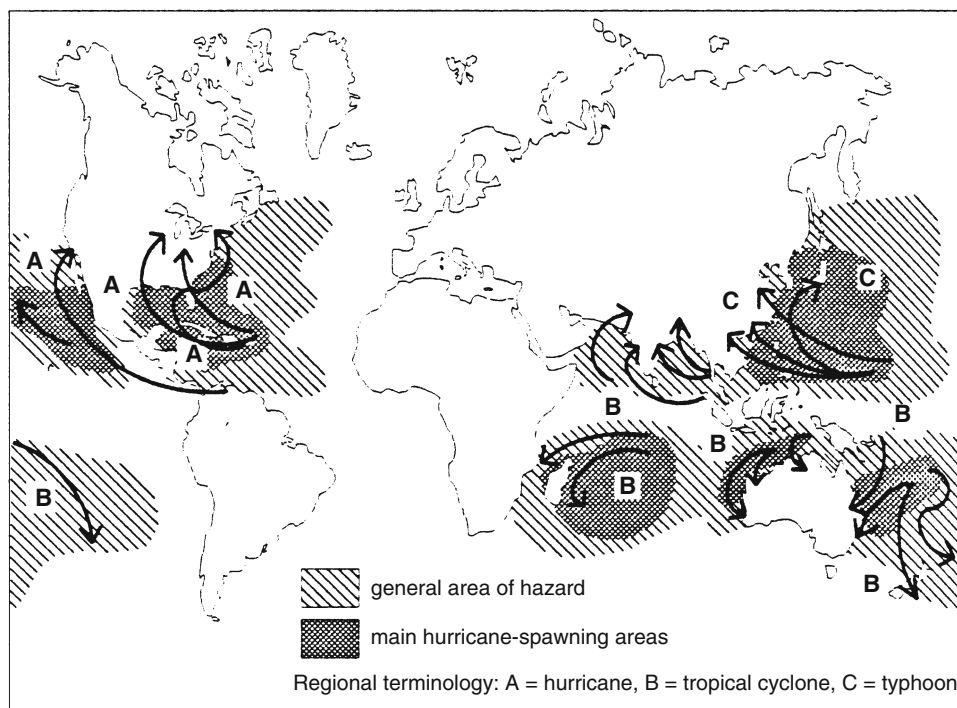
Rapid-Onset Hazards

Tropical Cyclones

Tropical cyclones (also called *hurricanes* in the Atlantic basin, *cyclones* in the Indian Ocean, and *typhoons* in the western Pacific) are major low-pressure systems that develop and intensify over the open ocean between $\pm 5^\circ$ and 25° latitude (Fig. 1). Hurricanes form where sea surface temperatures rise above 27 °C, under conditions of high relative humidity, strong evaporation rates, weak vertical wind shear, and atmospheric instability (Zehnder 2015). In the Atlantic basin, the hurricane season lasts from June to November, peaking between August and early October.

A tropical cyclone consists of a rotating atmospheric low-pressure system, ranging from less than one hundred to several hundred kilometers in diameter. At the center is the "eye" – a zone of calm atmosphere and nearly clear sky. It is surrounded by the eyewall, where the strongest winds of the hurricane are concentrated (Fig. 2). Spiraling around the eyewall are rainbands of high winds and torrential rains. In the Northern Hemisphere, the cyclonic system rotates counterclockwise around the eye, while simultaneously moving forward along a northwesterly track. Thus, winds are stronger on the cyclone's right side, due to the additive effect of the mean forward velocity of the storm and the rotational speed around the eye (Coch 2015).

Natural Hazards, Fig. 1 Location of tropical cyclone breeding areas (From Alexander 1993)



Natural Hazards, Fig. 2 Structure of a hurricane

A number of factors contribute to hurricane damage. One of these is the damage produced by very high wind speeds (minimum 119 km/year). In addition, hurricanes cause severe flooding due to the combination of heavy rainfall, waves, and storm surge. In coastal areas, a major hazard is flooding by the storm surge, which is a dome of water produced by the low barometric pressure and strong wind shear (see ► [Storm Surge](#)). The area at greatest risk lies to the right of the eyewall at landfall; the risk falls off with increasing distance. The hazard also depends on the mean forward velocity of the storm: a slowly moving storm can cause flooding over several tidal cycles. The angle at which the hurricane track crosses the coast also determines the hazard. A hurricane track which parallels the eastern coastline of a continent keeps the stronger

right side of the cyclone seaward, thereby reducing the damage potential to land. On the other hand, a coast-normal track crossing land may cause serious damage on the right side of the storm (Coch 2015).

The intensity of hurricanes is classified from 1 to 5 on the Saffir-Simpson scale, with 5 representing the most severe category (Table 1). This scale rates hurricanes on the basis of central air pressure (the lower the pressure, the more intense the storm), sustained wind velocity, storm surge potential, and potential for property damage. The three most powerful US hurricanes between 1900 and 2010 includes a category 5 storm that struck the Florida Keys in 1935, Camille in 1969 (category 5; Louisiana-Mississippi), and Katrina in 2005 (category 3; southeast Louisiana-Mississippi) (Blake et al. 2011).

The deadliest US hurricane (category 4) struck Galveston, Texas, in September, 1900, killing over 8000 people. The next most lethal hurricanes include one that swept over southeast Florida and Lake Okeechobee in 1928 (category 4), Katrina, 2005 (category 3), Audrey in Louisiana/Texas, 1957 (category 4), and an unnamed storm that hit the Florida Keys in 1935 (category 5) (Blake et al. 2011). In the United States, deaths caused by hurricanes have decreased substantially in recent years, because of the advance warning provided by weather satellite tracking. In other parts of the world, however, tropical cyclones continue to exert a high toll (see below).

Tropical cyclone frequency has not changed detectably since the 1970s, although the intensity of the strongest North Atlantic cyclones has increased (Elsner et al. 2008)

Natural Hazards, Table 1 The Saffir-Simpson scale of hurricane intensity (After NOAA 2013b)

Category	Central pressure (millibars)	Winds (km/h)	Surge (m) ^a	Damage
<i>The Saffir-Simpson scale</i>				
1	≥980	119–153	1.0–1.7	Dangerous
2	965–979	154–177	1.8–2.6	Extremely dangerous
3	945–964	178–208	2.7–3.8	Devastating
4	920–944	209–251	3.9–5.6	Catastrophic
5	<920	≥252	>5.7+	Disaster

^aEstimated ranges only. The actual surge depends on a number of factors, including sustained wind speed and direction, storm track, bathymetry, and near-shore topography

Effects

1. Dangerous: Some roof damage, broken branches, some damage to power lines and local power outages. Minor coastal flooding
2. Extremely dangerous: Damage to frame homes, some uprooted trees, serious power outages, coastal flooding
3. Devastating damage: Major damage to homes, many trees uprooted or broken, widespread power outages. Floodwaters may cover terrain below 1.5 m
4. Catastrophic damage: Severe damage to frame homes, loss of roof; most trees uprooted or broken, extensive power outages lasting weeks or longer; large areas uninhabitable for weeks to months. Potential flooding of terrain below 3 m, requires mass evacuation
5. Extremely catastrophic damage: Large number of destroyed framed homes in area; roof failure; fallen trees and broken power lines will isolate the area, rendering it uninhabitable for weeks to months. Mass evacuation may be necessary

and will likely continue to strengthen in the future (Knutson et al. 2010; Peduzzi et al. 2012). The most intense tropical cyclones have also shifted poleward within the past three decades (Kossin et al. 2014). The costliest US hurricanes have occurred in recent years, i.e., Katrina in 2005 (category 3, southeast Florida, Louisiana, Mississippi, \$153.8–\$108 billion), Sandy in 2012 (coastal North Carolina to Massachusetts, \$68.3–\$65 billion), Ike in 2008 (category 2, Texas, Louisiana, \$29.5–\$33.6 billion), and Wilma in 2005 (category 3, south Florida, \$21–23.4 billion) (Blake et al. 2011; NOAA 2017). Extensive coastal development over the last few decades has increased the amount of property at risk to devastating storms (Pielke et al. 2008). Increasing coastal populations, economic development, and greater cyclone intensity will magnify future flooding risks and damages (Peduzzi et al. 2012; Mendelsohn et al. 2012; Hallegatte et al. 2013; Woodruff et al. 2013; Estrada et al. 2015).

Extratropical Cyclones

Extratropical cyclones are the dominant type of storm responsible for major coastal flooding and beach erosion at mid-latitudes. These storms occur most frequently between October and April in the Northern Hemisphere, peaking in winter. Although wind speeds and surge heights are much lower than in hurricanes, extratropical cyclones inflict considerable

Natural Hazards, Table 2 Dolan-Davis scale for the classification of Atlantic Coast nor'easters (After Dolan and Davis 1994)

Storm class	Frequency (percent) (N = 1564)	Significant wave height (m)	Duration (h)	Power (m ² h)
<i>The Dolan-Davis scale</i>				
1 Weak	50.3	2.0	8	32
2 Moderate	25.1	2.5	19	107
3 Moderate	21.6	3.2	35	384
4 Severe	2.5	5.0	62	1420
5 Extreme	0.5	6.8	97	4332

Effects

1. Moderate beach erosion; minor dune erosion. No property damage
2. Minor beach erosion. No property damage
3. Significant beach and dune erosion. Moderate property damage
4. Severe beach and dune erosion. Overwash damage on low-profile beaches. Community-wide loss of structures
5. Extreme beach erosion, dune destruction. Massive overwash in sheets and channels. Extensive regional-scale property losses in millions of dollars

damage because of their greater spatial extent (typically over 1000 km versus 100–150 km for hurricanes), high waves, and duration spanning several tidal cycles at a given location (Colle et al. 2015; Dolan and Davis 1992). Areas at risk to flooding from extratropical cyclones include the Atlantic Coast of the United States north of 36°N, northwestern Europe, and northeastern Asia.

US Atlantic coast winter storms (*northeasters* or *nor'easters*) have been classified into five categories (5 is the most severe) based on significant wave height, $H_{1/3}$ (average of highest one-third of waves), duration,

D , and power, P , where $P = H_{1/3}^2 \times D$ (Dolan and Davis 1992, 1994; Table 2). (The energy of a wave, hence its destructive capability, is proportional to $H_{1/3}^2$). A qualitative assessment of the relation of storm class to coastal damage is also indicated in Table 2.

Nor'easters affecting the eastern United States have varied in number and severity over time, but do not show any long-term trends (Zhang et al. 2000; Colle et al. 2015). Significant nor'easters within the last 40 years include the “Ash Wednesday” storm (March 6–7, 1962), the Halloween storm (October 31, 1991), and several other powerful storms on December 11–12, 1992, March 13–14, 1993, and March 13–14, 2010. But the most devastating regional event was Superstorm Sandy (a hybrid extra-tropical/tropical storm) in late October, 2012. It struck New York City on October 29 and produced the highest recorded water level (3.38 m) at the Battery (southernmost Manhattan) in nearly 300 years (Talke et al. 2014; Orton et al. 2016), due to a combination of strong easterly winds, surge amplification, historic sea-level rise (0.44 m since 1856), and maximum storm surge coinciding with high tide and full moon. Storm damages totaled an estimated \$19 billion (SIRR 2013).

Northwestern Europe has also experienced severe winter storms. On February 1, 1953, a North Sea storm swept over the Netherlands, breaching dikes, inundating 150,000 ha, and killing over 1800 people. That storm also caused major damage in eastern Great Britain (Alexander 1993, p. 138; McGill University 2008). Storms during the winter of 2013/2014 generated the strongest wave activity along Europe's Atlantic coastline since 1948 (Masselink et al. 2016). High cyclone frequency and above-average cyclone intensity, due to an intense polar vortex and unusually strong North Atlantic jet stream, resulted in a season of exceptional storminess. Beach erosion along both sandy and rocky beaches was severe. Another fierce winter storm lashed west Europe on January 12, 2017, leaving over 300,000 homes in northwest France without power. The storm, named "Egon," whipped the northern port city of Dieppe with hurricane-force winds (France24 2017). The storm caused flooding in southeast England, heavy snow, rain, and cold throughout much of west Europe, and also resulted in extensive transportation disruptions.

Storm Surges

A storm surge is the elevation of ocean water level above that expected from astronomical tides, caused by a passing storm. The surge results from the reduced atmospheric pressure, which causes sea water to rise by around 1 cm for each drop in pressure of 1 mb, and from the wind stress, which literally pushes water toward the coast. Factors contributing to the surge height include the storm intensity, width and slope of coastal shelf, and coastal configuration. The total flood height also depends on the tidal stage (see ► [Storm Surge](#)).

The flood surge is a major cause of coastal erosion (see below). On exposed barrier islands, sand washes from the beach and dunes, and redeposits bayward in overwash fans. The surge also cuts channels or tidal inlets across the barrier. As the surge ebbs, seaward-flowing water creates further erosion. The flood surge damages buildings and other structures by direct wave impact, hydraulic lift of waves and water beneath raised buildings, wave energy reflected from protruding structures such as groins and jetties, as well as battering from floating debris (Coch 2015).

Regions of the world most vulnerable to surges from tropical cyclones include low-lying, deltaic coasts of south-east Asia, Bangladesh, China, the Philippines, the southeastern United States, the Caribbean, and northern Australia (Figs. 1 and 3). Surges from extratropical cyclones have also produced extensive flooding in the Netherlands, Germany, Italy, as well as the eastern United States (see ► [Storm Surge](#)).

The geographical location of Bangladesh at the apex of the Bay of Bengal, low-lying topography, and high population density makes it especially vulnerable to storm surges from tropical cyclones (Murty and Flather 1994). In 1970, a powerful storm killed half a million people; another intense tropical cyclone in 1991 claimed close to 140,000 lives. Since then, Bangladesh has instituted early warning systems, developed storm shelters and evacuation plans, constructed coastal embankments, restored mangrove forest cover, and raised public awareness. Thus, in 2007, a cyclone of comparable intensity to the two earlier ones took only around 4000 lives (Haque et al. 2012). In stark contrast, in neighboring Myanmar, where advance evacuation warnings and protective



Natural Hazards, Fig. 3 Major low-lying deltaic areas of the world. Tracks of tropical cyclones are also shown

measures were lacking, Cyclone Nargis killed over 138,000 people on May 2, 2008 (Fritz et al. 2009) (see also ► [Storm Surge](#)).

In Europe, the February 1, 1953, winter storm overwhelmed the Dutch dike defenses, causing over 1800 deaths in Holland, with an additional 300 fatalities in the United Kingdom (McGill University 2008). The disaster flooded 9% of Dutch farmland and led to the construction of the Delta Works – a massive engineering project designed to protect much of the coast from future coastal storm surges (Delta Committee 2008; Aerts et al. 2009). In Venice, a storm flood of 1.94 m above the local datum inundated most of the historic city on November 3–4, 1966. The frequency of a “very intense” 110-centimeter storm tide above the 1897 reference datum (which triggers a warning alarm system) has increased from once every 2 years to approximately 4 times per year during the twentieth century (Carbognin et al. 2009). This flood level could occur around 21 times a year for a 20-centimeter sea-level rise, and as much as 250 times a year for a 50-centimeter rise! To protect against more frequent flooding, Venice has constructed the MOSE system of mobile tidal gates that will be raised whenever tides over 1.1 m are predicted.

Tsunamis

A tsunami is a series of waves caused by the vertical displacement of ocean water, triggered by an earthquake, volcanic eruption, landslide, or more rarely, the impact of a bolide. Most tsunamis occur on islands and shores around the Pacific Ocean. In contrast to ordinary water waves, tsunamis have long wavelengths (from 150 km to 700 km), long periods (10–60 min), and low amplitudes (≤ 1 m). They propagate with enormous speeds (600–800 km/hr) over open ocean. The velocity of propagation is given by $v = \sqrt{g \cdot h}$, where v is velocity, g is the acceleration due to gravity, and h is water depth. The wavelength is $L = v \cdot T$, where L is wavelength and T is the period (Alexander 1993; Bell 1999).

The danger of tsunamis increases dramatically as the waves approach shallow water, where they slow down but increase tremendously in height. Water levels rise rapidly up to 20 m above normal sea level. Occasionally, a “bore” or wall of turbulent water moves up narrow bays or estuaries. The highest wave may occur hours after several lower ones have passed. The height of the wave as it arrives at the shore is related to the original amount of water displaced, distance from source, nearshore bathymetry, and coastal configuration (Bell 1999). The destructiveness is linked to the run-up, or maximum vertical height. A new tsunami intensity scale (Papadopoulos and Imamura 2001) qualitatively describes the destructive impacts of a tsunami on a scale of 1 to 12, similar to the Mercalli scale of earthquakes (Table 3). After the devastating Tohoku (Sendai) earthquake in Japan on March 11, 2011, a revised new scale was proposed, which

Natural Hazards, Table 3 A new tsunami intensity scale (After Papadopoulos and Imamura 2001)

Intensity		Wave height (m)
I	Not felt	0
II	Scarcely felt	0.0
III	Weak	<1
IV	Largely observed	<1
V	Strong	1
VI	Slightly damaging	2
VII	Damaging	4
VIII	Heavily damaging	4
IX	Destructive	8
X	Very destructive	8
XI	Devastating	16
XII	Completely devastating	32

Effects

- I. Not felt. Not felt, even under the most favorable circumstances. No effect; no damage
- II. Scarcely felt. Felt by few people on small boats. Not observed on the coast. No effect; no damage
- III. Weak. Felt by most people on small boats. Observed by few people on the coast. No effect; no damage
- IV. Largely observed. Felt by all on small vessels and a few on large vessels. Observed by most people on the coast. Some small boats move slightly ashore. No damage
- V. Strong. Felt by all on large vessels and observed by all on the coast. Some people are frightened and run to higher ground. Many small vessels move strongly ashore, a few crash into each other or overturn. Traces of sand layer are left behind under favorable conditions. Limited flooding of cultivated land, gardens, outdoor facilities of near-shore structures
- VI. Slightly damaging. Many people are frightened and run to higher ground. Most small boats shoved violently ashore, or crash into each other, or overturn. Damage and flooding of a few wooden structures. Most masonry buildings hold
- VII. Damaging. Most people are frightened and try to reach higher ground. Many small boats are damaged; a few larger ones oscillate violently. Objects of varying sizes and stability overturn and drift. Sand layer and accumulations of pebbles are left behind. Many wooden structures damaged, a few demolished or washed away. Slight damage and flooding to a few masonry buildings
- VIII. Heavily damaging. All people escape to higher ground; a few are washed away. Most small boats are damaged; many washed away. A few large vessels are moved ashore or crash into each other. Big objects drift away. Beach erosion and littering. Extensive flooding. Most wooden structures are washed away or demolished. Moderate damage to masonry buildings
- IX. Destructive. Many people are washed away. Most small vessels are destroyed or washed away. Many large vessels are moved violently ashore, a few destroyed. Extensive beach erosion and littering. Local ground subsidence. Heavy damage to masonry buildings
- X. Very destructive. General panic. Most people are washed away. Most large vessels are moved violently ashore, many are destroyed or collide with buildings. Small boulders from the seafloor are moved inland. Cars are overturned and drift away. Oil spills, fires start. Extensive ground subsidence. Destruction of many masonry buildings. Artificial embankments collapse; port water breaks damaged
- XI. Devastating. Lifelines interrupted. Extensive fires. Water backwash drifts cars and other objects to sea. Big seafloor boulders moved inland. Destruction of many masonry buildings
- XII. Completely devastating. Practically all masonry buildings demolished

adds more specific criteria to the earlier scale (Lekkas et al. 2013).

Tsunamis are usually caused by displacement of water due to vertical movement of the seafloor along faults, during major earthquakes associated with plate subduction or other tectonic events.

Waves are formed as the displaced water mass attempts to return to an equilibrium state. However, major volcanic eruptions can also set off tsunamis. For example, the catastrophic tsunami produced by the Krakatau eruption in 1883 generated waves up to 40 m high in places, and killed over 36,000 people (Winchester 2003). Less commonly, submarine landslides can also cause tsunamis, as occurred in Papua New Guinea in 1998, following a magnitude 7.1 earthquake (Lockridge et al. 2002).

Around 90% of destructive tsunamis occur around the Pacific Basin at an average rate of 2 per year (Alexander 1993). The tsunami caused by the 1960 Valdivia, Chile, earthquake (magnitude 9.5, the largest recorded earthquake) produced waves up to 10.7 m at Hilo, killing 61 people. Crescent City, California, was struck by waves up to 6.3 m high from the 1964 Alaska earthquake (magnitude 9.2), which caused 11 fatalities and US\$7.5 million in property damage. On December 26, 2004, another devastating tsunami, set off by a powerful magnitude 9.2 earthquake offshore of Aceh, northern Sumatra, swept across the Indian Ocean. The disaster affected northwestern Sumatra, Thailand, Sri Lanka, eastern India, and other countries bordering the Indian Ocean, killing over 230,000 people. A few years later, the Tohoku (Sendai) tsunami, triggered by a magnitude 9.1 undersea earthquake, created a deadly wall of water up to 40 m high in places that overwhelmed the northeast coast of Honshu island, Japan on March 11, 2011. Rushing inland, it wiped out nearly everything in its path, and killed an estimated 18,500–20,000 people. The tsunami also destroyed three nuclear reactors at the Fukushima power plant, necessitating a mass evacuation of a 20 km area surrounding the plant, because of possible radiation contamination (Pletcher and Rafferty 2011/2016). The Atlantic Coast of North America is also potentially at risk to destructive tsunamis, because of slope instabilities on the continental shelf. In November, 1929, a magnitude 7.1 earthquake triggered a turbidity current (suspended sediment in water) off the coast of Newfoundland that snapped a dozen transatlantic telegraph cables. The resulting tsunami caused 29 deaths (Lockridge et al. 2002). Discharge of pressurized natural gas and gas hydrates along fractures along the outer continental shelf edge off southern Virginia-North Carolina could eventually destabilize the shelf edge, causing massive submarine landslides and tsunamis (Driscoll et al. 2000). Other possible triggers include rapid ocean warming or reduced pressure during passage of a hurricane (Lockridge et al. 2002). Underconsolidated sediments on the New Jersey continental slope are also potentially

susceptible to slumping (Dugan and Flemings 2000). Massive collapse of the flanks of Canary Island volcanoes looms as another potential East Coast tsunami threat (Lockridge et al. 2002). Mega-landslides may occur if the volcano's slopes become too steep to remain stable, whereupon giant blocks slide into the sea, as has happened in the Hawaiian Islands, and possibly the Canaries as well. Giant limestone boulders perched 20 m high and 500 m inland on Eleuthera, in the Bahamas, may be the result of a Canary Island landslide and associated tsunami.

Slow-Onset Hazards

Coastal Erosion

Around 70% of the world's sandy beaches are retreating at present (Bird 2008). The prevalence of beach and cliff erosion from many parts of the world, even where human impacts are minimal, implies an underlying global cause, such as the recent sea-level rise (IPCC 2013, 2014; see below), which may have exacerbated other more localized processes. Among these are differences in rock resistance, coastal subsidence, diminution in sediment yield of rivers (because of reduced precipitation or upstream entrapment in artificial reservoirs), increase in wave attack due to greater number or severity of coastal storms, longshore drift, beach mining, or cliff quarrying, and presence of engineering structures that intercept sediment flow or enhance scour (Bird 2008). In the United States, the average erosion rate along the Gulf Coast is close to 3 m/year, with 61% of the measured shoreline retreating (Morton et al. 2004). Mean erosion rates in Louisiana are 8.2 m/year – a consequence of the high relative sea-level rise due to land subsidence (Morton et al. 2004). Erosion rates along the Atlantic Coast from Virginia to New England average 1.4 m/year; 23% of the measured coastline is eroding over 1 m/year (Hapke et al. 2011). The Atlantic side of the southern Delmarva Peninsula is an erosional “hotspot,” possibly related to anomalously high subsidence rates near the mouth of Chesapeake Bay and jetties at Ocean City, Maryland, which have curbed southward littoral drift to Assateague Island (Hapke et al. 2011; Schupp 2013). Widespread coastal erosion has also occurred as a result of major nor'easters (see above).

Coastal erosion is less severe on the Pacific Coast. California's most extensive coastal hazard comes from eroding cliffs and bluffs. Cliff failure arises from wave attack, subaerial erosion, mass movements, and earthquakes. Beaches and dunes are also prone to erosion, particularly following higher sea level, waves, and storm tides during El Niño events, such as in 2015–2016 (Barnard et al. 2017; see also ► [El Niño–Southern Oscillation \(ENSO\)](#)). Damming of many rivers has reduced the sediment supply to the sea, limiting beach recovery following episodic erosion events.

Beach nourishment and dune restoration are now a major means of shore protection against storm damage and long-term erosion (NRC 1995). Beach nourishment or restoration consists of placing sand that has usually been dredged from offshore or other locations onto the upper part of the beach. The process must be repeated at intervals which depend on local factors such as beach profile, average sand grain size, and anticipated losses due to storms and other processes. An estimated US\$2.4 billion (adjusted to \$1996) has been spent renourishing Atlantic, Gulf Coasts, and Great Lakes beaches since the 1920s (Program for the Study of Developed Shorelines 1999). Over US\$480 million have been spent replenishing East Coast beaches between 1990 and 1996 alone (Valverde et al. 1999).

Sea-Level Rise

Sea-level rise represents a slow-onset hazard whose main impacts will become apparent only decades from now. Mean global sea level has been increasing by around 1–2.0 mm/year over the last 100–150 years. This is the most rapid rate within the last few thousand years and is probably linked to the twentieth-century warming trend of $\sim 0.6^\circ\text{C}$ (IPCC 2013; Gehrels and Woodworth 2013; Cazenave and Cozannet 2013). The recent sea-level rise comes mainly from melting of mountain glaciers and thermal expansion of the upper ocean layers, with lesser contributions from polar ice sheets. Anticipated global warming could increase rates of sea-level rise by factors of 2–5 by the end of this century, greatly amplifying risks to coastal populations (IPCC 2013, 2014). The major consequences of accelerated sea-level rise are permanent inundation of the shoreline, more frequent coastal flooding, increased erosion, and saltwater intrusion (Cazenave and Cozannet 2013; IPCC 2014).

The impacts of sea-level rise will not be globally uniform, because of local variations in vertical crustal movements, topography, lithology, wave climatology, longshore currents, and storm frequencies. Low gradient coastal landforms most susceptible to inundation include deltas, estuaries, beaches and barrier islands, coral reefs, and atolls. Regions most at risk include the Low Countries of Europe, eastern England, the Nile Delta, Egypt, the Ganges-Brahmaputra, Irrawaddy, and Chao Phraya deltas of southeastern Asia, eastern Sumatra, and Borneo (Fig. 3). In the United States, the mid-Atlantic coastal plain, the Florida Everglades, and the Mississippi Delta will be especially vulnerable.

Coastal wetlands will be among the most severely affected ecosystems since they lie largely within the intertidal zone. The ability of a salt marsh (or mangrove) to keep pace with rising sea level depends on relative rates of submergence versus vertical accretion. In the United States, present rates of marsh accretion are generally keeping up with relative sea-level rise, except for parts of Louisiana and Chesapeake Bay (NRC 1987). Louisiana, with a relative sea-level rise of

~ 10 mm/year, is losing around $42.9\text{ km}^2/\text{year}$ between 1985–2010 (Couvillion et al. 2011). The recent increase in coastal land loss can be attributed to several severe hurricanes which have struck the state.

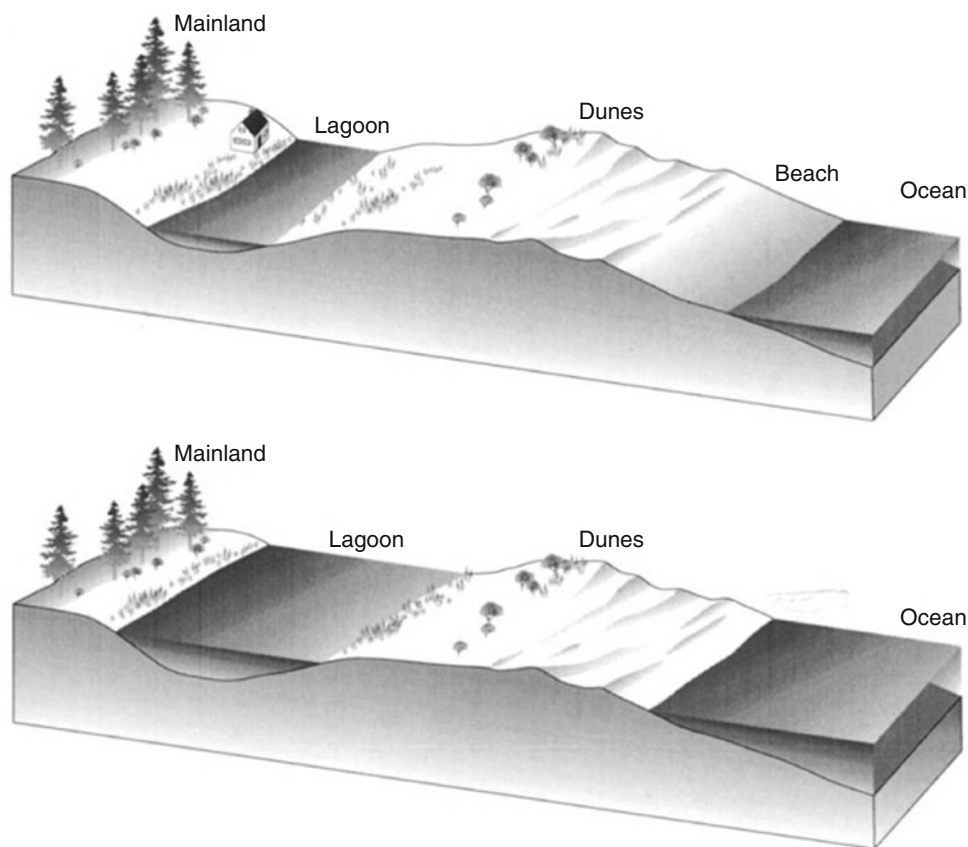
Coral islands, with average elevations of only 1.5–2 m above present sea level are also at high risk. If projected rates of sea-level rise approach 10–12 mm/year, near the upper limit of coral growth, some slow-growing species may drown, but a slight rise in sea surface temperatures may prove even more detrimental to coral reef survival (IPCC 2014). The additional stress of increasing sea-surface temperatures may inhibit the ability of corals to keep pace with sea-level rise. Increasingly frequent coral “bleaching” occurrences in recent years have been linked to warmer than normal sea surface temperatures, in part associated with El Niño events. Tropical ocean islands and reefs at risk include the Maldives, many Pacific and Caribbean islands, and Australia’s Great Barrier Reef.

Coastal flooding due to storm surges will increase in frequency with sea-level rise, even if the number and strength of storms do not change (Tebaldi et al. 2012). The flood return period is very sensitive to small changes in sea level. For example, by the 2050s, the return period of the 100-year flood in New York City would be reduced to around 28 years, for a projected sea-level rise of 30 inches (0.76 m) at the 90th percentile (Horton et al. 2015; see ► [Storm Surge](#)).

Coastal erosion is likely to increase as sea-level rises. Waves will be able to cut into cliffs at a higher level, triggering mass movements and slumping in poorly consolidated sediments. Barrier islands respond to rising ocean levels by eroding on their ocean sides and depositing sand by overwash on the bay side, a process called “barrier rollover” (Fig. 4).

The high-tide shoreline’s response to sea-level rise is often estimated using the Bruun Rule, which states that a typical concave-upward beach profile erodes sand from the beach-front and deposits it offshore, so as to maintain constant water depth (see ► [Beach Erosion](#) ► [Coastline Changes](#)). High-tide shoreline retreat depends on the average slope of the shore profile. A 1 m sea-level rise could cause the beach to retreat by as much as 100–200 m, depending on the slope. The Bruun Rule has been generalized to account for landward migration and upward growth of the barrier, but although still widely used, it has also been criticized for overlooking gains or losses due to longshore currents, or washover by storms. It also fails to account for other processes that affect shoreline position, such as sand supply, wave energy, or storm frequency. Therefore, some suggest it should be abandoned altogether (Cooper and Pilkey 2004). As sea level rises, saltwater will generally penetrate further up estuaries and rivers, as well as infiltrate into coastal aquifers, which could contaminate urban water supplies. Excessive groundwater pumping has already induced an upward migration of the saltwater-freshwater interface in many coastal localities. Upstream migration of

Natural Hazards, Fig. 4 Effects of sea-level rise on barrier islands. *Top*, present sea level; *bottom*, future sea level



the salinity front due to sea-level rise is analogous to that occurring under present drought conditions.

- [Meteorological Effects on Coasts](#)
- [Sea-Level and Climate Change](#)
- [Storm Surge](#)
- [Tsunami Deposits](#)

Conclusions

The shorelines of the world are exposed to a number of natural hazards, among which are those that provoke the greatest devastation: the high surge levels and intense winds of tropical cyclones and also tsunamis. Other coastal hazards include surges and high waves from nontropical storms, as well as slower-onset processes such as beach and cliff erosion and sea-level rise. Low-lying coastal plains, wetlands, major river deltas, and coral atolls are landforms most vulnerable to sea-level rise.

Cross-References

- [Barrier Island Landforms](#)
- [Beach Erosion](#)
- [Changing Sea Levels](#)
- [Coastal Changes, Gradual](#)
- [Coastal Changes, Rapid](#)
- [Coastline Changes](#)
- [El Niño–Southern Oscillation \(ENSO\)](#)

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Navigation Structures

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Nations with access to oceans or to large seas, rivers, and lakes utilize the resource of water-borne transport for defense, commerce, and recreation. Structures that protect and promote boat and ship traffic are among the oldest and largest engineered works along the coast. Masters (1996) discusses coastal harbors constructed in the Mediterranean from about 4,000 years ago, including a Phoenician harbor with almost continuous use for the past 3,000 years. Franco (1996) describes the engineering and functioning of pre-Roman harbors starting from the first harbor of Alexandria, Egypt, built by the Minoans some 3,800 years ago, and he gives examples of Roman and Italian design of harbor structures, including contributions by Leonardo da Vinci.

Safe passage through harbors, inlets, and river mouths requires (1) sheltering of vessels from breaking waves that

form the surf zone, (2) a relatively straight and permanent channel, and (3) a channel that does not unexpectedly fill with sediment (called sediment shoaling or simply “shoaling”). Engineering works that promote reliable navigation are called coastal navigation structures. The two major coastal navigation structures are jetties and breakwaters. This entry discusses the functioning of jetties and breakwaters, and their interaction with the coast.

Jetties

Jetties are typically placed at inlets and river mouths to maintain and promote navigability. A jetty is an engineered structure extending from the shore into a body of water and is usually constructed to serve three purposes. First, jetties partially or fully block waves, sheltering vessels as they pass through shallower water and the breaking waves of the surf zone to reach deepwater. In this way, jetties act as a breakwater, described in the next section. Second, jetties stabilize the location of the inlet, river, or harbor entrance, and in doing so they direct and confine the tidal and river current. By stabilizing the entrance location and directing the current, the location of the navigation channel tends to remain fixed in location and at least partially scoured by the current. Third, jetties block the movement of sediment transported along-shore (the littoral or shore drift), keeping the channel clear. Channel location and removal of shoals that impede navigation is accomplished by dredging.

If a single jetty stabilizes the navigation channel, it is built updrift of the entrance to block the predominant waves and littoral drift. A single jetty stabilizes the location of an inlet or river mouth by preventing its migration through spit extension from the updrift side. Typically, however, jetties are built in pairs that flank the harbor, inlet, or river mouth. As much as possible, jetties are aligned such that vessels exit directly into the incident waves. On a long and open coast, jetties are aligned perpendicular to the shore, because wave refraction causes wave crests to be parallel to shore. If the jetties protect a harbor or entrance near a headland or other area where the crests of the incident waves are not parallel to the coast, then one or more jetties may be constructed at an angle to the shore or have a seaward segment aligned at an angle to better block the waves.

Jetty length is usually such as to extend across the surf zone that can be present under navigable wave conditions. As a rule of thumb, the jetties are extended to a depth approximately that of the designed navigable depth. The elevation of the landward ends of jetties is typically above sea level to prevent sand moving along the coast from passing over the structure. The seaward end may gradually taper below the water and become submerged to allow currents and waves to pass around the structure without direct impact. However, the

seaward ends of most jetties are above water and armored with larger stone to withstand direct impacts by waves. In the United States, jetties are typically constructed of stone that is cut according to certain specifications and fit in place. These structures are referred to as rubble mound jetties.

Because jetties intercept sand moving along the coast, provision is usually made to bypass the beach-quality sand or sediment to the down-drift side (Seabergh and Kraus 2003). Sediment that falls into the channel is dredged and can be pumped to the downdrift nearshore or beach to prevent or reduce erosion. At longer jetties or where the wave climate is severe, beach-quality sediment may be dumped in the nearshore if it is infeasible or extremely expensive to place the material on the beach.

Another means of bypassing sand is by construction of a weir jetty. A weir is a purposeful low section in a jetty located near the shore. A typical weir is about 300 m long and with elevation about mean sea level or lower. A weir section is always designed together with a deposition basin that collects sediment passing over the weir before it reaches the navigation channel (Seabergh 1983; Weggel 1983). Sand can then be pumped from the relatively sheltered area of the weir to the downdrift beach.

Many large jetties have been in place for more than a century. As a result, they may induce a regional coastal response, similar to the response of a beach to a natural headland, producing updrift accretion and progradation of the beach, and downdrift erosion and shore recession. Construction or extension of jetties will also tend to move ebb-tidal deltas further offshore from their natural position (Pope 1991), with portions of the original shoal that is located to the side of the narrow ebb current tending to return to shore.

Breakwaters

A breakwater is an engineered offshore structure that protects a harbor, anchorage, beach, or shore area by creating a sheltered region of reduced wave height. Breakwaters are always sufficiently high to afford protection against storm waves, and they are typically constructed parallel to the shore. A general principle is to align them to create the greatest shadow zone for either the direction of the predominant waves or the direction of the larger incident waves. Breakwaters must withstand direct impact of waves and are constructed of large stone or pre-cast armor units. Because breakwaters tend to be long and high, sediment moving along the beach can collect behind the structure.

In the United States, breakwaters are typically constructed as rubble stone mounds or with pre-cast concrete armor units that come in various shapes. In some countries, especially those where large stone is unavailable, caisson breakwaters are common. Caissons are pre-cast hollow concrete

containers that are towed on site and then sunk in place by filling them with sand and small stone.

In coastal terminology, a mole is a massive solid-fill protective structure extending from the shore into deeper water, formed of masonry and earth or large stones, and serving as a breakwater or a pier. A mole is not designed to stabilize an inlet or river mouth and typically functions as a breakwater and part of harbor infrastructure.

Cross-References

- [Dikes](#)
- [Engineering Applications of Coastal Geomorphology](#)
- [History, Coastal Protection](#)
- [Longshore Sediment Transport](#)
- [Shore Protection Structures](#)
- [Shoreline Response to Littoral Drift Barriers](#)

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Nearshore Geomorphological Mapping

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Definition

Coastal (nearshore) seafloor geomorphological mapping is based on the recognition of submarine features, which in

turn make up bathymetric patterns interpreted in terms of topography of marine origin or the recognition of drowned terrestrial landforms. Mapping procedures were initially or historically based on visual reconnaissance in clear waters, lead and line sounding to make nautical charts, and in the twentieth century by echo sounding. However, current techniques include mostly remotely sensed imagery from aerial photographs, airborne digital remote sensing (laser depth sounding), and satellites. Results of modern mapping procedures are depicted in coastal seafloor geomorphological maps that usually interpret bathymetric patterns through cognitive and computerized autoclassification techniques (e.g., ArcGIS).

Introduction

Widespread mapping of seabed topography was made possible through the introduction of seismic methods during World War II. Methods have improved dramatically since then and large surface areas of the deep-sea floor and continental shelf are now routinely mapped as, for example, in the Geological Long-Range Inclined Asdic (GLORIA) project based on long-range sidescan sonar surveys used by the US Geological Survey to map large segments of the Exclusive Economic Zone (EEZ) at depths generally greater than 400 m. The digital system, designed at the Institute of Oceanographic Sciences (now the Southampton Oceanography Center, Challenger Division, United Kingdom), was specifically designed to map the morphology and texture of seafloor features in the deep ocean. The GLORIA imagery, for example, displays a plethora of subsea geologic and geomorphological features such as volcanic edifices, fault scarps, channels, levees, slump scars, and crustal lineaments. Textural and tonal differences in the digital imagery show large sediment bedforms and varying sediment types. Ironically, more information has been collected from the outer continental shelf and deep-sea floor than from shallow water close to shore.

The nearshore zone is often neglected due to the difficulty of working in this technically hostile environment, which is characterized by spatially and temporally dynamic biophysical conditions of high-energy coastal ecosystems, irregular and shallow depths, variable and dangerous surf conditions, and strong currents. There are thus many dangers associated with marine survey in shallow waters, the surf zone being particularly notorious for both men and equipment. With the advent of modern high-resolution satellite imagery (e.g., SPOT, IKONOS), airborne sensor platforms such as LIDAR (Light Detection and Ranging), and development of digital aerial photography that can be georectified on the fly, mapping nearshore submarine topography becomes not only feasible but also practical for multipurpose applications. Passive

satellite and photographic sensors that detect bottom reflected radiation restrict the method to clear, nonturbid Class II waters. Turbid nearshore waters along muddy coasts are unsuitable for classification of bottom types using satellite imagery and aerial photography. Muddy coasts, which are commonly associated with tropical mangrove systems and temperate salt marshes, are estimated to occupy about 75% of the world's coastline between 25°N and 25°S Latitudes (Fleming 2002). These regions and other muddy coasts mainly associated with large deltaic environments (Wang and Healy 2002) are generally excluded from satellite and airborne imagery. Coral reefs, sandy beach, and rock platform environments present more favorable settings where nearshore turbidity is minimal for at least part of the year. Under clear water conditions, maximum depth penetration is at least 30–40 m and so mapping can proceed in a shore-parallel swath of variable distance offshore, depending on the shoreface gradient.

Active remote sensing platforms such as airborne LIDAR are able to detect changes in bathymetry under a variety of conditions where the nearshore zone may be quite turbid. These developing methods, which allow rapid acquisition of topographic data, provide enormous amounts of information in the form of digital elevation points (laser repetition rates of 10,000 pulses per second can, for example, provide elevations of surface points with a nominal spacing of 2 m; RMS errors of less than 10 cm can be achieved for elevation points; overlapping lines can be flown to minimize nonscanned coastal segments and increase point density). These methods enable large-scale quantitative mapping of coastal features and are more cost effective than conventional surveying techniques. High-accuracy airborne LIDAR mapping, based on recent advances in Global Positioning System (GPS), Inertial Measurement Unit (IMU), laser ranging, and microcomputers, has been applied to wave, beach, sea cliff, coastline position, and nearshore bathymetric surveys (e.g., Hwang et al. 1998; Brock et al. 1999; Krabill et al. 2000; Stockdon et al. 2002). The SHOALS system (a combined bathymetric/topographic LIDAR data source based on the acronym: Scanning Hydrographic Operational Airborne Lidar Survey) is another similar application of scanning laser mapping in the coastal zone that has found particular application in the vicinity of inlets and passes (e.g., Irish and Lillycrop 1997, 1999). A 13-km²-test area in Florida Bay, for example, was surveyed in just 12 h using SHOALS to demonstrate that airborne LIDAR bathymetric technology can be a valuable and cost-effective tool for surveying large shallow water areas (Parson et al. 1997). Complex morphologic features identified in Florida Bay included extensive shallow water networks of mud banks, cuts, and basins, among other features. Instead of using contours or isobaths to show terrain variability, modern computer displays associated with these laser-mapping systems often employ digital color ramps that gradationally distinguish one elevation

from another by variation in colors. The result is a colorful map that shows transitional gradations in relief. Additionally, a technique of using false “sun illumination” can highlight artifacts, error sources, and features that are not always evident on color-coded bathymetry (e.g., Hogarth 2002). These maps convey maximum information when they are combined with morphologic units that are interpreted and sea-truthed (i.e., based on information derived from visual observation, grab samples, jet probes, vibracores, etc.) from satellite or airborne imagery. Still other digital mapping procedures that are proving to be extremely useful in coastal areas are those that can now merge land-based topographic with marine-based bathymetric data into a seamless digital elevation model (DEM) (e.g., Milbert and Hess 2001; Gesch and Wilson 2002). The advantages of integrating USGS topographic data and NOAA hydrographic data in coastal DEM areas are recognized as essential components of key baseline geospatial datasets for geologic, biologic, and hydrologic studies in the vicinities of tidal inlets and passes, lagoons, and estuaries. Highly accurate digital terrain models (DTMs), derived using high-accuracy photogrammetric techniques provide spatial richness that supports

detailed study of coastal dune and beach processes (e.g., Judge and Overton 2001).

Although high-resolution surveys in shallow waters are based on a wide range of sensor equipment and platforms (e.g., Gorman et al. 1998; Byrne et al. 2002), there is still a place for interpretation of aerial photography, especially for very detailed close order work in inshore areas. Aerial photography has a long history of application to document coastal features and of use in resolution of coastal problems such as flooding and erosion. Indeed, many coastal features are seen in some of the earliest commercially available aerial scenes as oblique images, as shown in the 1920s image of a long barrier spit in Broward County, Florida, near the present city of Fort Lauderdale (Fig. 1). The early imagery in this case is extremely important to studies of coastal geomorphology because without this information, it is likely that the true coastal barriers along this coast would have been overlooked and ignored. Studies by Finkl (1993), for example, emphasized the crucial role of aerial photography for interpreting coastal morphology. Dietz (1947), for example, was among the first researchers to present a comprehensive analysis of the



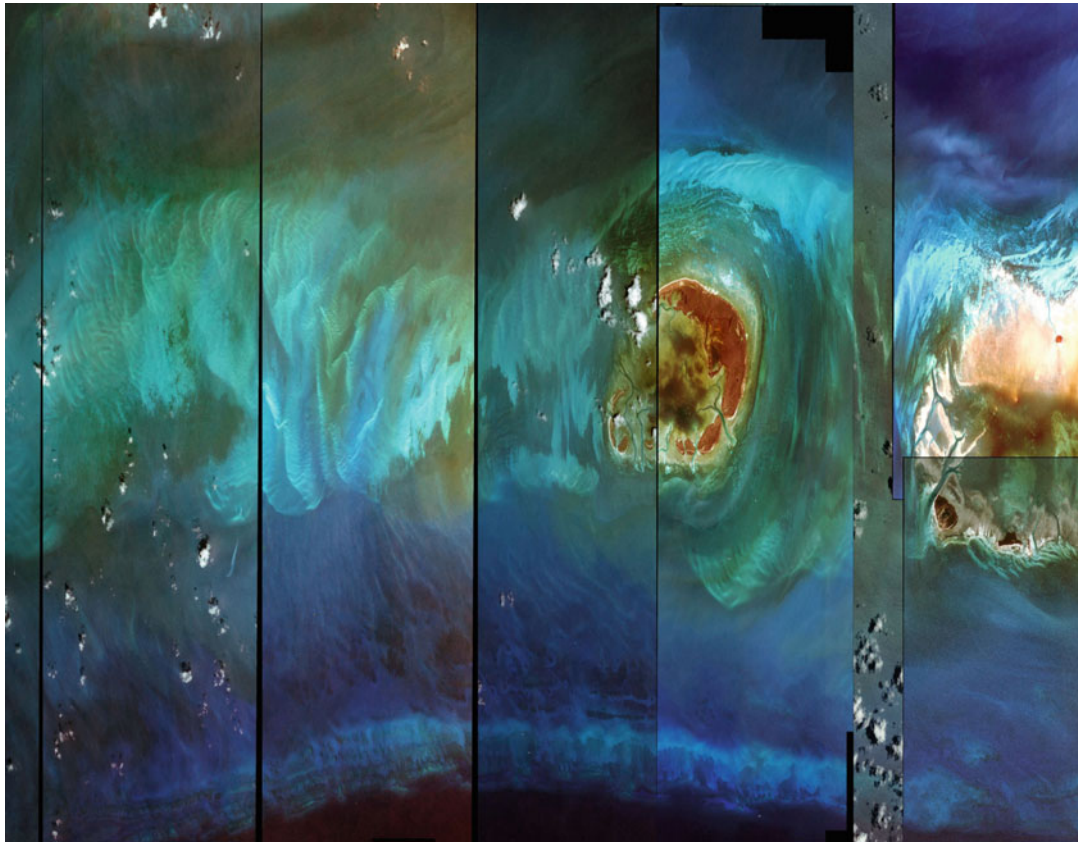
Nearshore Geomorphological Mapping, Fig. 1 Bay mouth bar and barrier spit near Fort Lauderdale, Florida, ca. 1925 (looking from east to west). The shallow open freshwater body on the right side of this oblique aerial photograph is Lake Mabel, now converted into the Port Everglades Turning Basin. A 9-km-long barrier spit, situated approximately 100 m offshore, continued alongshore from Fort Lauderdale to Dania passing in front of the bay mouth bar. Inlet cutting in Fort Lauderdale beheaded the updrift part of the barrier spit turning it into a barrier island that migrated shoreward over a 3–5-year period. During the migratory phase or barrier island rollover, approximately $6\text{--}10 \times 10^6 \text{ m}^3$ of sediment was moved

shoreward (Finkl 1993). Shown here is welding of the barrier spit to the bay mouth bar. The remaining southern (left side of photo) portion of New River Sound opens to a temporary natural inlet. By 1930, the entire barrier island (formerly a barrier spit) was welded to the mainland. When the Intracoastal Waterway (ICWW) was dredged through the tidal marsh (left center of photo), the ICWW-beach tract of land was mistaken for a barrier island and is still so designated today. (Partial scanned image from an original 1925 print by Fairchild Aerial Surveys, Inc., Long Island City, New York)

importance of aerial photographs in the study of shore features and processes. He pointed out the scientific value of aerial photographs in studying underwater features. A cogent and informative summary of aerial photography in the coastal zone is provided by El-Ashry (1977), who collected a range of examples covering multiple applications in different parts of the world. A more recent example of comprehensive use of aerial photographs may be found in Owens (1994) where aerial surveys covered over 40,000 km of Canadian coastline to assist in the compilation of oil spill response manuals. Oblique and vertical aerial photographs are used throughout the manual to illustrate different types of coastal environments, morphologic features, and shore processes. Other examples of new technologies, advanced techniques, and new applications (not limited to aerial photography) are found in a series of international conferences on remote sensing for marine and coastal environments sponsored by ERIM, a subsidiary of Veridian, Inc. (PO Box 134008, Ann Arbor, MI 48113). Vertical, stereo-paired aerial photographs find many applications, singly or in combination with other techniques, in coastal studies not only to relatively long-term shore variation in coastal planforms or coastline positions (e.g., Smith and Zarillo 1990; Shoshany and Degani 1992; Thieler and Danforth 1994; Moore 2000), but also in studies of short-term coastline changes that use fully automated techniques for determining the high-water line (e.g., Shoshany and Degani 1992). Jiménez et al. (1997), for example, used temporally close order aerial photography (a series of seven aerial surveys over 4 months) to analyze highly dynamic coastal features in very flat, microtidal deltaic areas in Spain. Acquisition systems that collected temporal data to monitor the behavior of a dissipative multiple bar system in a nearshore zone near Terschelling, The Netherlands, were based on vertical aerial photogrammetry and echo sounding (Ruessink and Kroon 1994) as were similar efforts that monitored nearshore submarine morphology at Duck, North Carolina (Guan-Hong and Birkemeier 1993), and along the Wanganui coast, New Zealand (Shand et al. 1999). Alam et al. (1999), by way of another example, used high-resolution vertical black and white aerial photographs (1:20,000), using a zoom transferscope, to prepare a preliminary terrestrial geomorphological map of Cox's Bazar Coastal Plain, southeast Bangladesh.

As the space program was born in the second half of the twentieth century, so was a new coastal imaging acquisition technology with the use of orbiting satellites. Through the utilization of hyperspectral and multispectral sensors, satellites provide a continuous stream of coastal images without the logistical hardships of deploying a vessel or aircraft every time visual renderings are to be taken. Instead of acoustic or light reflectance, satellite sensors create an image-based visual approach to discerning physical and biological bottom features of the nearshore benthic substrate. Typically, hyperspectral sensor datasets constitute a range of 100–200 spectral

bands of relatively narrow bandwidths (5–10 nm). On the other hand, multispectral sensor datasets are only composed of a few spectral bands ($n = 5–10$), but have a relatively large range of bandwidths (70–400 nm). The visual detection of seafloor marine features is then dependent on the spectral coverage of the spectrometer and the overall spectral resolution (i.e., the pixel size of the satellite image covering the earth's surface) of the acquired images. There are many satellite sensors currently in orbit around the Earth today, some of which maintain a high spatial resolution (i.e., 0.6–4 m), while others utilize a medium spatial resolution (i.e., 4–30 m). For example, the IKONOS satellite, which was launched in 1999, provides high-spatial resolution images on a global scale (Dial et al. 2003). Achieving a 0.8-m panachromatic resolution and a 3.2-m multispectral resolution, IKONOS uses five spectral bands that include blue, green, red, near infrared, and panachromatic. Many marine habitat studies have used IKONOS satellite imagery as the basis of their mapping efforts (e.g., Maeder et al. 2002; Mumby and Edwards 2002; Dial et al. 2003; Hochberg et al. 2003; Palandro et al. 2003; Finkl and Vollmer 2011 (Fig. 2); Steimle and Finkl 2011; Makowski 2014; Finkl and Makowski 2015; Makowski et al. 2015, 2016, 2017) (Fig. 3). Andrefoet et al. (2003) incorporated multiple IKONOS satellite images of the Arabian Gulf, Indian Ocean, Indo-Pacific, Pacific, and Caribbean biogeographic zones in order to map geomorphologic zones (e.g., reef flats, forereef, patch reef, lagoon) and biological communities (e.g., seagrass beds, macroalgae coverage, coral overgrowth). Their mapping results helped to justify the use of IKONOS satellite supplied images as a means to appropriately interpret and classify tropical coral reef environments. Similarly, Maeder et al. (2002) utilized the blue, green, and red spectral bands from IKONOS images of one fixed location offshore of Roatan Island, Honduras (Central America) to maximize the water-penetrating capabilities of the satellite sensor and map the general coral reef structure of the nearshore benthos. Maeder et al. (2002) helped to prove that the geomorphological structure of a typical Caribbean coral reef and the biota growing upon it could be mapped using the high spatial and radiometric resolution of the IKONOS imagery. In addition to IKONOS, there are many other satellite sensors (e.g., ASTER, LANDSAT-7, SPOT-5, etc.) that provide high or medium spatial resolutions, and subsequently, a multitude of remote sensing studies have used these kinds of images for their nearshore geomorphological mapping efforts (e.g., Finkl and DaPrato 1993; Bour et al. 1986; Dobson and Dustan 2000; Hochberg and Atkinson 2000; Andrefoet et al. 2001; Klemas 2011a, b; Capolsini et al. 2003; Hochberg and Atkinson 2003; Joyce et al. 2004; Purkis and Riegl 2005; Yamano et al. 2006; Muslim et al. 2007; Everitt et al. 2008; Makowski 2014; Makowski et al. 2015).



Nearshore Geomorphological Mapping, Fig. 2 A composite image of multiple IKONOS satellite scenes taken over the Southern Key West National Wildlife Refuge, Florida, USA. In order to map sub-bottom features effectively, different IKONOS image frames were selected for this mosaic that voided cloud cover, waves, turbidity, and algal blooms.

Finkl and Vollmer (2011) used these IKONOS images to map the Marqueses and Quicksands regions of the Southern Key West National Wildlife Refuge, and their results were published in the *Journal of Coastal Research* (JCR). (Source: Finkl and Vollmer 2011)

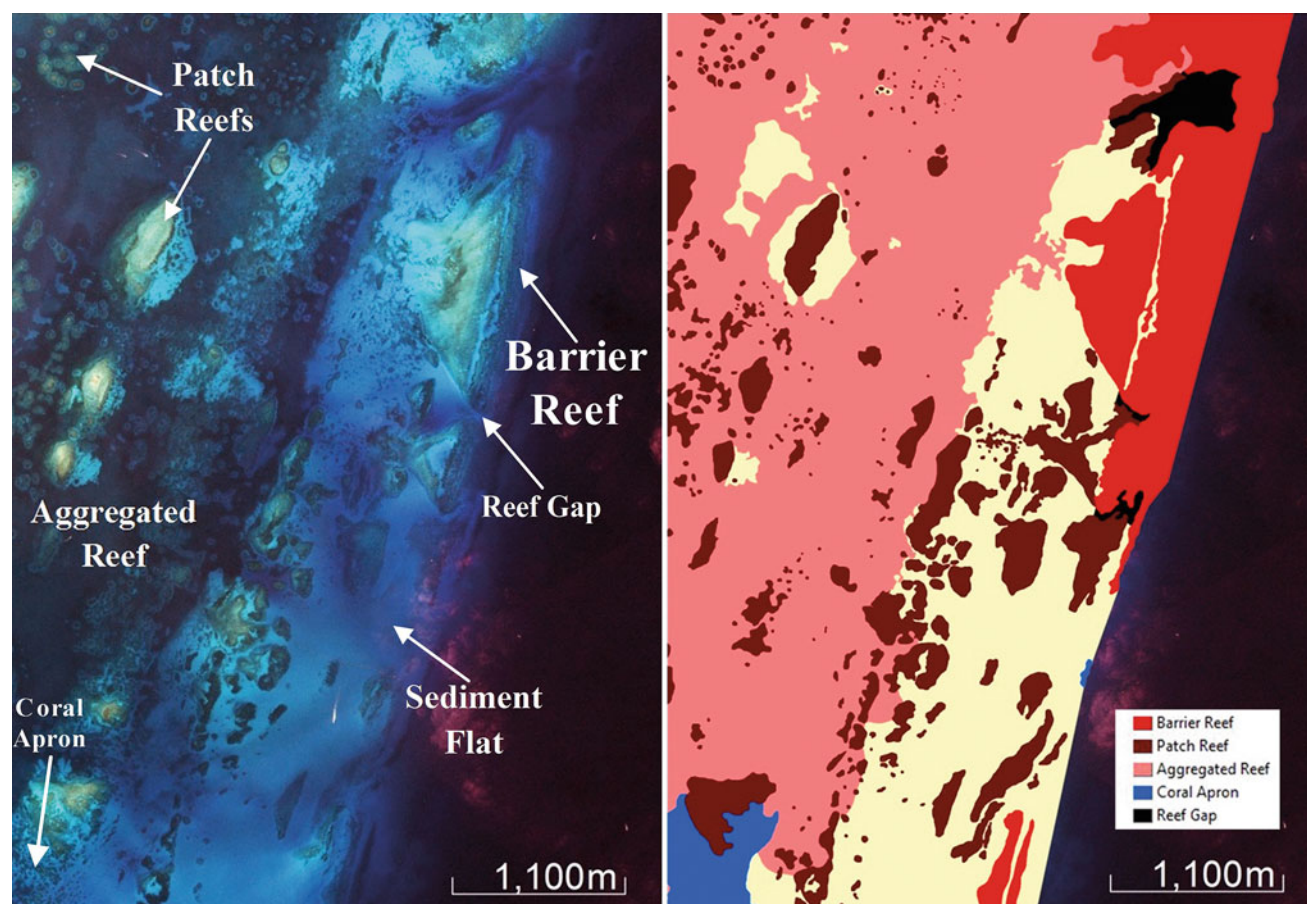
Although the subject matter of nearshore geomorphological mapping is global in extent, this discussion necessarily focuses on the use of digital aerial photography for submarine geomorphological mapping along the nearshore zone of southeastern Florida. The example elucidated here is representative of many coastal areas that contain sandy beaches and rocky shores that front seabeds characterized by hardgrounds and unconsolidated sediments. The methodology is thus not restricted to tropical and subtropical carbonate environments, but applies to mid- and high-latitude coasts with low nearshore turbidity.

The Nearshore Environment of Southeast Florida

Some of the more obvious topographic features of the inner continental shelf along the southeast coast of Florida (Fig. 4) were appreciated several decades ago when they were mapped by the US Army Corps of Engineers as hardgrounds and inter-reefal sand flats (Duane and Meisburger 1969). The shore-parallel “reef” (comprised by rock and coral-algal

components) system, composed of inshore exposure of the Pleistocene Anastasia Formation (a cemented, quartzitic, molluscan grainstone that formed in beach and shallow-water nearshore environments) (Stauble and McNeil 1985; Davis 1997) and coral-algal reef tracts (referred to as the Florida Reef Tract) (Lidz et al. 1997), increases in depth offshore as a giant staircase. These parabathic reef tracts extend southwards into the Florida Keys and represent approximate positions of paleocoastlines extensionally offshore and vertically within particular tracts as prior sea-level stands were revisited through time (Finkl 1993, 1994). Sedimentary troughs that contain admixtures of clean, free-running sands, discontinuous lenses or stringers of silts and clays separate the reef tracts, or carbonate rubble accumulations deposited in association with paleo-inlets that cut through the reef tracts.

Most of the Holocene coast was characterized by extensive spits that extended downdrift and alongshore for tens of kilometers (cf. Fig. 1) (Finkl 1993, 1994). The spits, which rose several meters above mean sea level, were stabilized by herbaceous and phanerophytic vegetation and protected



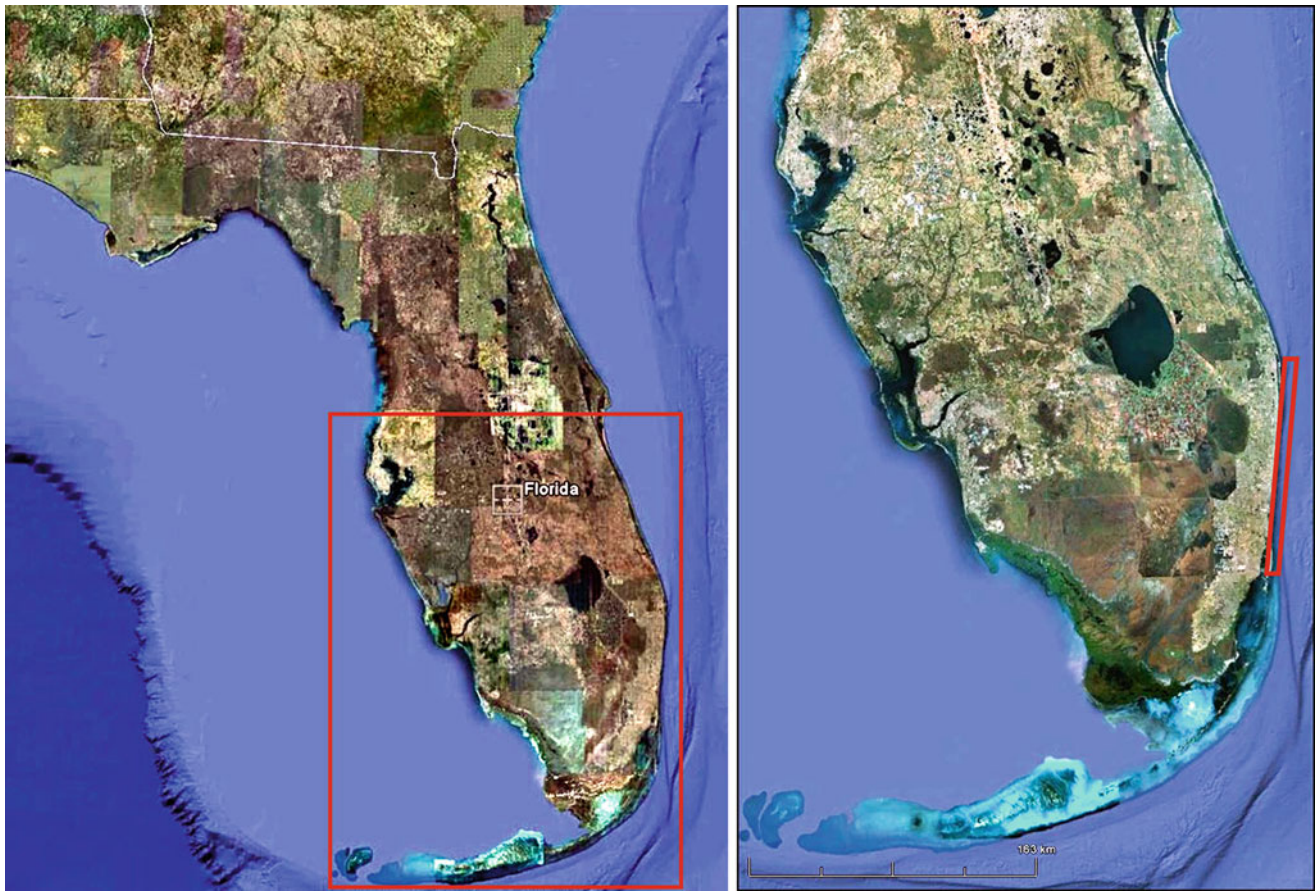
Nearshore Geomorphological Mapping, Fig. 3 Example of a near-shore geomorphological mapping effort using IKONOS satellite imagery (from Makowski et al. 2017). In order to interpret and classify the separate components of the Coral Reef Geoform, the left panel shows associated landform examples of Barrier Reef, Patch Reef, Aggregated Reef, Coral Apron, and Reef Gap within the IKONOS-2 satellite image. The right panel overlays color-coordinated classification units, as shown in the legend, on top of the IKONOS-2 to visually demonstrate the spatial distribution of landforms. In addition to the colors and units

provided in the legend, yellow areas in the right panel represent the Planar Bed Forms and Ripples Landform, which is a subclassification unit of the Sediment Flat Geoform. The blacked out portion on the right-hand side of both panels represents where the water depth exceeded the threshold by which the IKONOS-2 sensor could visually register any benthic spectral signatures. The satellite image and interpretive overlay panel were exported out of ArcMap program at a nominal scale of 1:24,000 within the layout view feature. (Source: Makowski et al. 2017)

shoreward sounds leading to coastal bays and estuaries. Inlet cutting and stabilization in the early 1900s initiated the demise of barrier spits (cf. Fig. 1) and islands by jetties and canalization of coastal wetlands, as described by Finkl (1993), to form the present shore. Bedrock of the Anastasia Formation is exposed onshore or buried at a shallow depth below present-day beaches (Fig. 5). Most berms contain beach sands less than 2 m in thickness so that some beaches are stripped of sediment during storms to expose the underlying bedrock during part of the year (Finkl 1994). Dunes fronting back beaches were commonly leveled for high-rise development so that today, incipient dunes only develop where buildings or infrastructure is setback from the shore. Seawalls that preclude dune formation back many beaches along this developed shore.

Morphologic Features

The methodology employed in the identification of coastal geomorphic features was based on interpretation of digital, color aerial photographs at an acquisition scale of 1:3,900. In addition to on-screen digitizing of black and white images (derived from scanned color prints to a resolution of 300 dpi), the stereo-paired color prints were also utilized to enhance visual inspection of details and to resolve complicated photo patterns. Most photographs showed about 25% land and 75% ocean to provide dune-beach features and assist with location on the ground. Prior to mapping, 257 aerial photographs (23 × 23 cm color photographic prints) for the Palm Beach County shore were visually perused to determine the range of coastal features present



Nearshore Geomorphological Mapping, Fig. 4 Location diagram showing the Florida peninsula within the conterminous United States (left panel) and the general Florida study area in Palm Beach, Broward,

and Miami-Dade counties along the southeast coast (right panel, red box). (Source: Google Earth)

N

in the study area. Compilation of a summary list of geomorphic features was organized as a general or organizational classification scheme with major categories for beach, bar and trough, dunes, rock outcrops, seabed morphosedimentary features, geologic structures, suspended sediments, and engineered structures. The morphologic forms and engineered structures that were noted are listed in Table 1.

Identification of the various morphologic features was based on direct recognition image interpretation strategy, but also employed field observations, information from grab samples and drill core, interpretation by inference, and probabilistic interpretation. Campbell (1996) provides a cogent summary of different image interpretation strategies and details disciplined procedures that enable the interpreter to relate geographic patterns on the ground (and under water) to their appearance in the image. Subaerial features such as dunes and infrastructure are identified by discretely different visual information. Because much of the native dune system has been destroyed by construction, many areas are interpreted through a priori knowledge. Visible

dune sands and dune vegetation in other areas permitted application of a direct recognition strategy. Features in the backshore regions were identified by field observation and direct recognition. Beach berms and wrack lines were easily identified on the photographs due to distinct tonal variations. Interpretation of foreshore features was somewhat more difficult because this zone is partially submerged and acted upon by swash. The beachface is a predominant feature of the foreshore and can be conveniently divided into the upper (light-toned) and lower (dark-toned) beachface. Interpretation of geomorphic features of the inshore is sometimes hindered by complexity of spatial patterns. Sharp tonal contrasts facilitate interpretation of bars and troughs, which are the main sedimentary components of the inshore zones. Hardground features are identified by their dark tones, regular structural (lineament) patterns, and coarse texture of the rock surfaces. Many of the offshore features are identified using probabilistic interpretation as well as direct recognition. A priori knowledge of the existence of coral reefs (as mapped by Duane and Meisburger 1969) in the offshore zone increases the probability of seeing reefs in the aerial



Nearshore Geomorphological Mapping, Fig. 5 Erosion of perched beach and dunes in Boca Raton, central Palm Beach County, Florida, showing exposure of bedrock. (a) Remains of a subaerial beach that has been stripped off the underlying rock platform and cut into the Anastasia Formation by an Atlantic Northeaster winter storm. Wave attack from the northeast tends to erode beaches and deposit the sediments offshore. (b) Rocky platform that underlies the perched beach shown in (a), upper part of photo center. (c) Reentrant surge channel that has cut across the

berm and is shown here nipping foredune sediments. The surge channel, which follows a cross-shore fracture in the bedrock, brings turbulent water to the backshore. Note the salient environments along this eroded shore where perched beaches (a) with berm thicknesses generally less than 2 m are stripped away by storm surge to expose rocky platforms (b) that lie at about mean sea level and how ingress is facilitated by structural weakness in the bedrock (c). (Photo: Lindino Benedet)

photographs. Smooth tonal variations usually permitted differentiation of sandy bottom features from rock outcrops that are associated with disrupted blocky patterns and coarse textures.

Although a variety of interpretive techniques are available to assist in the identification of landforms in the coastal zone, it is necessary that the persons preparing the morphologic map have knowledge of coastal features prior to embarking on a project such as this. Preparation of a comprehensive legend before mapping (cf. Table 1) assists in the rapid completion of the project. Modification of some morphologic units can take place as mapping proceeds, however.

The preceding paragraphs briefly outline some of the salient techniques for interpreting coastal morphologic features from vertical, stereo-paired aerial photography. Interpretation assumes familiarity with landscapes and seascapes, including polygenetic landform sequences that are so common along the world's coastline. There is no substitute for careful fieldwork, which must at some stage precede interpretive efforts that attempt to deduce what is on the ground from aerial imagery. Of the numerous reference works that show in pictorial or graphic formats physical features of the coast, the following are noteworthy for their wide scope of

topical coverage, geographic focus, clarity of presentation, or detailed orientation to specific classes of morphologic features: Bird (1976), Bradley (1958), McGill (1958), King (1959), Russell (1967), Zenkovich (1967), Emery and Kuhn (1982), Schwartz (1982), Pethick (1984), Trenhaile (1987), Guilcher (1988), Hopley (1988), Nordstrom (1990), Nordstrom et al. (1990), Short (1993), Forbes and Syvitski (1994). In contrast to these general reference works are those that focus specifically on the southeast Florida coastal zone viz. Duane and Meisburger (1969), White (1970), Finkl (1993), Finkl and DaPrato (1993), Finkl (1994), Finkl and Esteves (1997), Finkl and Bruun (1998), Brown (1998), Khalil (1999), Warner (1999), Finkl and Khalil (2000), Finkl and Benedet (2002), etc. Prager and Halley (1997), while conducting a survey of bottom types in Florida Bay using sequential aerial photography, studied the distribution of sea grass and its influence on wave energy impacting mud banks. Bottom types in this shallow, low-energy embayment that are identified by Prager and Halley (1997) include: open sand, hard bottom, open mud, mud bank suite, mixed bottom suite, sparse sea grass, intermediate sea grass, and dense sea grass. The subtropical, open ocean coast of southeast Florida provides a wider range of morphologic types than might

Nearshore Geomorphological Mapping, Table 1 Classification of salient coastal morphological forms and engineering structures identified from large-scale aerial photographs for the southeast coast of Florida, centered on Palm Beach and Broward counties

Level 1 ^a	Level 2 ^b	Level 3 ^c
Beach	Berm	
	Cusp	
	Beachface	Lower beachface
		Upper beachface
		Sub-beach step
Bar-and-trough topography	Channels in bar	
	Crescentic bar, trough	
	Longshore bar	Welded
		Transverse
	Rip channel	
	Low tide terrace	
	Trough	Beachface trough
		Shoreface trough
		Crenulated trough
		Infilled nearshore trough
Dunes	Undifferentiated dunes	
Rock outcrop (intertidal and submarine)	Abrasion platform	Beachface outcrop
	Coral-algal reef (reef tract)-Hardground (rock reef) [Complex unit]	Outer coral-algal rock reef
		Fluted rock reef
		Fluted offshore rock reef
		Parabathic hardground stringer
Sedimentary morphological features	Sandy bottom	Light-toned running sand
		Dark-toned running sand
		With rock outcrop
		Featureless sandy bottom
		Structurally controlled sandflat
	Tidal delta (ebb, flood shoal)	
Suspended sediments	Turbidity plume	
Engineering structures	Dredged inlet Seawall Rock revetment Groin Pier Submerged breakwater Jetty	

^aLevel 1 groups the most general morphologic forms and engineering works

^bLevel 2 categorizes the broad forms of Level 1 into specific types

^cLevel 3 incorporates specific types of Level 2 that have distinguishing characteristics that permit further differentiation

at first be anticipated for a coast that is commonly regarded as a sediment-rich, sandy beach shore. The frequent occurrence of rock forms, both subaerial and submarine, lends diversity to an otherwise potentially monotonous stretch of shore. Examples of morphologic units occurring in this subtropical zone of the western North Atlantic Ocean are first described in terms of the hard rock units comprised by coral and algal reef materials and lithified coquina and sandy deposits. These hard rock morpho-units are emphasized because antecedent topography influences present coastal configuration in Florida (e.g., Evans et al. 1985; Kelletat 1989; Finkl 1993; Finkl and DaPrato 1993; Finkl 1994; Lidz et al. 1997), and as described for other areas, for example, by Sanlaville et al. (1997), Pilkey (1998) and Dillenburg et al. (2000). The occurrence and distribution of sedimentary geomorphic features is then considered, often in relation to the geologic framework of these coastal environments in southeast Florida.

Reef Tract and Hardground Morphologic Features

The term *hard shore* refers to the occurrence of sand, gravel, cobbles, boulders, or bedrock as opposed to the antonym *soft shore*, which identifies shore composed of peat, muck, mud, soft marl, or marsh vegetation (Bates and Jackson 1979). *Hardground*, on the other hand, denotes a zone on the seafloor, usually a few centimeters thick, where the sediment is lithified to form a hardened surface that is often encrusted, casehardened, or solution-ridden. Colloquial usage in the Florida environment restricts hard shore and hardground to subaerial and submarine outcrops of bedrock or lithified sediments. The term hardground is further unrestricted in definition as it is used in reference to coral reefs, coral-algal reefs, and exposure of bedrock such as the Anastasia Formation. When corals or coral-algal reefs are known, explicit identifying terminology is applied but often, especially in remote

sensing analyses, differentiation of specific units is not possible without groundtruthing and hence the general term is applied. The term *reef tract* specifically refers to coral and coral-algal reefs whereas hardground refers to any hard surface, regardless of composition or origin.

Because coral reefs and coral-algal reefs, which comprise the Florida Reef Tract, and hardgrounds appear as many distinct forms on the aerial photographs, they are divided into different map units as follows: abrasion platforms, fluted inshore/offshore rock reefs, parabathic hardground stringer, coral-algal rock reef, and sandy bottom with rock outcrop. Linear disjunctive segments that parallel the coast make up what are locally referred to as the first, second, and third reefs that collectively characterize this part of the Florida Reef Tract. Sandy deposits cover parts of the coral reef and inshore Anastasia Formation.

Fluted Inshore/Offshore Rock Reefs

Several morphologically distinctive types of hardground occur in southeast Florida. Some coastal segments are characterized by *patterned hardgrounds* (rock reefs) with somewhat regularly spaced distinct cross-shore channels, referred to here as *flutes* (Fig. 6). The flutes average about 1–10 m in width and their length extends the width of the outcrops that can range upwards of 30–50 m. The flutes are clearly evident in the photograph (left) but the unit is mapped (right) without detailing the position of flutes. Some of the more prominent examples of fluted rock reefs are contiguous for more than 400 m in a shore-parallel direction. The flutes occur in outcrops of the Anastasia Formation that are often cracked or fractured into large blocks that make up a general reticulate pattern due to increased differential erosion of these zones of weakness. The small diabathic channels or flutes follow the initial or pre-existing structural depressions that were later subaerially etched into the carbonate rock surface prior to drowning by the Flandrian sea-level rise.

Abrasion Platforms

Abrasion platforms (Fig. 6), a type of shore platform (e.g., Pethick 1984; Trenhaile 1987; Sunamura 1992) that occur in the intertidal and inshore zone, are developed on outcrops of the Anastasia Formation, as described by Duane and Meisburger (1969) for the coast of southeast Florida. The platforms, which show signs of weathering, solution, and wave-induced erosion, are generally quite narrow (less than 50–100 m) and may extend intermittently alongshore for distances of 1,500 m or more. These calcareous platforms also occur as blocky, jagged outcrops arranged in sets along the shore and are separated by sandy flats, as shown in Fig. 6. Structural patterns (left photograph) in the Pleistocene

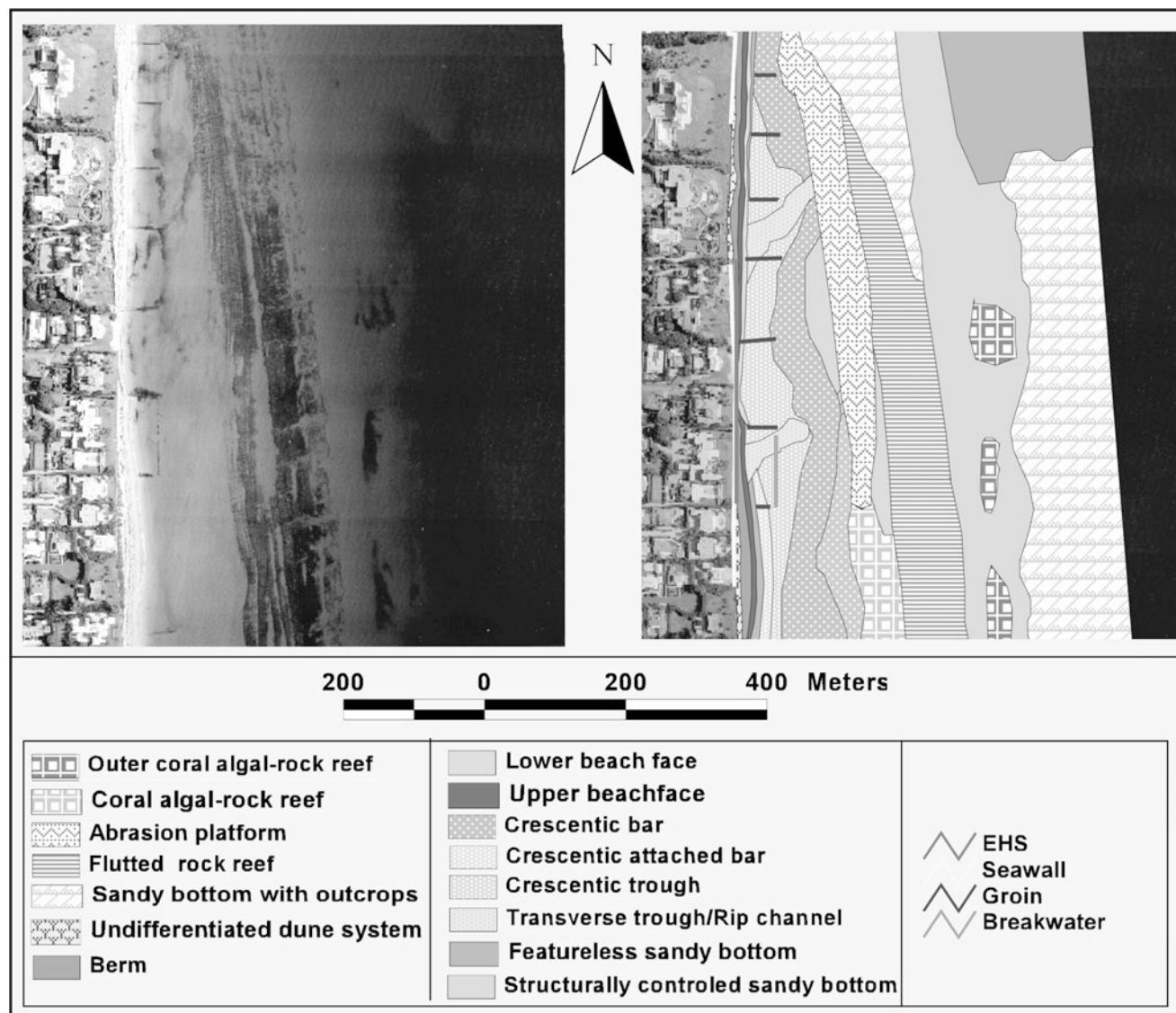
bedrock show clear shore-parallel lineations with sand partially filling chemically weathered depressions that lie below more resistant facies that have been abraded by wave action to produce concordance of ridge levels.

Parabathic Hardground Stringer

Another common type of hardground that occurs in this region is referred to as a parabathic hardground stringer (Fig. 7). These narrowly elongated morphological features are referred to as *stringers* because they tend to occur as long, ridged rock outcrops surround by sandy seafloor. These inshore rock outcrops are narrow parabathic forms that are usually less than 7–10 m wide but may be more than 200 m long. These stringers usually occur in groups and are separated by sandy deposits. These topographically positive seabed features represent more chemically resistant and structurally competent facies of the Anastasia Formation or coral-algal reef. The stringer shown here (dark lines in center of the left photograph) trends in a general north-south direction along the axis of strike of the now-lithified banks (including coquina beds) of the Anastasia Formation and is partly overlain by coral-algal rock reef (bottom center of photograph, left, and geomorphological map, right). Extensions of buried hardground that are emergent above surrounding sand formations, such as bars, often form hardground stringers.

Coral-Algal Rock Reef

The northern extension of the Florida Reef Tract (see Lidz et al. 1997), which extends northward from the Florida Keys, contains a seriously degraded coral reef environment that has been stressed by a variety of factors (Lidz and Hallock 2000) to the point where many coral communities have died and their skeletons now serve as foundations for algal growth. The algal encrustations form a kind of biogenic carbonate rock that is mixed with surviving corals; the complex morphological structure (coral reef plus algal encrustations) is thus referred to as a *coral-algal rock reef* in an effort to identify its true status as a conditionally stable but degrading coral reef system. Bottom morphologies associated with this unit are thus complex intergrades that are undifferentiated due to interpretive difficulties to consistently separate true corals from coral-algal units. Outcrops of the Anastasia Formation sometimes occur in close proximity to these biogenic units (see Figs. 6 and 7), further complicating the interpretive process. Varieties of these parabathic forms, which are mapped for simplicity as one unit, are visually contiguous as a distinct bottom type for distances ranging up to 4 or 5 km alongshore. Bottom sands separate successional coral reef systems (viz. the first, second, and third reefs) and also reefs from outcrops of the Anastasia Formation.



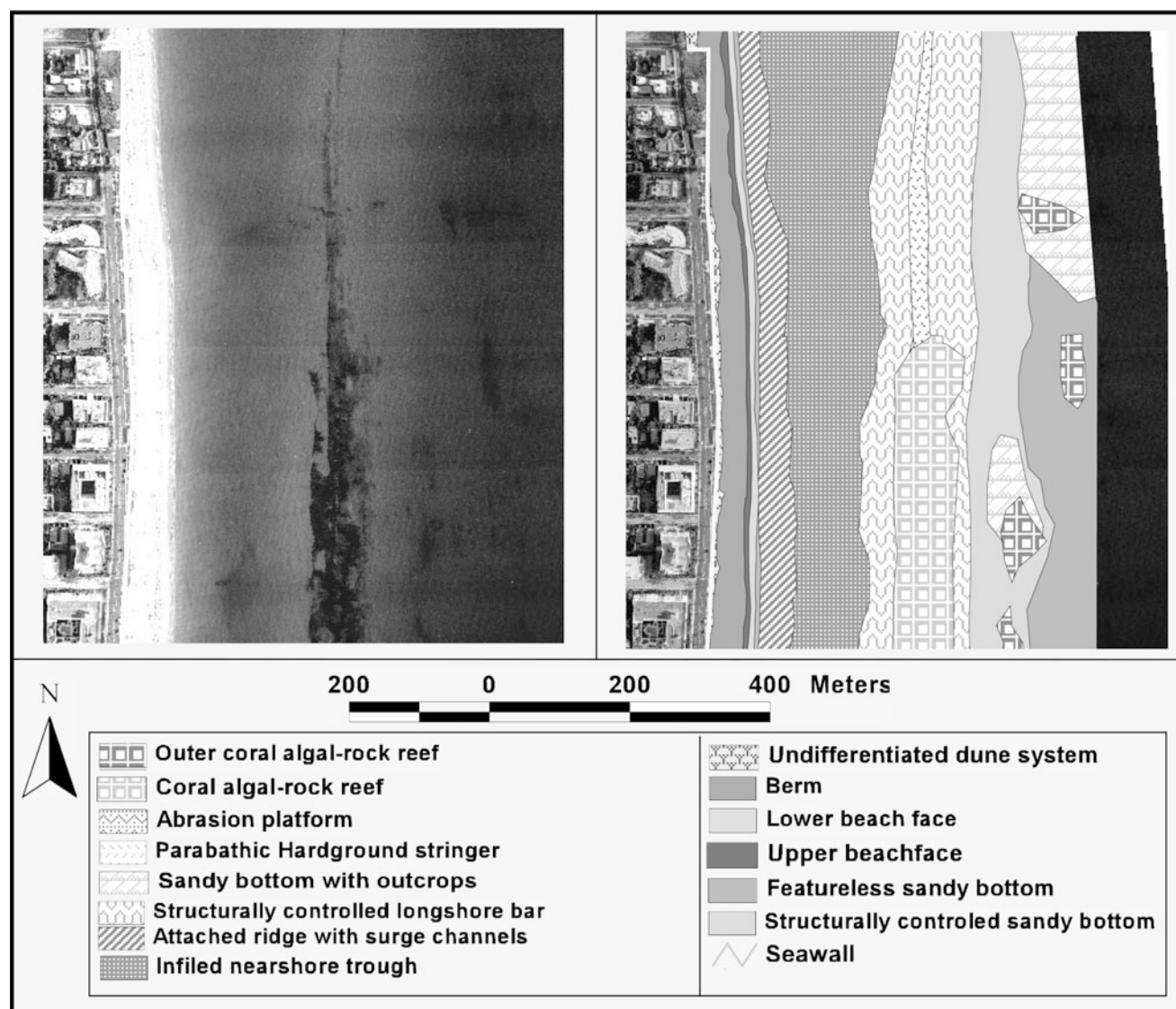
Nearshore Geomorphological Mapping, Fig. 6 Rock outcrops and sandy bottoms along the southeast coast of Florida in the Town of Palm Beach, Palm Beach County. The complexity of geomorphic units is seen in parabolic and diabolic spatial distribution patterns, that is, differentiation of bottom types alongshore and across the shoreface. The occurrence of rock units on the seafloor is a most striking feature that is denoted by several categories of hardground: coral-algal reef, abrasion platforms, fluted rock reefs. The map of geomorphic units (right), interpreted from the aerial photograph (left), identifies complex features associated with the landward-most portions of the Florida Reef Tract, first and second reef systems. Isolated blocks of coral-algal rock associated with the second reef system lie about 400 m offshore and poke through a thin veneer of sand that overlies slightly lower levels of the

same rock unit (referred to as structurally controlled sandy bottom). The main features of the first reef are a beveled abrasion platform, rough carbonate rock reef of the Anastasia Formation, and a seaward portion identified as fluted rock reef that contains diabolic channels. Sedimentary features, composed of mixed siliciclastics and biogenic carbonate sands, include sandy bottoms and bar-and-trough topography. The rhythmic wave pattern of a large crescentic bar-and-trough system migrates alongshore, landward of the first reef. Rip current channels pass through the crescentic trough in a general NE direction. The beach is subdivided into berm, upper beachface, and lower beachface. A groin field and an erosional hot spot (zone of accelerated shore retreat) are additionally identified on the map

Sandy Bottom with Rock Outcrop

This is a complex, intergraded bottom type (see Figs. 6, 7, and 8) with spatial patterns that are too intricate or relationally detailed to warrant separation at the mapping scale. The unit is essentially a variety of hardground bottom type with a veneer

of sand that ranges in thickness from a few millimeters to several tens of centimeters in thickness. The sandy veneer is so thin that structural features of the underlying rock are still visible, especially where slightly positive relief features emerge through the sand cover. Unconsolidated, free-running sands that overlie portions of the reef tract or intermittent rock



Nearshore Geomorphological Mapping, Fig. 7 Sandy shore with submarine rock outcrops along the southeast coast of Florida a few kilometers south of the Town of Palm Beach, Palm Beach County. The aerial photograph (left) and interpreted geomorphological map (right) are companioned here to show salient characteristics of this shore. This sandy coastal segment is characterized by beach (berm, upper and lower beachface), bars and troughs, parabathic hardgrounds (coral-algal and rock reef), and sandy bottoms. Bedrock, which is close to the surface of the seabed here, is a component that is noted in composite units viz. sandy bottom with outcrops (which shallowly flanks coral-algal rock reefs) and structurally controlled sandy bottom (which shallowly overlies bedrock). The dominant parabathic geomorphic distribution patterns

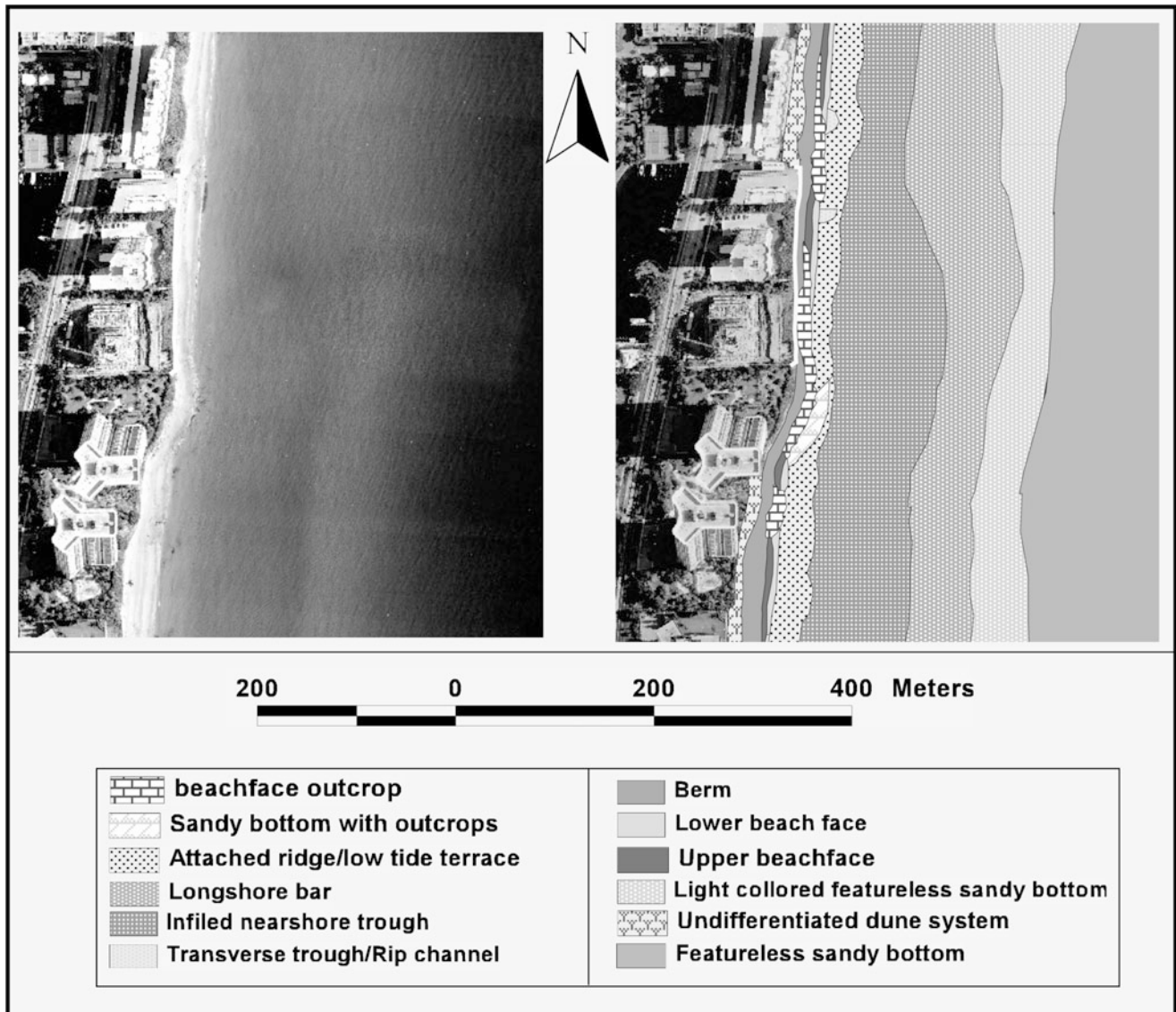
outcrops characterize these bottom-type features. Because corals that have been buried by sediment are now dead, they are simply referred to as carbonate rock. Rock outcrops occur primarily as parabathic features, but diabathic patterns with sediment between the outcrops can be found in some areas. Sedimentary infills immediately adjacent to the rock outcrops are relatively thin, usually less than 1 m thick. Sandy bottom with rock outcrop may occur in almost any geographic

reflect shore-parallel rock outcrops on the seafloor. The parabathic attached ridge (welded to the lower beachface) with diabathic surge channels merges seaward with a partially infilled nearshore trough that together makes up the sandy nearshore environment. Structural control of sedimentary features and geomorphic patterns is evident in the vicinity of the coral-algal rock reef materials of the first Reef System in the Florida Reef Tract, the landward boundary of which lies about 300 m offshore. Note the northward extension of the hardground as a parabathic hardground stringer (about 500 m in length) in the upper central portion of the map (right). The black area on the seaward-most portion of the geomorphological map is uninterrupted due to depth constraints

location along the shore as a reminder that bedrock or buried coral reef is never far away.

Sediment-Composed Morphologic Features

A large part of the inner continental shelf, loosely analogous to what some researchers refer to as the shoreface (e.g.,



Nearshore Geomorphological Mapping, Fig. 8 Example of beachface outcrops along the southeast coast near Boca Raton, Florida. This coastal segment shows beach sediments perched on top of bedrock (Anastasia Formation) that often helps to stabilize the shore and sometimes forms a natural salient, as seen in the lower left hand corner of the photograph (left) and in the interpreted map (right). Bedrock outcrops in parts of the low tide terrace where the sand cover is relatively thin and

lineations in the rock structure are evident. Transverse troughs or rip channels transect the low tide terrace in the upper part of the photograph and are shown in the map. A bar and trough system up to 200 m in width grades seaward into flat lying featureless sandy bottom. Inshore rock outcrops are a prominent hard bottom feature along the shore and their presence is a clear influence on coastal configuration, when seen in plan view

Wright 1995), contains sedimentary covers that are surficially modified by currents to form distinct morphologic units. The various kinds of morphologic features, composed of siliclastic and carbonate sediments, occur in beach sands or as bars and troughs. Under conditions of relatively stable or slowly rising sea level, typical of the last 6,000 years, the shoreface may approach conditions of steady-state equilibrium (as defined by Schumm 1991). On shorter time frames, instantaneous or event timescales, the upper shoreface is subject to continual change with relaxation times on the

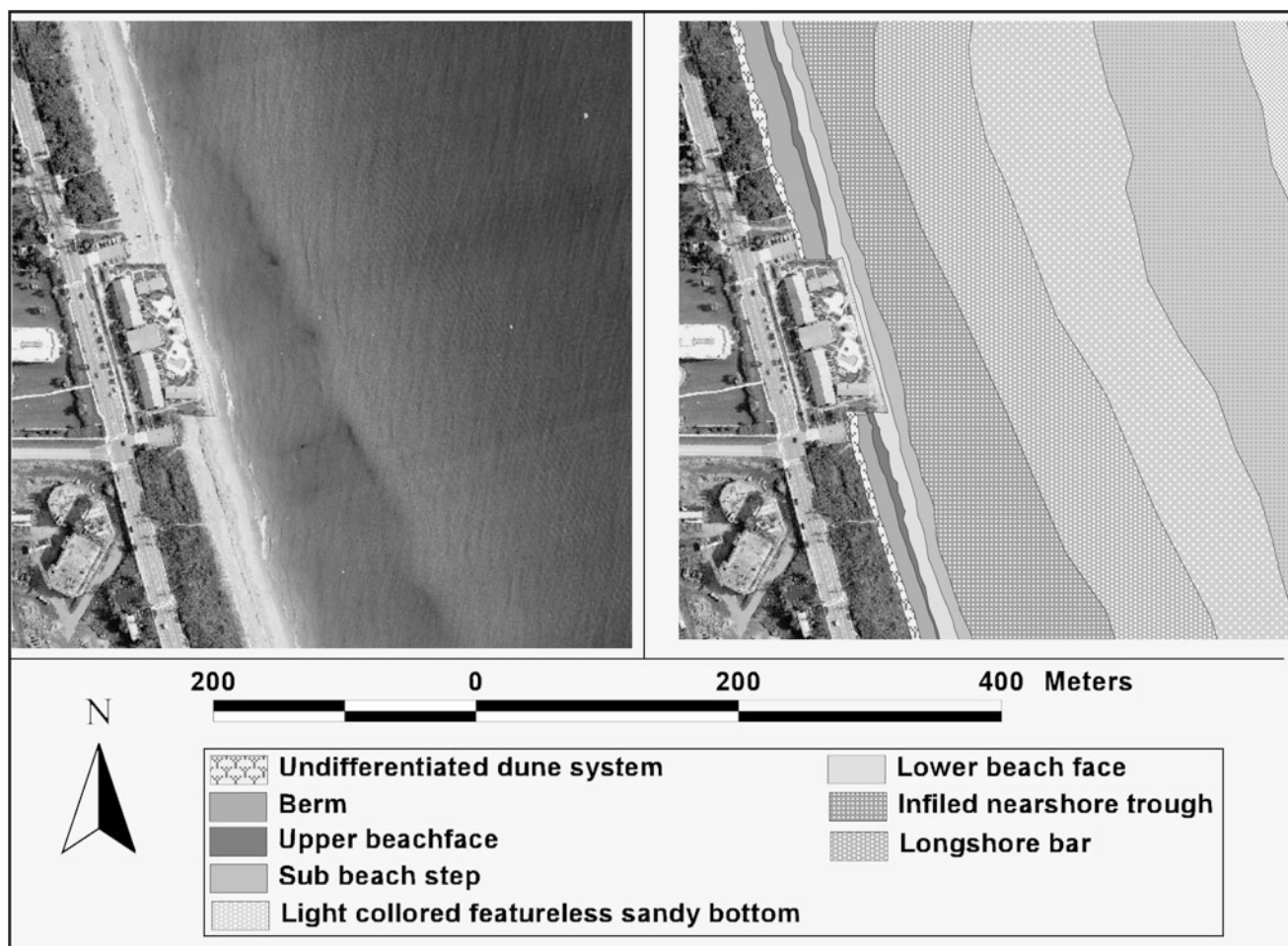
order of days (Larson and Kraus 1995). As Cowell et al. (1999) point out, the upper shoreface chases a dynamic equilibrium (Schumm 1991) as boundary conditions vary stochastically, without ever quite making it. This section of the southeast Florida coast is no exception as beaches and inshore-offshore sedimentary features are seen to change over the long run, from winter to summer wave climates, to occurrences of winter northeasters and summer tropical storms. This morphologically dynamic zone features a range of sedimentary morphodynamic subaerial and submarine

types that are identified from aerial photography in terms of coastal dunes, beaches, bar and trough systems, deltas, and sandy bottoms.

Coastal Dunes

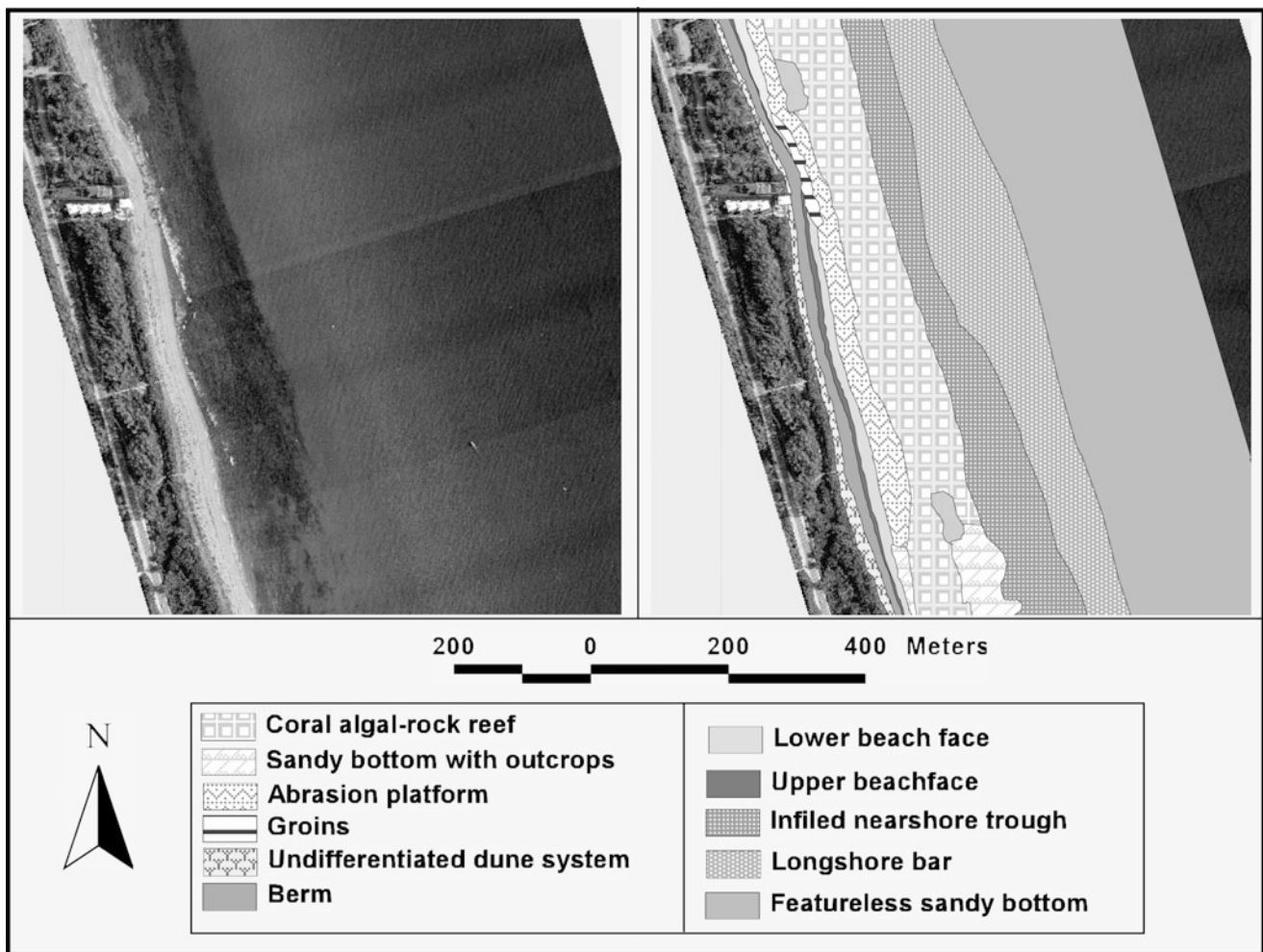
Dunes along this shore are generally compromised by construction for coastal infrastructure, recreational facilities, condominiums, or private estates. With the destruction of the primary dune system by development along this coast in the early part of the twentieth century, there is little space left for contemporary dune building processes to operate. Nevertheless, some areas contain small foredunes that

develop within a limited framework of environmental controls. Coastal dunes are found above the high water marks of sandy beaches and are formed by wind transport of predominantly loose, sand sized (2.0–0.05 mm) sediment of different origins, as described by Carter et al. (1990). Because of their limited developmental status and restricted geographic occurrence, no attempt is made here to differentiate types of coastal dunes. Dunes are mapped here, as shown in Figs. 6, 7, 8, 9, and 10, as undifferentiated dune system. Severely restricted in geographic occurrence, the dunes may be somewhat wider (up to 40–50 m in width) (Fig. 6, geomorphic map on right) to somewhat narrower (less than 20 m) (Figs. 9 and 10; geomorphic maps on right).



Nearshore Geomorphological Mapping, Fig. 9 Example of a simple sandy shore with a series of shore-parallel sedimentary morphoforms viz. coastal dunes, beach, bar and trough, and featureless sandy bottom near Jupiter, Florida. The mapped polygons (right) correspond to major features interpreted from the aerial photograph (left) showing a truncated dune system restricted by the coastal highway (A1A), a wider berm on the northern updrift side of the recreational complex built on the beach, a sediment starved beach on the southern downdrift side of the structure, and the presence of a well-developed lower beach face and sub-beach

step. The infiled nearshore trough flanks the seaward portion of the rectilinear beach face. Bars and troughs alternate in a seaward direction as depth increases. The Florida Reef Tract lies farther offshore and is not shown in this diagram. This coastal segment typifies a sediment-rich sequence of geomorphic forms and illustrates a gradation of bottom types that are commonly experienced along so-called soft coasts. The impression is, however, misplaced because the southeast coast of Florida is rock controlled



Nearshore Geomorphological Mapping, Fig. 10 Example of alongshore rock outcrops on the southeast coast of Florida near Tequesta, Palm Beach County. Hardbottom types are clearly evident in the aerial photograph (left) by their dark tones, coarse texture, and linear patterns. The interpreted map (right), showing salient geomorphological units, delineates the presence of coral-algal rock reef seaward of the well-developed abrasion platform that is about 100 m in width and 800 m in length. The beach is perched on top of the abrasion platform, which

extends landward under the beach. A bar and trough system flanks the seaward side of the coral-algal reef, part of the first reef in the Florida Reef Tract. Undifferentiated sandy bottom extends at least 500 m offshore to about 30 m depth, below which photointerpretation is not possible. When hardbottoms occur this close inshore, sands from the perched beaches are often stripped away by storm waves and currents during northeasters leaving an extensive rock surface exposed to the base of coastal dunes

Beach

As explained by Wright and Short (1984), there is considerable variability in concept, definition, and morphological properties of beaches. Although a common feature of many coasts, beaches can be deceptively complex systems that require careful attention to their biophysical characteristics and environments of formation. The term *beach* was originally used to designate the loose wave-worn shingle or pebbles found on English shores and is still used in this sense in some parts of England (Johnson 1919, p. 163). In common parlance, the term is used locally for the seashore or lakeshore area. More closely defined usage refers to the temporary accumulation of loose water-borne material that is in active

transport along, or deposited on, the shore between the limits of low and high water. From a geomorphological point of view, the beach is comprised by the unconsolidated material that covers a gently sloping zone, typically with a concave profile, and which extends landward from the low-water line to a place where there is a change in material or physiographic form (e.g., dune) or a line of permanent vegetation. Relying on these general concepts of beach as a morphodynamic unit, unconsolidated sandy material deposited along the shore zone between low and high water was mapped as beach (see units for berm and beachface in Figs. 6, 7, 8, 9, and 10). Beaches in this area contain admixtures of siliciclastics and carbonates of biogenic origin from shells and calcareous algae. In the southern part of the southeast coast of Florida, beach sands

are mostly carbonates and in the northern part they are mostly silicates. In the central part of this coastal segment, the beach sands are nearly equally differentiated into silica and carbonate components. All beach sands are shallowly underlain by the Anastasia Formation, which extends several kilometers inland (Finkl and Esteves 1997) and continues as a shelf deposit along the southeast coast of the United States to at least the latitude of central North Carolina. Morphodynamically, these beaches key out as mostly intermediate types in the beach classification system developed by Short (1999) but the unit is not differentiated for mapping purposes. Based on beach width and planform configuration, it is evident that beaches are eroding, accreting, or stable. An example of an eroding beach, that is, one with a retreating coastline, is shown in Fig. 9 where a condominium complex just out into the sea. This facility sits astride the dune system, berm, and upper beachface; its seaward-most portion terminates on the lower beachface. The berm is widened on the northern updrift side of the complex and narrowed on the southern downdrift side that is now starved of sediment because this structure acts as a partial littoral drift blocker. Figure 7 shows a stable or accreting beach with a berm nearly 90 m in width.

Rhythmic features and berms were interpreted from the aerial photographs and mapped as separate units. The *beachface*, that section of the beach normally exposed to the action of wave uprush (i.e., the foreshore of the beach), was subdivided into two units, the *upper and lower beachface*. This distinction was interpreted from the aerials on the basis of geomorphic expression of the cross-shore profile, the *beach scarp* providing a consistently recognizable division, except along beaches where municipalities deliberately destroy the scarp by scraping. Beach scarps of 1–2 m can be hazardous barriers that block access to the water and they are thus destroyed, especially when they form in newly renourished beaches.

Cusps

These ephemeral features occur intermittently throughout the study area and belong to the class of “typical beach cusps,” (vs. storm cusps) as defined by Dolan and Ferm (1968). Cusp embayments average about 6 m wide (measured between horns) and have wavelengths of about 13–15 m. Beach cusps occur sporadically along the beach, usually in groups of rhythmic topography about 300 m long. These crescentic, concave-seaward features at the shoreline occur at different locations at different times of the year; in general, however, cusped features seem to occur along coastal segments where reef tracts or hardgrounds outcrop in the inshore zone. Their morphometries (quasiregular spacing alongshore) differ according to localized oceanographic processes that are influenced by bathymetry and bottom type.

This rhythmic shoreline topography is associated with *intermediate beach systems*, as described by Hughes and Turner (1999). Beach cusps are not shown in any of the figures.

Berms

Berms are inter- to supratidal shore-parallel terraces or ridges that consist of a steeper seaward sloping beachface, topped by a shore-parallel crest. As explained by Hesp (1999), the term *berm* has come to be synonymous with the intertidal beach and backshore, as an asymmetric landform comprising a seaward swash slope (usually the steepest component; the “riser” in terrace morphological terminology), a crest, and a low, often linear, sometimes concave beach top slope (the “tread”). The term *berm* has often been taken to mean the nearly horizontal portion of the beach (i.e., the tread; see Davis 1982). These low, impermanent, nearly horizontal or landward sloping benches, which occur as a backshore terrace on a beach, are formed by material deposited by storm waves. Some narrow native beaches (beaches that have not been renourished) have no berm while some of the wider, accreting beaches immediately updrift from jettied inlets may have several. Well-developed, wide berms are usually associated with renourished beaches that may extend more than 100 m seaward of frontal dunes. Eroded beaches may have narrow berms and along some critically eroded coastal segments there is no beach. Figure 8 (geomorphic map, right), for example, shows a rock outcrop immediately seaward of a seawall with no intervening beach. The seaward portion of the rock outcrop is partly overlain by a low-tide terrace. Berm crests define the seaward boundary of the dry beach. Some alongshore segments were mapped as a berm/upper beachface complex unit because it was not possible to differentiate these two units on the aerial photographs. Small beach scarps separating the berm and upper beachface, which did not allow consistent identification on the aerials, compromised interpretation. On some beaches, the scarp is mechanically swept, which combines the berm and upper beachface into a continuous unit, to make a user-friendlier beach. Beach morphology is extremely variable and the units mapped here (berm, upper/lower beachface) relate to the time of photo acquisition.

Upper/Lower Beachface

The *beachface* is that section of the beach that is normally exposed to the action of wave uprush, that is, the foreshore of the beach. The beachface is defined as being in a state of dynamic equilibrium when the net sediment transport, averaged over several swash cycles, is zero (Hughes and

Turner 1999). Thus, if offshore sediment transport is dominant, the beachface profile becomes flatter; and conversely, if onshore transport prevails, the profile becomes steeper. Beachface profiles rarely exhibit a uniform and planar gradient; the typical beachface is concave seaward but the beachface slope may vary alongshore. In this study, the *upper beachface* was distinguished from the *lower beachface* because they are morphologically distinct and are easily separated on aerial photographs due to different tone and pattern (cf. Figs. 6, 7, 8, 9, and 10). The upper beachface extends from the landward limit of the lower beachface to the berm crest or foot of the beach scarp. The lower beachface is the seaward sloping portion of the foreshore where swash action begins to occur, that is, where waves break for the last time and swash moves up the beachface as a bore. The width and form of the beachface is significant because this is the feature that ultimately reflects or dissipates wave energy. In most cases, the lower beachface is less than 5 m wide.

Bar and Trough Systems

Study of cross-shore profiles has resulted in the notion of “winter” and “summer” profiles; winter storm waves removed sand from the berm forming a breakpoint bar, whereas calmer wave climates in summer induced landward migration of the bar with subsequent welding to the beachface to form a non-barred cross-shore profile (e.g., Fox and Davis 1973; Aagaard and Masselink 1999). Subsequent research suggested that beach response to wave climates was cyclic, rather than seasonal, and so the terms “barred” and “non-barred” appeared more appropriate (Greenwood and Davidson-Arnott 1979). The separation between barred and non-barred profiles is now reported to relate to the direction of cross-shore sediment transport. Offshore sediment transport results in erosion of the beach and formation of barred profiles, whereas onshore sediment transport causes beach accretion and non-barred profiles (Aagaard and Masselink 1999). Relationships between nearshore bar morphology and wave climates is complex and site specific, but periods of high or low wave energy impart distinctive bar types and sequences alongshore. Aagaard and Masselink (1999) report that in general, transverse bar morphology is found at the low energy end of the spectrum, whereas linear bar morphology characterizes high wave energy levels. Crescentic bars develop under intermediate wave energy conditions. All of these classical types of bar and trough systems are seen along the southeast Florida coast.

In addition to classical bar and trough forms, there are local or site-specific morphological variations and combinations that resulted in identification of the following main units: longshore bar, structurally controlled longshore bar, crescentic bar, crescentic attached bar, attached ridge with surge

channels, crescentic trough, transverse trough/rip channel, and infilled nearshore trough. Longshore bars are low, elongate sand ridges that are built up mainly by wave action and which occur some distance from the shoreline. Longshore bars generally extend parallel to the shoreline and are submerged at least by high tides and are separated from the beach by an intervening trough (see Figs. 6, 7, 8, 9, and 10). Several terms have been applied to these kinds of features (e.g., ball, offshore bar, submarine bar, barrier bar) as described here in their simplest situational occurrence. There is a wide range of morphological variation where bars and associated troughs occur as single linked units (e.g., Fig. 10) or in multiple associations in rectilinear to strongly curved (wavy) planforms (e.g., Figs. 6, 7, 8, 9, and 10). Bars and troughs may be continuous along the shore for many kilometers or they may occur as disjunctions continuously or sporadically along the shore (cf. Figs. 6 and 10), parallel to shore or inclined at an angle (e.g., Fig. 6). When longshore bars have migrated shoreward and become welded to the beachface as relatively high ridges of sand, they are termed welded bars or, as referred to here, as *attached ridges* (Fig. 7).

Longshore Bar

Along the southeast Florida coast, longshore bars occur in the inshore zone as single bars or as part of multiple bar and trough systems (Figs. 6, 7, and 8). The bars are sediment ridges of running sand that generally parallel the shore. Some bars are rather narrow, for example, about 15 m wide, and continue without break for distances longer than 1.5 km. Occasionally, one end of transverse bars are welded to the shore either singly or in multiple sets (Fig. 6).

Crescentic Bars and Crenulated Troughs

Crescentic bars are rhythmic shore features that tend to have hyperbolic wavelengths on the order of 115–160 m and are usually located about 12–60 m from the shoreline (e.g., Fig. 6). Crescentic bars are sometimes welded to the shore (Fig. 6) but welded or not, they are associated with troughs that are either infilled nearshore or crenulated. The term *crenulated trough*, as applied here, describes certain inshore coastal segments where sedimentary troughs occur in the inshore zone adjacent to crescentic bars. The adjacent border of the crescentic bar forms their borders so that a convex seaward portion of a crescentic bar forms the concave seaward crenulated trough. Crenulated troughs may form on either the landward or seaward side of crescentic bars, and they may occur singularly (inside one cycle of the crescentic bar) or they may connect to form a crenulated trough that stretches 4 km or more alongshore.

Infilled Nearshore Trough

As defined here, infilled nearshore troughs are elongate sedimentary depressions that are associated with bars. Because the troughs tend to become partially infilled under subsequently calmer conditions than during their formation during storms, they are referred to as being infilled. These depressional features of the inshore zone range in width up to 150 m cross-shore and may extend for several kilometers alongshore (Figs. 8, 9, and 10). As reported by Lippman and Holman (1990), longshore-bar-trough states are variable and the features may be highly mobile under certain wave conditions, whereas Wright et al. (1985) concluded that straight or rhythmic longshore bar and trough states could remain relatively stable under certain conditions. The bar and trough systems described here seem to remain positionally stable from year to year, based on observations of interannual photography. The presence of reefs and rock outcrops on the seafloor in the study area apparently complicates present comprehension of bar-trough positional stability.

Shoreface Trough

These troughs, lying seaward of the lower beachface, contain variable water depths but all are several meters deeper compared with the longshore bars or hardground features to which they are adjacent. Shoreface troughs may extend uninterrupted for long distances, for example, up to several kilometers alongshore. Occupying areas of lower elevation compared with surrounding morphologic features, usually coral-algal rock reefs or structurally controlled sandflats, these troughs are commonly floored by hardbottom with thin sand veneers. Structural features of carbonate hardbottom are often evident in the aerial photographs, verifying the thin cover of overlying sand. The troughs more or less parallel landward variations in the Florida Reef Tract or follow inshore lineations in the Anastasia Formation.

Sandy Bottom

Sandy sediments are a dominant feature of the seafloor on the inner continental shelf, occupying about 30–40% of the mapped areas (e.g., Brown 1998; Warner 1999), the remainder being coral-algal reef or exposure of bedrock as hardgrounds. The sandy deposits are mostly sheets overlying the karstified bedrock as basin infills. Corollary data from seismic survey shows that the sand lying between reefs in the Florida Reef Tract are shallow basinal-type deposits with stringers or lenses of fine grained (silt plus clay) sediments and occasional coarse rubble mounds in the vicinity of paleoinlets (now submerged reef gaps) (e.g., see Finkl and Bruun 1998; Khalil 1999). Sedimentary infills here average several

tens of meters in thickness, but present a generally monotonous flat topography at the surface between reefs. Because of the flat, rippled surface patterns seen on the aerial photographs, these features are referred to as sandflats or simply as sandy bottom. The sandy deposits are more or less of uniform grain size at the surface and thus appear similar and are easily recognized in the interpretive process. Some areas, however, are partially covered by anchored or bottom-drifting algal mats or less commonly with sea grass beds giving a dark-toned appearance in the imagery. Because these areas are so distinctive in appearance and thus mutually exclusive, they were separated as different mapping units and referred to simply as light- and dark-toned running sands. Areas of sand sheets appear similar to the inter-reefal sand flats but instead of occurring offshore they tend to occur more inshore and shallowly overlie karst limestone. The underlying limestone is evident in the imagery by its reticulate structural patterns that poke through the thin sedimentary cover. These sandy bottoms thus present complicated image patterns that are combinations of sedimentary cover over outcrops of rough bedrock surfaces. Although wide ranges of variable patterns are associated with this unit, they are grouped into one category of occurrence for mapping purposes.

Light- and Dark-Toned Running Sand

Unconsolidated running sands without vegetation or algal mats appear as smooth, fine textured, bright patterns on the aerial photographs. The “light-toned” photograph interpretation descriptor is used to differentiate this unit from the darker-toned seabed sands. These clean running sands occur mainly in the offshore zone (e.g., Figs. 8 and 9), as rather wide (about 60–120 m in width) bars that parallel the shore or as sand sheets. Collateral data (e.g., side-scan sonar) indicates that these sandy bottoms may extend to several (6–7) meters in thickness and overlie carbonate substrates of the Florida Reef Tract. Dark-toned running sands are differentiated on the basis of the areal distribution and density of benthic algae or biological material that is integrated within the surface-most sediments. These dark-toned running sand sheets occupy large areas of seafloor and in contradistinction to the parabathic light colored bars and sheets, they display many different irregular shapes that are determined by the presence of organic matter at the sediment surface. These sands also infill depressions between reef tracts and occur as sheets of variable thinness over hardgrounds.

Structurally Controlled Sandflats

As defined here, these submarine sandy areas of flat lying sediments are bordered on both landward and seaward

margins by hardground ledges. Figure 6, for example, shows marine sediments that have infilled solutional depressions in rock reefs. These sandy areas are laterally and vertically controlled by preexisting or antecedent topography that was previously subaerially exposed and then drowned by the Flandrian Transgression. The bounding ledges, which are clearly evident in aerial photographs, generally rise about 1–2-m above the seafloor to interiorly trap sediment. In some areas, the karst structures of the underling limestone show through the thin sedimentary veneer or are propagated upwards through the sedimentary cover to show the limestone's reticulate structural pattern. The structurally controlled sandflats tend to be irregular in overall shape, but their borders create zigzag or saw-tooth patterns. These sandy areas rarely contain algal mats and the sands are clean.

Tidal Delta

Shoals associated with tidal inlets occur as flood-tidal deltas and ebb-tidal deltas along this microtidal (less than 2 m) coast with semidiurnal tides. Tidal prism normally has a large influence on the size and shape of natural inlets, but inlets in the study area are cut through bedrock of the Anastasia Formation making them positionally stable (Finkl 1993). Because the major ports have stabilized deep navigational entrances with cuts 15–18 m deep and jetties, which in turn results in practically no bypassing of sediments, only smaller inlets have well-developed deltas of the classical form in shallow water. These wave-dominated ebb-tidal deltas (Davis 1997), always completely submerged along this coast, are sometimes deformed southwards due to structural impacts associated with the downdrift curvature of jetties and the net littoral drift from north to south. Bypass bars are often associated with the ebb-tidal deltas.

End Products, Applications, and Insight

Detailed coastal mapping, based on interpretation of aerial photographs, can result in the preparation of special purpose morphologic maps along specified coastline segments. Mapping areas may be defined on the basis of coastal physiography or administrative units such as counties. Interpretation of georectified digital aerial photography has the advantage of being incorporated into spatial analysis programs such as ArcView®. Maps prepared in this manner can be compared with preexisting coverages for research and visual display. Mapping of coastal morphology provides new insight into the topologies of coasts that provides better understanding of process-form relationships. Classification of coastal morphologies elucidates associations of natural features and differentiates groups of related features. The delineation of

coastal morphologies provides a basis to model coastal landform development and provides baseline data for subsequent change detection analysis.

Detailed coastal morphological maps (e.g., Brown 1998; Khalil 1999; Warner 1999) are useful because they provide a record of coastal biophysical conditions that in turn may find application, for example, in: (1) searches for beach-quality sand that is required for beach renourishment, (2) studies of coastline stability where offshore morphology may determine onshore anchor points or erosional hot spots (localized zones of accelerated beach erosion), (3) in the development of computer models of coastal processes, (4) assessment of the structure and integrity of biological habitats, (5) background geological and geotechnical information for engineering works, (6) coastal hazards research related to surge flooding, overwash, sediment erosion and deposition, location of rip currents, and (7) investigations of coastal geomorphology, evolution of coastal configurations, and coastal/marine landform classification. The methodology is applicable to clear Class II waters where there is low turbidity in the water column during parts of the year when aerial photography can be acquired. Many rocky, sandy beach, and coral reef coasts are thus amenable to study using this technique. Muddy coasts and highly turbid waters are excluded from interpretation of bottom features using aerial photography.

These kinds of morphological studies emphasize the importance of underlying geomorphology to coastline configuration (in planform) and submarine beach profiles. With decreasing natural availability of littoral drift sediment resulting from dredged inlet construction, the sand layer overlying the Anastasia Formation will continue to decrease in thickness. Consequently, an increased importance of underlying geology in describing profile variability and coastline configuration should be expected. It also follows that a greater portion of the beachface will be comprised of rock outcrops and a greater amount of littoral sediments may be permanently lost to the offshore. This positive feedback mechanism persists at the expense of artificial beach renourishment projects (Brown 1998). As discussed by Finkl (1993), the limestone base will naturally inhibit shore erosion at points where exposed as submerged hardground in the nearshore zone (Fig. 11) and thus control the alteration of beach profiles, shoreline configuration, as well as nearshore processes. Submarine headlands are thus seen as representing paleo-topographic highs in front of modern shores and modify the incoming wave climate (Pilkey et al. 1993). Morphological mapping here clearly shows that rock-control of a sediment-starved beachface is a reality. As shown in Fig. 11, perched beaches may be eroded away during storms and dunes threatened by waves but the rock platforms protect the shore. Sand eventually becomes reestablished on top of the bedrock and a new beach is again perched on top of

Nearshore Geomorphological Mapping, Fig. 11

Exposure of the Pleistocene Anastasia Formation in Boca Raton, Florida. This marine abrasion platform has truncated the coquina facies of the limestone bedrock base that occur along the Atlantic shore. The seaward margin of the platform along this rocky shore, which occurs as a pronounced ledge up to a meter or so in height, serves as a natural seawall. Note waves breaking offshore in the surf zone, tripped by hardgrounds, and the wave in the photo center that has crashed against the vertical wall of the ledge. There is a duality to this rock platform in that it helps to protect the shore from wave attack but at the same time inhibits onshore sediment, except under high-energy conditions. (Photo: Lindino Benedet)



the platform. These snapshots confirm on the ground what is interpreted from the aerial photographs and verifies the spatio-temporal importance of coastal materials that influence shoreline morphology.

It is the interpretation of aerial photography that makes this morphologic data crucial for coastal morphodynamic studies. Given more than one set of digital aerial photographs, multiple GIS coverages of the delineations of coastal geomorphology can be used in numerous overlay operations in order to gain insights into the planform coastal dynamics (e.g., Finkl and DaPrato 1993; Brown 1998). Essential to this topic is the development and implementation of methodologies by which to continually map and assess the coastal data acquired by present and future reconnaissance efforts.

Conclusion

In this example of detailed coastal mapping and classification, several important points emerge to illustrate the advantages of making morphological inventories based on interpretation of aerial photographs. That mapping based on georectified digital imagery is faster and more cost effective than conventional survey methods is hardly surprising. There are, however, many surprising elements to remotely sensed coastal surveys such as this one. Facts derived from visual and digital records are more reliable and informative than general impressions or suppositions that are often used in coastal planning, management, and engineering schemes. The southeast Florida shore is commonly referred to as a sandy beach coast with abundant sedimentary deposits offshore. Results of mapping here showed that beaches are shallowly perched

over a limestone base, beach deposits are relatively thin at <2 m in thickness (visible after large storms when beach sands are washed from buried shore platforms), there are many indicators of cross-shore processes (e.g., fluted rock reefs), there is a large expanse of hard bottom types in inshore and offshore zones, and that extremely narrow subaerial beaches occur inshore of reef gaps. Overall observations, based on analysis of the resulting morphological maps, indicate that this coastline morphology is lithologically controlled, the seabed is composed of drowned karst topography overlain locally by unconsolidated sedimentary bodies, and that coastal barriers such as the Florida Reef Tract modify coastline configuration.

Additional advantages that accrue from this kind of detailed mapping of coastal morphological features include the ability to define morphodynamic zones. Recognition of coastal process zones provides a rational basis for coastal management because certain topographic forms are morphodynamically significant, allowing interpretation of dominant processes. Although the technique is limited to clear Class II waters, significant stretches of the world's coastline warrant application of aerial photographic interpretation in conjunction with other methods of investigation.

Cross-References

- [Bars](#)
- [Beach Features](#)
- [Beach Processes](#)
- [Coasts, Coastlines, Shores, and Shorelines](#)
- [Coral Reef Coasts](#)

- Coral Reefs
- Photogrammetry
- Remote Sensing of Coastal Environments
- Rhythmic Patterns
- Shoreface
- Shoreline and Coastal Terrain Mapping
- Surf Zone Processes

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Nearshore Sediment Transport Measurement

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Definitions and Key Parameters

Sediment transport in coastal environments is driven by the combined motions of waves and currents. Due to the orbital nature of wave motion, methods developed for measuring sediment transport in unidirectional flow field may not be directly applicable in measuring coastal sediment transport. Furthermore, wave breaking and associated intense turbulence play a significant role in nearshore sediment transport. In the following, the term nearshore is used in a general sense, representing various coastal environments with shallow water.

Generally, sediment transport rate is calculated as

$$Q = \frac{1}{t} \int_0^t \int_0^{x_n} \int_0^h C_s(x,z,t) U_s(x,z,t) dz dx dt \quad (1)$$

where x and z = horizontal and vertical coordinates, t = time, x_n = width of a certain nearshore zone over which the transport is computed, U_s = particle velocity, and C_s = sediment concentration. The origin of z (i.e., $z = 0$) represents a level at which the sediment is not moving. It is often below the seabed surface because a thin layer of sediment may be moving at the seabed. Therefore, h equals to the water depth from the immobile sediment to the water surface. The resultant Q is the total transport rate integrated over h across a horizontal scale x_n and averaged over a time period of t . Sediment concentration C_s is a scalar quantity. Under most circumstances, C_s is greater near the bottom and decreases logarithmically (or exponentially) upward. Therefore, it is important for sediment concentration measurement to extend as close to the bottom as possible. In order to use Eq. 1 to calculate sediment transport rate, simultaneous

current measurement is also important. Sediment transport rate, the product of a scalar C_s and a vector U_s , is a vector in the same direction as U_s .

In coastal zones, especially in environments controlled strongly by wave motions, U_s and C_s vary rapidly with time and space. Measuring U_s and C_s simultaneously with adequate temporal and spatial resolutions is difficult. The product of time-averaged U_s and C_s yields sediment transport rate

$$\overrightarrow{F_s(x,z)} = \overrightarrow{U_s(x,z)} \times \overline{C_s(x,z)} \quad (2)$$

Under many circumstances, directly measuring the sediment transport rate is technically easier than measuring the U_s and C_s separately. However, precise measurements of U_s and C_s will not only yield transport rate but also shed light on transport processes. In coastal waters, the sediment concentration C_s and particle velocity U_s are influenced by many hydrodynamic and sedimentary factors. It is beyond the scope of this article to examine these factors. A comprehensive summary was provided by Van Rijn (1993).

Sediment transport is often described in three modes: bed load, suspended load, and wash load. As often as these terms are used, they are not as clearly defined, especially from a measurement perspective. The wash load describes the portion of fine particles that are transported by the water but are not represented in the bed. Wash load is typically neglected or incorporated into suspended load during transport measurements. Bed load is typically defined as the part of the total load that is in frequent contact with the bed during transport. It primarily includes grains that roll, slide, or jump along the bed. The suspended load is the part of the total load that moves without frequent contact with the bed controlled by the fluid turbulence.

The concepts of bed load and suspended load provide a representation of the modes of particle movement. In practice, bed load and suspended load as defined above cannot be distinguished during field or laboratory measurements. From a measurement perspective, bed load is sometimes defined as the part of the total load that travels below a certain level, which may be defined differently by different researchers, often influenced by the measurement device. Therefore, bed load and suspended load measured during different studies may not be directly comparable due to different definitions. Caution should be exercised in using these terms in nearshore sediment transport measurement, and clear definition should be provided.

Fundamentals of Nearshore Sediment Transport Measurement

One of the first and important decisions in planning a nearshore sediment transport measurement project is to determine

the appropriate temporal and spatial scales. "Coastal zone" is a general term describing many different types of coastal environments, including tidal inlets, surf zones, estuaries, tidal deltas, etc. Sediment transport is dominated by different processes in these different environments. The spatial scale should be adequate to describe the characteristics of the specific coastal environment. Coastal sediment transport often demonstrates apparent periodicity controlled by the cycles of water motion and sediment supply, such as wave cycles, tidal cycles, seasonal wave conditions, seasonal cycles of riverine sediment supplies, etc. The temporal scale should be able to resolve the periodicity that is of interest to the specific study.

The appropriate temporal and spatial scales are also influenced by the objectives of the sediment transport measurement, and different measurement methods are suitable for different scales. Generally speaking, finer temporal and spatial scales may provide more accurate measurement at the target location during the study period. However, fine scale studies tend to be short term with limited regional coverage and may not accurately capture regional characteristics and yield a representative regional long-term transport rate. Therefore, fine scales tend to be used in studies where the main objective is to quantify transport physics. On the contrary, coarser temporal and spatial scales focus on providing regional and long-term transport rates, although local details and short-term variations tend to be neglected. For studies with the objective of obtaining long-term regional transport rate, coarse scales may be more suitable than fine scales. Determining proper temporal and spatial scales comprises an important portion of the nearshore sediment transport measurement. Coastal sediment transport demonstrates various periodicities; it is important to incorporate certain periodicities.

Both current and sediment concentration profiles typically demonstrate logarithmic trends with rapid changes in the vertical direction. The errors in determining elevations may be exaggerated during vertical interpolation and calculation of sediment flux. The logarithmic shape of the profile should be taken into consideration during the vertical interpolation and sum and calculation of depth-averaged values.

Since sediment transport rate is a vector, determining its direction is as important as determining its magnitude. The direction of the sediment transport is equal to the direction of particle velocity or current velocity in most cases. The majority of current sensors measure the velocity along two perpendicular axes, and the vector sum yields the magnitude and direction of the total current. A proper coordinate system provides many conveniences in describing the transport direction. One of the most commonly used coordinate systems aligns one axis parallel to the shoreline and one perpendicular. However, this longshore and cross shore system may not be the most efficient for coasts with complicated features such as

inlets and river mouths, where careful consideration of the dominant regional features may be necessary.

Methods for Measuring Nearshore Sediment Transport

Based on its definition (Eq. 1), sediment transport rate can be obtained by simultaneously measuring sediment concentration and particle velocity. However, in dynamic coastal zones, simultaneous measurements of sediment concentration and velocity are sometimes not possible due to technical limitations, and as a result, alternative indirect methods are often used. In the following, nearshore sediment transport measurements are discussed in two categories: direct and indirect measurement. Direct measurements obtain sediment transport rate or via the product of sediment concentration and particle velocity. Indirect measurements involve measurements of other quantities, such as morphological changes, from which sediment flux can be deduced.

Direct Measurement of Sediment Transport Rate

Numerous methods are available for measuring sediment concentration and particle (or flow) velocity. Often, the difficulty is measuring concentration and velocity simultaneously at the same location and near the bottom. A straightforward measurement of sediment concentration is to determine the weight of sediment particles in a unit volume of water. A commonly used method consists of an array of sampling bottles with a trigger device to open and close the bottles (Kana 1979). A considerable area can be covered by each sampling event due to the low cost. Temporal coverage of this kind of measurements is limited because of the manual sampling. Also, sampling interval is generally short limited by the typically small amount (no more than several liters) of water sample. In a wave-dominated environment, this method may not be able to provide a representative average over an entire wave cycle due to its short sampling. The short sampling interval also makes the simultaneous velocity measurement difficult. In less dynamic environments with limited wave influence, such as semi-enclosed estuaries, where sediment concentration may not change quickly, the instantaneous sampling may provide reasonable representation of average value. Direct deployment of sampling bottles is often replaced by a pump-suction device. By regulating the rate of intake, the sampling duration can be increased to provide measurement of average value over several wave cycles (Nielsen 1984).

A variety of turbidity sensors are available for measuring sediment concentration. The optical backscatter (OBS) is commercially available and is probably the most commonly used. OBS sensors can be deployed and synchronized with current sensors to provide measurement of sediment transport rate. The automated and self-recording concentration sensors

are capable of sampling at high frequency to resolve sediment suspension events, such as those driven by wave motions (Beach and Sternberg 1992; Osborne and Greenwood 1993). A limitation of sensors like OBS is that the output is significantly influenced by sediment properties, and therefore, it requires calibration using in situ sediment. Bottom sediment from the study site is often used for the calibration. However, it is not uncommon that bottom sediment is considerably different from sediment in suspension (Wang et al. 1998a). Therefore, calibration using bottom sediment may introduce uncertainties in computing the OBS concentration. Comparison with direct concentration measurement is always desirable. The self-recording sampling allows long-term measurements; however, biological fouling, especially during the summertime, necessitates periodic cleaning of the sensors. The spatial coverage of the turbidity sensors is often limited by their relatively high cost, since at least three are required for the measurement of concentration profile.

In order to obtain sediment transport rate, currents must be measured simultaneously with concentration. Measuring current velocity can be more complicated than measuring sediment concentration since velocity is a vector. A commonly used current sensor is the electromagnetic current meter (EMCM) which measures currents along two perpendicular axes. Acoustic Doppler velocimeter (ADV), which measures 3-D particle velocity directly, is another commonly used sensor. The acoustic Doppler current profiler (ADCP), which is capable of measuring current profiles throughout the water column, is becoming a standard equipment used in tidal channels and in deeper water.

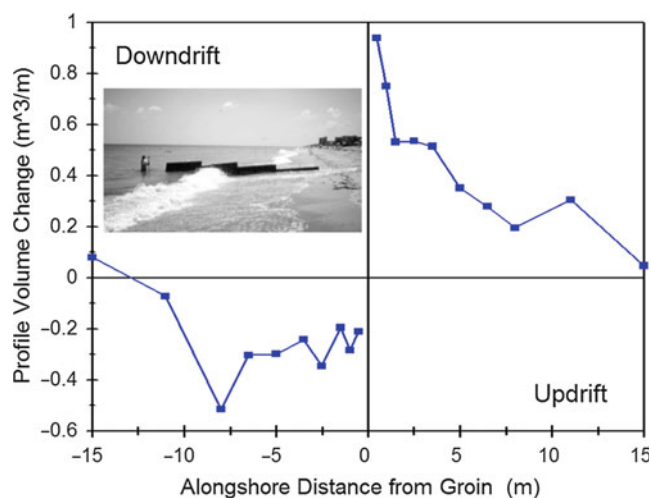
Sediment transport rate can be obtained using traps without measuring sediment concentration and flow velocity separately. Generally, there are two types of sediment traps, categorized by their functionality. Type 1 functions by reducing the intensity of the hydrodynamics to such an extent that the sediments that are moving with the fluid settle into the trap. Type 2 functions by providing a selective obstruction to sediment movement, for example, via the use of sieve cloth. The selective obstruction blocks the movement of the sediments and retains them in the trap while allowing the fluid to pass through. Because it allows the passage of fluid, type 2 traps generally impose less disturbance to hydrodynamic conditions than type 1. Disturbance to hydrodynamic conditions may have a significant influence on the efficiency of sediment traps, as well as on the transport processes.

Due to the complexity of coastal hydrodynamics, no universal trap design functions well in all coastal environments. In the following, sediment traps used to measure longshore sediment transport in the surf zone are discussed as examples. An example of type 1 trap is described in Wang and Kraus (1999). A temporary shore-perpendicular structure was deployed to block longshore sediment transport, causing the sediment to be impounded updrift of the structure.

The sediment transport rate was obtained by quantifying the volume of sediment accumulated at the updrift during the period of experiment (Fig. 1), which ranged from 2 to 6 h for the measurements conducted by Wang and Kraus (1999). Since the structure blocked longshore sediment supply, downdrift erosion occurred. Due to mass conservation, the updrift accumulation should equal to the downdrift erosion. The achievement of mass balance during the measurement provides a check for data quality, which is valuable for non-

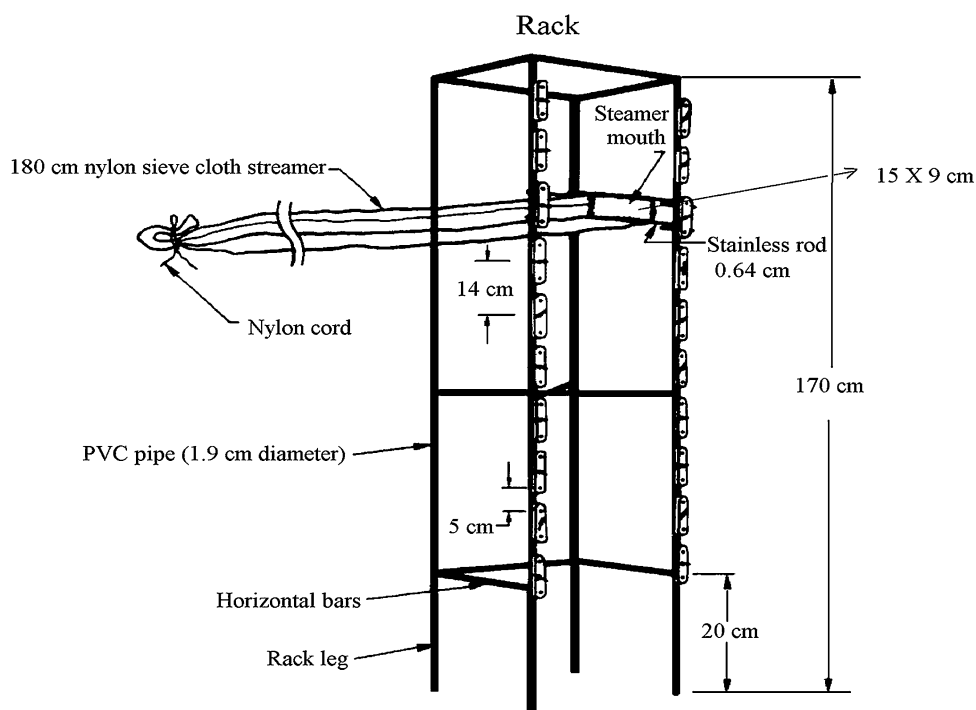
repeatable and non-controllable field measurement. Some permanent shore protection structures may serve as traps. Quantification of beach changes in the vicinity of these structures may provide reasonable estimates of sediment transport rate averaged over several months to a year (Dean 1989).

The streamer sediment trap developed by Kraus (1987) and Wang et al. (1998b) provides an example of the type 2 trap (Fig. 2). The streamer traps are made from sieve cloth of a mesh size dependent upon the size of the sediment in the study area. The traps not only allow water to flow through the sieve cloth but also move with the cross shore wave motion and, therefore, impose minimal disturbance to the hydrodynamic conditions. The sediment transport rate is measured by the weight of sediment trapped in the streamer bag over the period of the experiment. The traps are operated manually, and the sampling interval ranged from 3 to 15 min, representing a value averaged over 60–200 waves. The longshore component of the sediment flux can be measured by orienting the trap opening in the longshore direction. The streamer traps are mounted on a vertical rack, providing measurement of the vertical distribution of sediment flux. By deploying the trap array at different locations in the surf zone, cross shore distribution of longshore sediment transport can be measured (Wang 1998). Both the short-term impoundment and streamer sediment traps are fine-scale methods yielding nearly instantaneous measurements of longshore sediment transport and are therefore more suitable for studying transport physics than obtaining regional long-term rates.



Nearshore Sediment Transport Measurement, Fig. 1 Sand accumulation at the updrift of a short-term impoundment experiment (photo) and erosion at the downdrift

Nearshore Sediment Transport Measurement, Fig. 2 Design of the streamer sediment traps and the frame (Modified from Wang et al. 1998b). The sizes can be modified based on the measurement conditions



Indirect Measurement of Sediment Transport

Generally, indirect methods can be classified into two categories. Method 1 provides an estimate of sediment transport rate via tracing the movement of a small amount of specially marked sediment particles (sediment tracers), from which the movement of all the grains is deduced. Method 2 estimates the rate of sediment transport by quantifying morphological changes resulting from net sediment transport.

Although an indirect method, sediment tracers were the primary method for measuring longshore sediment transport for over four decades (Ingle 1966). The principle of tracer method can be simplified as that a relatively thin layer of sand is assumed to move along shore at an advection speed. The total longshore transport rate in the surf zone is calculated as

$$Q = x_b Z \frac{P_2 - P_1}{t} \quad (3)$$

where Z = thickness of the moving layer, x_b = surf zone width, P_1 and P_2 = locations of the tracer mass at the beginning and end of the experiment, respectively, and t = duration of the measurement. Accurately estimating Z through the vertical distribution of the tracer can be difficult (Kraus 1985). Determination of location P_1 is straightforward, controlled by the initial tracer injection. Location P_2 can only be determined statistically and is influenced by many factors including the sampling scheme (White and Inman 1989). The theoretical foundation and practical difficulties of the tracer method have been critically examined by Galvin (1987) and Madsen (1987).

Transport rate over a longer term can be estimated from morphological changes. The fundamental assumption is that morphological changes are caused by sediment accumulation or erosion, which are related to the rate of net sediment transport. For example, longshore transport rate can be estimated from the alongshore migration of barrier spits. From the volume change of ebb-tidal deltas, sediment transport in the vicinity of tidal inlets can be estimated (Davis and Gibeau 1990). These indirect methods are typically of large temporal and spatial scales and tend to provide estimates of regional, long-term transport rate. Morphological changes may be dominated by energetic events. Therefore, linking the measured transport rate with hydrodynamic condition can be difficult.

Summary

Measuring nearshore sediment transport rate is a difficult task and no single method is universally applicable. A sound understanding of regional hydrodynamics and sediment transport processes is important in adopting and executing a suitable method for sediment transport measurement. The first

step in planning a transport measurement project is to determine the proper temporal and spatial scales, which are partially controlled by the objective of the study. Some methods, such as sediment tracing or manual sediment trapping, are suitable for short-term measurement in a small region, while automated monitoring with self-recording instrumentation is more suitable for longer-term study in a small, focused area. The few methods that are suitable for long-term, regional study are predominantly indirect techniques involving the examination of regional morphological changes. It is worth emphasizing that sediment transport rate is a vector and is described by not only a magnitude but also a direction. Sometimes, the directionality is simplified by studying the transport component in a predetermined direction, such as transport in the longshore or cross shore direction, respectively.

Cross-References

- [Beach Erosion](#)
- [Beach Processes](#)
- [Cross-Shore Sediment Transport](#)
- [Erosion Processes](#)
- [Gross Transport](#)
- [Longshore Sediment Transport](#)
- [Monitoring Coastal Geomorphology](#)
- [Nearshore Wave Measurement](#)
- [Sediment Budget](#)
- [Shoreline Response to Littoral Drift Barriers](#)

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Nearshore Wave Measurement

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Definitions and Key Parameters

Generally, water waves are described by two length parameters, wave height and wavelength, and one temporal parameter, wave period. Wave height is the vertical distance between the wave crest and trough. Wavelength is the horizontal distance between two successive wave crests or wave troughs. Wave period is the time needed for two successive crests or troughs to pass a spatial reference point. Direction of wave propagation is also an important parameter and critical when computing wave-induced current and sediment transport (Wang et al. 1998, 2002). Scientific convention describes the wave direction as “direction to which it propagates” measured clockwise from the x -axis. In practice, however, wave direction is often reported as “direction from which the waves propagate,” similar to the description of wind direction. It is necessary to specify the wave direction convention to avoid potential confusion. Depth over which the waves propagate is also important and is necessary in linking wave height and length to other parameters such as wave-induced

water particle velocities and accelerations. Based on linear wave theory, wavelength and wave period are related through the dispersion relation as

$$L = \frac{g}{2\pi} T^2 \tanh \frac{2\pi h}{L} \quad (1)$$

where L = wavelength, g = gravitational acceleration, T = wave period, and h = water depth. Although nonlinear properties become more significant in shallow water than in deep water, they have rarely been considered during general wave measurements. Small amplitude (also known as Airy or Linear) wave theory has been used widely in coastal science and engineering research (Dean and Dalrymple 1991).

A commonly used 2-D description of a small amplitude progressive wave is

$$\Phi(x, z, t) = -\frac{H}{2} \frac{gT}{2\pi} \frac{\cosh k(h+z)}{\cosh kh} \sin \left(kx - \frac{2\pi}{T} t \right) \quad (2)$$

where Φ = velocity potential, k = wave number defined as $2\pi/L$, x and z = horizontal and vertical coordinates, respectively, and t = time. The horizontal (u) and vertical (w) water particle velocities can be determined from the velocity potential as

$$u = -\frac{\partial \Phi}{\partial x} \quad (3)$$

$$w = -\frac{\partial \Phi}{\partial z} \quad (4)$$

Describing waves in the real world using one sinusoid, such as Eq. 2, is an oversimplification. A more realistic description would be a superposition of a large number of sinusoids. In practice, the above wave parameters are described statistically. The purpose of this entry is to provide a summary of field methods used to measure key wave parameters including statistical wave height, wave period, and direction in nearshore regions. It is acknowledged here that wave measurement, and particularly its direction, is one of the more difficult tasks in observational coastal engineering and science.

Fundamentals of Wave Measurement

For the convenience of discussion, wave measurement is described in the following two categories: (1) measuring the sinusoidal parameters including wave height, period, and orbital velocities, etc. and (2) measuring the direction of wave propagation. Measuring wave direction is more costly and instrumentation-intensive and is more complicated with regard to data analysis, than measuring nondirectional waves.

Nondirectional Wave Measurement

Generally, three approaches have been used to measure non-directional waves and include measuring: (1) the vertical acceleration of water particles; (2) orbital velocities of water particles; and (3) fluctuations of water surface position. Approach (1) involves the use of wave-rider buoys, which are most commonly used in deep and intermediate depth wave measurement. Buoy use in coastal waters is limited by its relatively low sampling frequency and inability to measure water depth. Approach (2) requires complicated data processing procedures to obtain wave height. Approach (3) is most commonly used in nearshore wave measurements and provides the most direct measurement of wave height and period. Most applications use wave height and period to describe waves, while orbital velocities are depth dependent and are usually calculated based on a certain wave theory, for example, using Eqs. 3 and 4 if linear wave theory is applied.

Directional Wave Measurements

In order to obtain wave direction, it is necessary to measure the nondirectional parameters simultaneously with one of the following: (1) the direction of water-surface slope or (2) the direction of the water particle velocity. Measuring the water surface slope is accomplished using wave-rider buoys by recording the pitch and roll of the buoy. The water surface slope can also be measured by deploying a specially arranged array of at least three water-level sensors. Measuring the velocity vector is more frequently used to obtain directional nearshore waves by combining a velocity sensor with the water-level sensors.

Wave direction can also be obtained through remote sensing techniques, such as radar and satellite imagery. A major advantage of this approach is that it provides a 2-D view, instead of a point measurement. Remote sensing methods remain to be innovative and promising for future wave measurements.

Wave Sensors

For the convenience of discussion, wave sensors are divided into two categories, including: (1) surface sensors and (2) subsurface sensors. There are two types of surface sensors, wave buoys, moving with the water surface and recording the vertical acceleration of the wave motion, and surface piercing wave staffs which are generally attached to a fixed structure and measures the water-level fluctuations. The most commonly used subsurface sensors are the pressure transducers, measuring the pressure variations induced by water-level fluctuations. Water surface movement can also be measured with upward looking acoustic sensors mounted on the bottom, which measure the elevation of the air-sea interface.

Surface Sensors

Wave buoys are typically used in deep water, where structures are not available to attach wave staffs and the water is too deep for bottom-mounted sensors. The application of buoys in coastal waters is limited by its inability to measure water depth. Nevertheless, most ocean observing systems rely heavily on buoys for sea state measurement.

There are three main types of wave staff: resistance wire, capacitance, and electromagnetic transmission line staffs. The operating principle of this instrument is that the staff is a component of an electronic circuit which produces voltage changes proportional to either the length of the staff that is or is not submerged. In turn this provides a measure of water level fluctuation. Wave staffs are often used in laboratory measurements. For field measurements, locations of wave-staff deployment are confined by mounting requirements. The mounting confinement also limits the options of incorporating directional measurements using wave-staff arrays. Biological fouling and electronic drifting limit the long-term applications of wave staffs.

Subsurface Sensors

The most commonly used wave sensor in coastal waters is the bottom-mounted pressure transducer, which measures the pressure variation induced by water-level fluctuations. Directional waves can be measured by combining a pressure sensor and a velocity sensor. This combination comprises the most commonly used directional system for nearshore wave measurement and is often called a PUV gage (Morang et al. 1997).

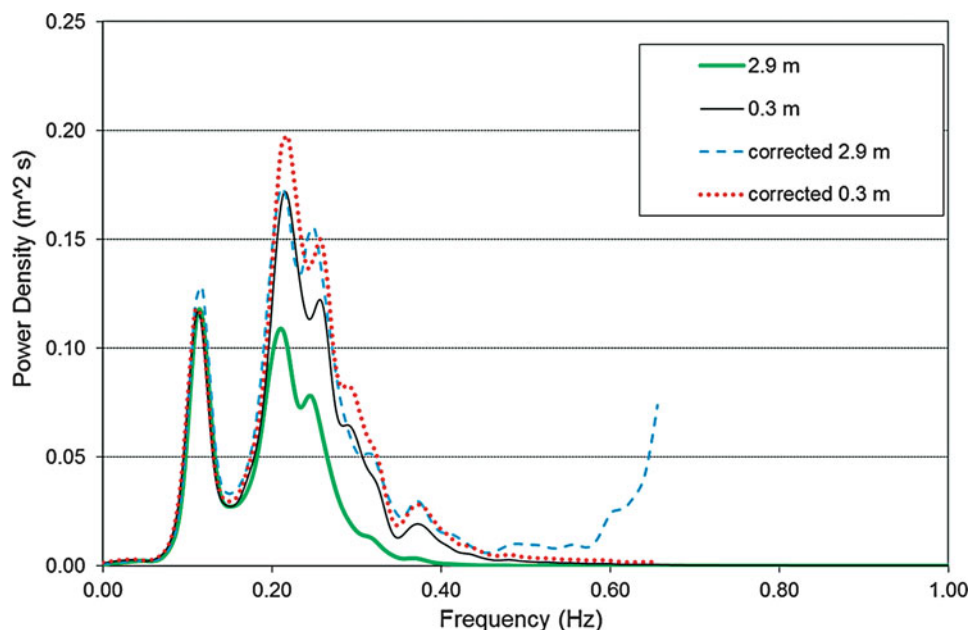
A primary limitation of bottom-mounted pressure and current sensors is the increased attenuation of signal strength with increased water depth, referred to as depth attenuation. This results in a decreasing signal-to-noise ratio with increasing water depth. The degree of depth attenuation is a strong function of frequency, with high frequency signals attenuating more rapidly than low-frequency signals. Depth attenuation of signal strength can be quantified by comparing measurements taken near the water surface and near the bottom (Fig. 1). For the example shown in Fig. 1, similar energy spectral density was measured at different water depth for low-frequency components, while for high-frequency components, greater energy spectral density was measured near the surface than that measured in deeper water.

Depth attenuation can be corrected during data analysis. A commonly used procedure, derived from linear wave theory, applies a frequency-dependent depth attenuation factor, $K(f)$

$$K(f) = \frac{\cosh[k(z_s + h)]}{\cosh(kh)} \quad (5)$$

Nearshore Wave

Measurement, Fig. 1 Wave spectral measured at different water depths, 2.9 m and 0.3 m below surface, respectively, and depth-attenuation correction



where z_s = depth of the sensor below the water surface (negative downward). Wave number and wavelength cannot be directly measured by the pressure sensor, or other point sensors, but are calculated based on the dispersion relation (Eq. 1). The measured power spectral density, $E_m(f)$, is corrected as

$$E_c(f) = \frac{E_m(f)}{K^2(f)} \quad (6)$$

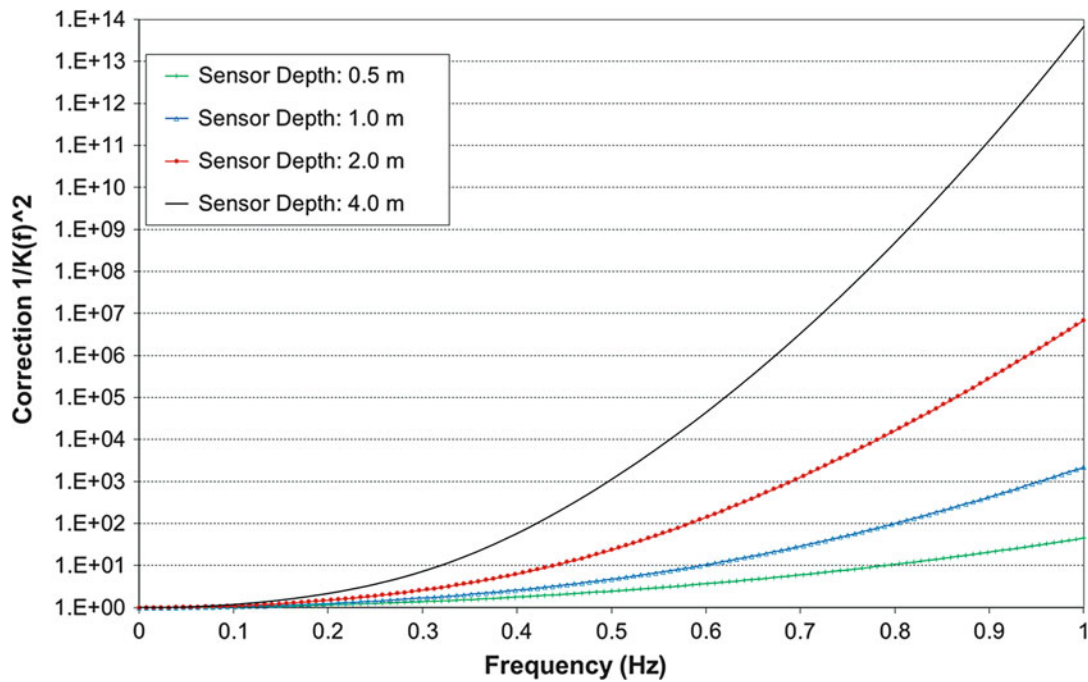
where $E_c(f)$ = corrected power spectral density. The attenuation-correction factor $K(f)$ increases exponentially as the frequency increases (Fig. 2).

The depth-attenuation correction Eqs. 5 and 6 is a strong function of wave frequency. This is shown by the sharp increase toward the tail of the corrected power spectrum for the sensor that was 2.9 m below the surface (Fig. 1). Since the attenuation correction amplifies both the signal and noise, the high-frequency noise can be amplified substantially, causing considerable error in computing wave height and sometimes, peak wave period. Therefore, the corrected spectra should be truncated at a certain frequency, above which excessive noise amplification has occurred. Noise amplification and the determination of a high-frequency cutoff depend on a variety of factors including the characteristics of the individual sensor and data collection system, the depth of the sensor, and the height and period of the waves.

Proper determination of the cutoff frequency has significant influence on the calculation of significant wave height and other statistical wave parameters in coastal waters, such as in estuaries, where local-wind generated short period waves often dominate. It is common for these waves to have

peak periods of 2–4 s (0.50–0.25 Hz). As shown in Fig. 2, if the sensor is mounted 4 m below the surface, a correction of over 10^3 will be applied to signals that are greater than 0.5 Hz. In other words, the noise beyond 0.5 Hz will be amplified 1000 times if the sensor is 3 m deep. For open-ocean waves, components shorter than 2-s period are typically negligible. High-frequency cutoffs ranging from 2 s to 4 s are often applied for open-ocean wave measurements. However, in coastal waters, especially in semi-enclosed coastal water bodies, wave components with a period less than 2 s may not be neglected. Determination of the proper high-frequency cutoff may be dictated by local conditions and careful study is necessary. The use of a fraction of the peak wave period may not be appropriate for bimodal spectra, which are common in coastal waters.

Recently, the commonly used Acoustic Doppler Current Profiler (ADCP) has been improved to measure directional wave. The rather complicated method and computation algorithm are discussed by Terray et al. (1999) and Strong et al. (2000) and are beyond the scope of this entry. The directional spectrum of wave energy can be estimated from both the ADCP velocity measurements and the direct echo-location of the surface along the four slant beams. The ADCP wave package is capable of yielding comprehensive point measurement of marine hydrodynamics including directional wave and current profile. A disadvantage of the ADCP wave package is that it requires a certain water depth, for example, deeper than 1.5 m, to function properly and it is also quite costly. In addition, its sampling rate is typically less than 2 Hz. These limit its application in some coastal environments which are characterized by shallow water and local wind waves, for example, in small estuaries and near the surf zone.



Nearshore Wave Measurement, Fig. 2 Increase of depth-attenuation correction with increasing frequency

Design of a Wave Measurement Project

Generally, the design of a wave-measurement project includes three aspects: (1) sampling rate of the sensor, often expressed in term of samples per second, or Hz; (2) length of the sampling; and (3) interval between each sampling events. The sampling rate and length are determined by the characteristics of the waves to be measured. The interval between samples is determined by several factors including the objectives of the research and capabilities of power supply and data storage.

The sampling rate should be rapid enough to capture the wave-induced water level fluctuation. The minimum frequency that can be detected by a certain sampling rate, referred to as the Nyquist (f_N) or folding frequency, is

$$f_N = \frac{1}{2dt} \quad (7)$$

where dt = sampling rate. For example, if the sampling rate is 2 Hz, the minimum frequency that can be resolved is 1 Hz. However, it is worth mentioning that 2 Hz sampling may not capture the full magnitude of the 1 Hz signal, although the frequency is resolved. It is always desirable to sample at a faster rate than twice of the Nyquist frequency, and the latter should be determined with adequate knowledge of the shortest frequency that needs to be resolved, not the peak wave frequency. Each sampling event is typically called a wave burst. The duration, or length, of a wave burst is

typically based on the peak wave period. Generally, in order to obtain reliable statistical values of wave parameters, the length of a wave burst should be at least 150 times the peak period. Another important factor that needs to be considered in determining the sampling duration is the low-frequency oscillation, related to wave grouping. The sampling duration should be long enough to ensure that low-frequency oscillations or wave grouping would not have significant influence on the wave measurement.

A general knowledge of the study area is important in determining the optimal sampling rate and wave-burst length. It is desirable to have a rapid sampling rate and long wave burst. However, these are often limited by the capacities of the power supply and data storage, especially for the bottom-mounted systems, which are typically powered by batteries and have a fixed amount of data storage. Optimal sampling design balances the storage-consuming fast sampling and the power-consuming long wave burst with the most effective frequency coverage and reliable statistical properties.

The interval between wave bursts is another important parameter in planning a wave measurement project. Wave conditions in coastal waters can change quickly, thus it is preferable to have shorter intervals between wave bursts. However, more frequent wave bursts require more power and data storage. If the main goal of a wave measurement project is to obtain long-term characteristics of wave conditions, longer intervals between wave bursts, such as 3 or even 6 h, may be acceptable. If the main goal is to capture wave growth or development and dissipation of storm waves,

shorter intervals, such as 1 h or shorter, are preferred. More than one sampling mode can be used, for example, a storm mode with 1 h interval and regular mode with a 3 h interval.

Wave Data Analysis

Nondirectional Wave Analysis

As discussed earlier, one sinusoidal form is not adequate to describe realistic waves in the ocean. Waves are best presented in a statistical manner using spectral energy density and probability distribution. Numerous procedures have developed for wave data analysis, largely based on concepts from time-series analysis and statistics. Details of wave analysis, especially that of directional waves, is complicated and beyond the scope of this entry. In the following, a procedure recommended by the Coastal and Hydraulic Laboratory of the US Army Corps of Engineers (Earle et al. 1995) is discussed as an example. This procedure has been widely used in computing wave parameters, such as wave height, period, and direction. It is important to be consistent and to adopt a commonly used data processing procedure to ensure compatibility.

A key component of wave data processing is spectral analysis. Spectral analysis determines the distribution of wave variance as a function of frequency. Wave variance is proportional to wave height squared, which in turn represents wave energy. Thus, wave variance spectra are often referred to as wave-energy spectra. Spectral analysis is conducted using the Fourier Transformation. Wave spectra are typically calculated with units of variance/frequency so that integration over a frequency range yields the total variance of the wave components within the frequency range. Significant wave height (H_{mo}), which was originally defined as the average of the one-third highest waves in a wave record, can be calculated as

$$H_{mo} = 4 \times \sqrt{\sum_1^m E_c(f) df} \quad (8)$$

where m = number of discrete Fourier frequencies in the frequency band, and df = small frequency interval over which the $E_c(f)$ is calculated. Another commonly used wave parameter is the peak wave period, which is defined as the reciprocal of the frequency of the maximum wave variance. Other wave height parameters, such as the root-mean-square wave height (H_{rms}) and 1/10 maximum ($H_{1/10}$) wave height can be calculated assuming a Rayleigh probability distribution of wave heights:

$$H_{rms} = \frac{\sqrt{2}}{2} H_{mo} \quad (9)$$

$$H_{1/10} = 1.80 H_{rms} = 1.27 H_{mo} \quad (10)$$

Directional Wave Analysis

Wave direction is important for applications such as harbor and structure design, wave refraction studies, and calculation of sediment transport rates (Massel 1996; Wang and Beck 2012). A directional wave spectrum, $E(f, \theta)$, is defined so that $E(f, \theta) df d\theta$ is the wave variance in a small frequency range df , and direction range $d\theta$. Integration over all directions provides the frequency spectrum $E(f)$:

$$E(f) = \int_0^{2\pi} E(f, \theta) d\theta \quad (11)$$

It is beyond the scope of this article to examine in detail the measurement and determination of directional wave spectra (Earle and Bishop 1984). Based on derivations of Longuet-Higgins et al. (1963), the directional spectrum can be simplified and represented by the first five Fourier coefficients

$$E(f, \theta) = \frac{a_o}{2} + \sum_{n=1}^2 [a_n \cos(n\theta) + b_n \sin(n\theta)] \quad (12)$$

The first five Fourier coefficients are calculated based on the cross spectrum analysis of two time series, such as water elevation and horizontal velocities. A cross spectrum consists of a cospectrum and a quadrature spectrum. The cospectrum represents the variance of the in phase components of the two time series. The quadrature spectrum represents the variance of the out of phase components.

Since PUV gages are commonly used in coastal water to obtain directional spectra, a simplified procedure of PUV data analysis is discussed in the following as an example. The first five Fourier coefficients are calculated as

$$a_o(f) = \frac{C_{pp}(f)}{K_p^2(f)\pi} \quad (13)$$

$$a_1(f) = \frac{C_{pu}(f)}{K_p(f)K_u(f)\pi(2\pi f)} \quad (14)$$

$$b_1(f) = \frac{C_{pv}(f)}{K_p(f)K_v(f)\pi(2\pi f)} \quad (15)$$

$$a_2(f) = \frac{C_{uu}(f) - C_{vv}(f)}{K_u^2(f)\pi(2f\pi)} \quad (16)$$

$$b_2(f) = \frac{2C_{uv}(f)}{K_u^2(f)\pi(2f\pi)} \quad (17)$$

where $C_{pp}(f)$ = power spectral density of pressure at frequency f , $C_{pu}(f)$ and $C_{pv}(f)$ = co-spectrum densities between pressure and velocities u and v , respectively, $C_{uv}(f)$ = co-spectrum density between u and v , $C_{vv}(f)$ and $C_{uu}(f)$ = power spectral densities of v and u , respectively, and $K_p(f)$ and $K_u(f)$ = depth attenuation correction coefficients for pressure and velocity, respectively, which can be calculated using Eq. 5. There are two methods to calculate wave angle ϕ using a_1 and b_1 , or a_2 and b_2

$$\varphi_m = \tan^{-1} \frac{b_1}{a_1} \quad (18)$$

or

$$\varphi_p = \frac{1}{2} \tan^{-1} \frac{b_2}{a_2} \quad (19)$$

The result from using Eq. 19 is often referred to as the principal wave angle to differentiate it from the mean angle from Eq. 18.

Summary

Measuring waves in coastal zones involves a complicated procedure and requires careful planning. Wave conditions in coastal zones can vary significantly from region to region. Coastal waves may also be substantially different from typical deep ocean waves in terms of spectral composition. Hardware and software developed for deep-ocean swell measurement, such as large diameter disk buoys, may not be applicable for coastal wave measurements. Planning a coastal wave measurement project should include several considerations including objectives of the measurement, a general knowledge of the waves in the study area, types of instrumentation, sampling scheme, and data retrieval and analysis procedures. For wave measurements in semi-enclosed environments with high frequency waves, it is important to have sensors close to the water surface to minimize the loss of important high frequency signals. A sound knowledge of the technical properties of different wave sensors and wave analysis procedures are also important in assembling an optimal system for a coastal wave measurement project.

Cross-References

- Coastal Climate
- Wave Climate
- Wave Environments
- Wave Hindcasting
- Wave Power
- Wave Refraction Diagrams
- Waves

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Net Transport

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Net transport refers to the difference between the total up-coast and the total downcoast movement of sand approximately parallel to the shore over a specified period, often one year (see entries on *Cross-Shore Sediment Transport* and *Longshore Sediment Transport*). Waves approaching the shore at an oblique angle, such that the crests of the breakers are at an angle to the shoreline, generate longshore currents that convey the mobilized sand along the shore (see entry on *Waves*). In many locales, the wave approach direction varies over time such that this transport can be in either direction relative to the shore. Consequently, sand may move down-coast (positive direction) for an interval and then, as wave conditions change or the lesser effects of tidal or wind-driven currents intervene, reverse and move upcoast (negative

Richard J. Seymour: deceased.

direction). Summing the volumes of sand transport over the period of interest and taking into account their signs yields net transport. The customary units of net transport are volumetric rates (cubic meters per year). The terms littoral drift and longshore drift are occasionally used to mean either net transport or gross transport (see entry on *Gross Transport*) so caution must be taken in their interpretation.

In some special locations, because of physical barriers or climatic anomalies, waves approach from only one direction. In this case, the net transport can be identical to the gross transport. In most instances, however, this is not the case. Studies of the potential for longshore transport, based upon directional measurements of nearshore waves at a large number of coastal locations, have shown that the net transport is often a very small difference between large values of positive and negative transport. For example, at Oceanside Harbor in California, during the entire year of 1980, the net transport potential observed was less than 1% of the gross transport (Castel and Seymour 1986).

Cross-References

- [Cross-Shore Sediment Transport](#)
- [Energy and Sediment Budgets of the Global Coastal Zone](#)
- [Gross Transport](#)
- [Longshore Sediment Transport](#)
- [Sediment Budget](#)
- [Waves](#)

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New Zealand, Coastal Ecology

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Introduction

New Zealand has the fourth largest exclusive economic zone (EEZ) in the world (over 4 million km²), and a long indented coast. The EEZ ranges from subtropical areas to subantarctic areas, though we confine our review to mainland New

Zealand (i.e., North, South, and Stewart Islands). Our discussion considers areas from the upper limit of saltwater penetration to the edge of the continental shelf (mean depth 200 m), though we focus on coastal regions. Coastal regions supply New Zealanders with food, recreation, and a livelihood, with high-value near-shore capture fisheries and aquaculture being important contributors to the nation's economy.

The important features of the complex current patterns around New Zealand are becoming better known. The west coast is generally characterized by turbid, cold, productive seas, served in some areas by upwelling of nutrient-rich waters. Subtropical water impinges on northern New Zealand, and may reach down both coasts of the North Island in summer, influencing catches of pelagic gamefish. Interannual variation in the extent of those currents is large, and in recent years many previously unrecorded fishes and other taxa have been observed. The Subtropical Convergence (STC), a global oceanic front separating warmer subtropical waters to the north from cooler subantarctic waters to the south, intersects New Zealand's South Island. The STC passes around the southern end of the South Island, then follows the east coast of the South Island before moving offshore near Christchurch.

Intertidal habitats of New Zealand's coast include mangrove forests, sand and mudflats, salt marshes, surf beaches, and rocky platforms. Subtidally there are rocky reefs, harbor channels, sandy beaches, sounds, and fiords. Kelp forests occupy reefs, and solitary corals occur, and although there are no true coral reefs, reef-building corals do occur at the Kermadec Islands, 800 km northeast of Auckland. Over the 25 years after 1975, knowledge of coastal ecosystems has increased greatly, though Schiel (1991) was still able to suggest that very few processes structuring nearshore communities were understood.

Review of Major Ecosystems

Estuaries

The major biological habitats occupying intertidal areas in estuaries are mangroves, *Spartina* grass, and mud- and sand-flats. Mangroves occupy estuaries in the northern half of the North Island, and are thought to have important sediment-stabilizing and ecological roles there. Studies of mangrove populations have quantitatively and experimentally examined demography, litter production, and the inter-relationships between sediments and mangroves, but there are no comprehensive analyses of their effects on other fauna. Marsh grasses have been most intensively investigated in more southern areas, where attempts to control the introduced *Spartina* have had variable success. Sea grass (*Zostera* sp.) beds also form important habitat on intertidal areas throughout New Zealand. In recent years, some areas of sea grass are thought to have declined, perhaps due to various human activities,

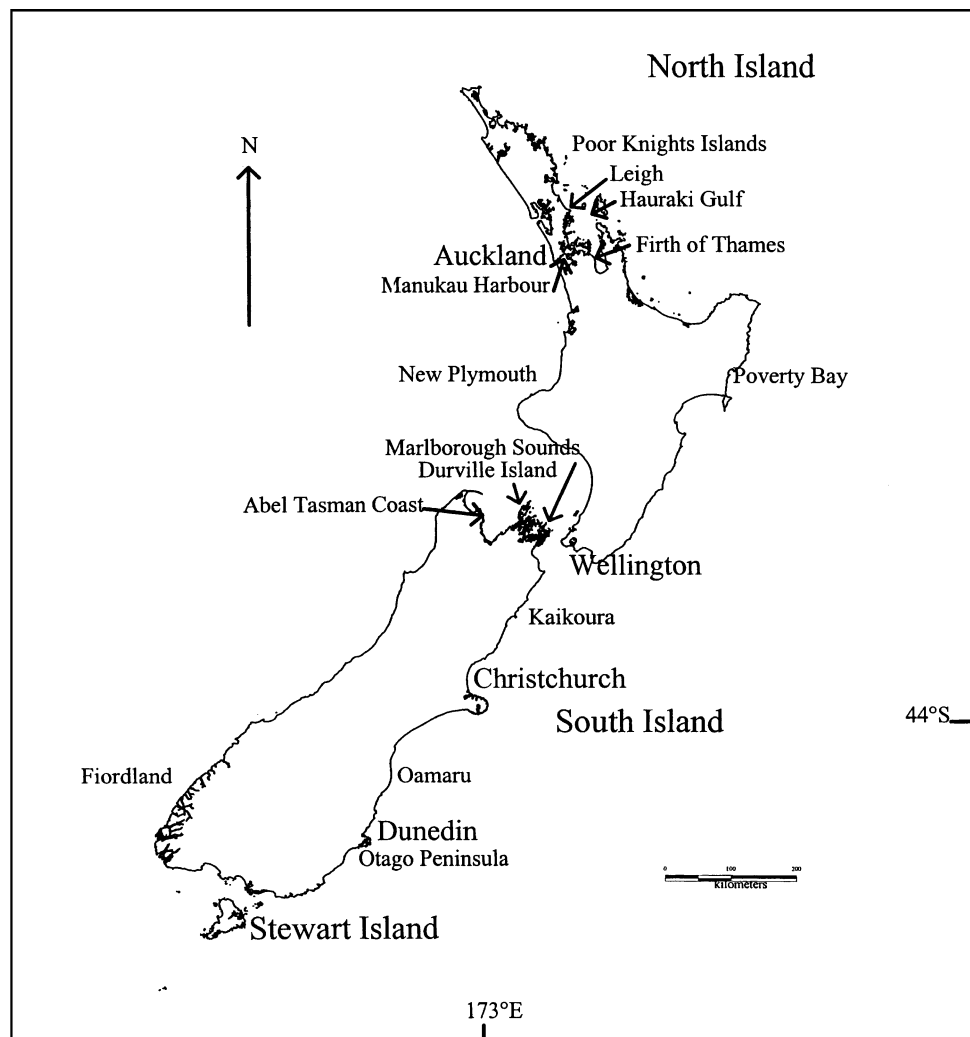
such as trampling on rocky intertidal platforms, or more subtle effects, perhaps related to water quality at the level of entire catchments. The other important plant of harbors is the green seaweed *Ulva lactuca*. Blooms of this species have occurred in several New Zealand estuaries, and are thought to indicate eutrophication. The seaweed also accumulates on shores of harbors after storms and rots, causing problems for local homes.

Throughout New Zealand the cockle *Austrovenus stutchburyi* is prominent in intertidal faunas of more sheltered shores, being harvested commercially throughout its latitudinal range. Several estuaries (e.g., Manukau, Firth of Thames; Fig. 1) are important foraging areas for populations of birds that migrate to the Northern Hemisphere. The best-known faunal components of those areas are clams, and intensive studies of interactions between sedimentology and biology have been undertaken in the Manukau Harbour (e.g., Thrush et al. 1997).

The dominant subtidal features of northern harbor channels are dense populations of pipi *Paphies australis*, often in

association with abundant 11-armed starfish *Coscinasterias muricata*. The dense covers of shell associated with such areas probably modify sediment mobility, and the abundances of shellfish are possibly high enough to influence water quality. Harbors are thought to be important to stocks of commercially important fishes, and the fish fauna of harbors and estuaries in northern New Zealand is being investigated in a comprehensive program at present (2000). The fauna of Otago Harbour, near Dunedin and areas of Stewart Island have also been described, and feature broadly similar organisms to those in more northern areas. One important aspect of population biology that is not well understood is how the population biology of even common species varies along latitudinal gradients. Casual observations suggest large differences in the densities and sizes of seaweeds, intertidal bivalves, echinoderms, and fishes between northern and southern populations, but as yet there have been few focused studies on latitudinal patterns. One study has compared sites on the east and west coast of the South Island, but it is not clear whether such differences relate to broader-scale

New Zealand, Coastal Ecology,
Fig. 1 Map of New Zealand,
showing localities mentioned in
the text



oceanographic processes or some other environmental difference between coasts.

Intertidal Rocky Shores

Considerable research investigating the rocky reef platforms in eastern South Island has been undertaken by University of Canterbury, with some additional comparative information available for the west coast. On the east coast, sea grass *Zostera* sp. and Neptune's necklace seaweed *Hormosira banksii* may occupy large areas of reef. Experimental investigations indicate complex patterns of demography of the sea grass that relate to position, and physical variables such as standing water and temperature. The seaweed is affected by trampling, but recovery is complex and influenced by season, location, and presence of corallines. Similar effects have been shown for trampling of coralline turfs in northeastern New Zealand.

Lower reaches of wave-exposed rocky shores of the southern half of the South Island are occupied by very large laminarians (Schiel 1994). Earlier studies indicated geographic differences in the influence of grazers on the extent occupied by the bull kelp *Durvillaea antarctica*, and other patterns in morphology that were related to wave action. Research into the seaweed karengo at Durville Island is investigating the possibility of enhancing populations for iwi (local Maori).

Subtidal Reefs

Most knowledge of subtidal reef ecology stems from work done in northeastern New Zealand, though additional information exists for New Plymouth, the Wellington coast, the Marlborough Sounds, the Abel Tasman coast, and Kaikoura (Fig. 1). The northern studies have emphasized the role of seaweeds in structuring habitats, the importance of small mobile invertebrates such as amphipods in contributing to the productivity of those reefs, and the role of humans in modifying the populations of organisms on those reefs (see below).

Seaweed morphology has been shown to be important in several disparate ways in northeastern New Zealand. Surface area of seaweeds may be important to photosynthetic processes, as ammonium metabolism appears confined to surface layers of cells. A series of studies have shown that the secondary productivity of reefs in northeastern New Zealand is highest in stands of finely structured seaweeds, and that under certain circumstances the excretory products of the small animals might be of value as a nitrogen source to the seaweeds. Studies near Leigh documented large differences in life expectancy, morphology, and palatability to grazers for wave-exposed and wave-sheltered populations of the large fucal *Carpophyllum flexuosum*. Investigations currently being undertaken near Dunedin examine the influence of wave action on large laminarian seaweeds directly by measuring wave forces.

The sea urchin *Evechinus chloroticus* forms prominent "barrens"—areas devoid of seaweeds—in northeastern New Zealand. Further south the extent of those barrens may be diminished, and in some areas, such as near New Plymouth, extensive barrens can occur in the presence of small abundances of *Evechinus*. In northern New Zealand, the large diadematid sea urchin *Centrostephanus rodgersii* may be abundant, and in deeper water the endemic diadematid *Diadema palmeri* may also be common. The other prominent grazers of New Zealand reefs are gastropods, which may be highly abundant both on the rocky seabed and on the fronds of seaweeds. Experimental investigations suggest that it is the grazing activities of *Evechinus* that structure seaweeds on reefs, and that gastropods mainly respond to the grazing of echinoids.

Seaweeds have also been shown to be important in structuring populations of fishes. Some fishes recruit into seaweed stands on reefs, others use detached seaweeds floating offshore as shelter (and may be moved onshore as a result of that association), still others feed mainly in areas where those seaweeds do not occur, and there is a growing base of knowledge regarding the biochemical pathways whereby herbivorous fishes process seaweeds. Small mobile crustaceans such as amphipods that contribute most of the secondary productivity on reefs, form an important component of diet for nearly all reef fishes.

Open Coast Beaches and Inner Shelf

The dynamics of fauna occupying sediments of the inner shelf are relatively unknown, though much research attention has been focused on the effects of dredging of scallops and trawling. Again, dense populations of bivalves occur, some of which are implicated in structuring populations of other organisms. The best known of these is the horse mussel *Atrina zelandica*, dense beds of which occur throughout New Zealand. *Atrina* has a fragile shell, appears to have strong and weak year classes, is known to influence sedimentation and meiofauna, and to be influenced by physical disturbances such as dredging and anchoring. Other bivalves are important in other habitats; dense populations of the venerid *Tawera spissa* occur in northeastern New Zealand, and in Marlborough Sounds large areas are occupied by live and dead individuals of the dog-cockle *Glycymeris laticostata*. Beds of both *Tawera* and *Glycymeris* are sufficiently dense that they potentially structure the fauna of areas. Although large amounts of drift seaweed may occur on the seabed offshore, most investigations have examined its decay onshore, or its role in concentrating plankton at the sea surface.

Studies of sewage outfalls and dredge spoil disposal in offshore areas such as Tauranga and Gisborne generally suggest that the influence of those disturbances are localized and/or difficult to separate from natural variation. Biological

investigations of coastal sediments have lagged behind physical oceanographic studies; in recent years the physical oceanography of the East Auckland Current, Manukau Harbour, Poverty Bay have all been described in some detail. The biological consequences of those processes are only well understood for the East Auckland Current, however, and there are few, if any, studies of the influences of such physical processes on organisms occupying sediments.

Fiords

The fiords of southwest South Island have an estuarine circulation system, which is maintained by the input of freshwater from the prodigious rainfall (>6 m per annum). Fiordland has a long complex coastline of inlets and islands, providing shelter from the prevailing westerly winds. Rainwater runoff from forests leach tannins from leaves and stains the surface fresh-water layer (up to 6 m thick) brown, limiting light penetration. Divers must pass through a murky freshwater layer in which visibility is poor, before bursting through into the clear saltwater below. In the low-light environment antipatharian corals may commonly be found as shallow as 10 m, and other deepwater emergent species also occur there. There are strong gradients in the distribution of several species along the length of the fiord (e.g., large brown seaweeds), and the freshwater layer influences grazing on mussels by starfishes. Starfishes are sensitive to salinity, whereas the mussels are able to resist freshwater incursions to some extent, so that mussels have a refuge in shallow water.

The Fiordland area is a renowned tourist destination, fishing area, source of hydro-electric power, and has been mooted for freshwater export. There is a readily accessible underwater observatory and a proposal for submarine tours. Visitors are concentrated in one readily accessible fiord, and other areas are subject to much lower visitor densities. However, some renowned dive destinations are under pressure from the large numbers of divers visiting. That effect is magnified as several of the species that are drawcards (such as red and black corals) are fragile and may be broken by careless divers. Such problems are likely to emerge with continuing diver pressure on areas such as Poor Knights Islands and Cape Rodney—Okakari Point Marine Reserve, near Leigh.

Marine Birds and Mammals

The best known New Zealand higher marine vertebrates are Hector's dolphin, seals, and penguins in southern New Zealand, and the huge flocks of wading birds in northern harbors. Hector's dolphin is endemic, permanently coastal, and thought to be endangered due partly to its susceptibility to set nets. Marine mammals and penguins are important tourist drawcards in southern New Zealand, particularly on the Otago Peninsula. Sperm whales are also abundant off the

Kaikoura coast feeding in the productive waters overlying the Kaikoura Canyon. Conflicts between fisheries and Hooker's sea lion occur in some offshore fisheries, and modifications to fishing practices are used to minimize seal mortalities. In some areas of northeastern New Zealand feeding by birds in intertidal areas may be sufficiently intense to locally reduce abundances of clams.

Fisheries and Aquaculture

New Zealand promotes a broad range of capture fisheries and aquaculture round its coast. Important coastal capture fisheries include spiny lobster *Jasus edwardsii* and snapper *Pagrus auratus*. Value of exports is increasing faster than the volume caught, reflecting at least partly improved handling techniques. Studies in several areas suggest that fisheries (both recreational and commercial) are a major influence on the marine environment and populations of marine animals. Environmental concerns are emerging regarding the environmental effects of some fisheries, and the role of commercial fishers is increasingly one of stewardship. The indigenous Maori people also have an important and growing stewardship role. Recreational fisheries can be important to populations of both the fish and the fishers, with snapper in the Hauraki Gulf, blue cod in the Marlborough Sounds, and kahawai throughout the country being among the most targeted species.

Studies of the benthic impact of trawl and dredge fisheries have revealed reductions in faunal richness and damage to other species. Some of those studies have revealed patterns at broad spatial scales that are consistent with fisheries being a dominant influence on benthic ecology. Concern regarding benthic impacts has led to closure of areas in northern New Zealand to scallop dredging. There are clear advantages for highly targeted, "clean" fisheries, which minimize bycatch. The benefits of such fisheries may extend into the overseas markets, where increasingly greater returns can be obtained for animals or plants that are captured in environmentally benign ways. Further, capture in such ways may have advantages in product quality and ultimately price, as greater shelf life follows from selective, highly controlled harvest.

The aquaculture of GreenShell™ mussels *Perna canaliculus* is a major industry, particularly in the Marlborough Sounds, but farms are found from the north of the North Island to Stewart Island. Mussels are grown on longlines, from which droppers bearing the mussels hang down (usually 10–12 m). Growers hope to harvest mussels within 14–17 months of seeding out, and aquaculture leases in productive areas are highly sought after. The environmental effects of such activities have been little investigated, and concerns regarding sustainability and benthic impacts are mounting. Localized depletion of phytoplankton within

farms has been documented, but increasingly there are concerns of depletion at the level of entire embayments. Measurements of currents, phytoplankton abundances, and computer models are being used to assess the viability of proposed farms. Environmental effects of salmon farming have been studied, but that activity occurs in only a few locations, mainly in southern New Zealand. Other cultured species include Pacific oysters, cockles *A. stutchburyi*, and developed paua (abalone) *Haliotis iris* ventures. As yet few pharmaceutical products are being harvested from aquaculture ventures. As yet few pharmaceutical products are being harvested from aquaculture ventures, though these have considerable potential for high returns.

Introduced Species

Some of New Zealand's important coastal fisheries are for introduced species. The Pacific oyster was introduced in the late 1960s or early 1970s, and is now an important aquaculture industry. Sea-run trout and salmon provide valuable recreational fisheries in the South Island, with a trawl bycatch of salmon having been important in the past. However, not all such introductions have positive outcomes, and international experiences (some as close as Tasmania) indicate that introduced species can have major negative effects on many aspects of coastal ecology. Other introduced species will colonize New Zealand shores, and it appears that managing the inevitable arrivals is the focus of current efforts. Pamphlets detailing candidate invaders are circulated by NZ Ministry of Fisheries, with the aim of detecting exotic arrivals early.

A pest management plan has devised for the Asian kelp *Undaria pinnatifida*. However, that species is atypical of introduced species in many ways, in that it is commercially valuable, and that it has an extremely dispersive phase, linked to human activities. *Undaria* has spread throughout southern New Zealand, as far north as Gisborne in Poverty Bay, and in a wide range of environments from harbors to wave-exposed shores. Intense efforts are being made to prevent its spread at Stewart Island, although the success of the seaweed elsewhere has been variable. In Marlborough Sounds, *Undaria* has flourished in sheltered areas, persisted on shallow sills that are impacted by ferry wash, survived well on marine farms that it has recruited to, but failed to persist in subtidal areas adjacent to farms in several areas. It appears that human activities such as boating and aquaculture are the major vector, that grazers on natural shores are able to limit the abundance of *Undaria* in sheltered area but perhaps not in wave-exposed areas (e.g., Oamaru), and that the abundance and species composition of animals that occupy *Undaria* are distinct from natural seaweed stands on the same shores. There are conflicting opinions regarding whether the species should be exploited as a valuable resource or whether it should be extirpated.

Another invader, the Asian date mussel *Musculista senhousia*, is confined to northern harbors and appears to have limited capacity for longterm occupation of areas. It forms matlike localized beds, but the beds appear not to persist. In the Marlborough Sounds some marine farms have had serious problems with recruitment of a solitary ascidian *Ciona intestinalis*. It is thought to compete with mussels for food, and heavy recruitments are sufficient to diminish growth. Concerns regarding biosecurity are mounting, and an awareness of the threat posed by organisms in ballast water has prompted studies of the fauna of ballast tanks, and investigations of ways of treating it to ensure that foreign organisms would not survive. However, other species are able to survive externally on ship hulls (e.g., *Undaria*), and to protect the value of New Zealand's marine resources, great emphasis will have to be directed into surveillance of our borders, and studies of the biodiversity and susceptibility of natural populations. The experience with *Undaria* indicates that studies elsewhere are not able to be directly transferred to the New Zealand situation.

Marine Reserves

New Zealand has been at the forefront of marine reserve creation since 1975, and as of October 2000 has 16 marine reserves in which all marine life is protected under the Marine Reserves Act 1971. One of the limitations of that legislation is that it permits marine reserves to be established solely for the purpose of scientific study, and public enthusiasm for marine reserves as a conservation method seems to have outstripped the need for scientific investigation. Several other areas are protected under different legislation, and designated as marine parks. There are also several taiapure and mataitai reserves, which are under the control of local Maori to varying degrees. Marine reserve legislation is currently under review to determine whether the current Marine Reserves Act 1971 is appropriate. It seems likely that it will be reshaped so as to give a broader range of reasons for protecting particular areas.

These protected areas have provided considerable information regarding ecological interactions in coastal habitats. The best-known marine reserve is the Cape Rodney to Okakari Point Marine Reserve, near Leigh, north of Auckland, which has been fully protected for almost 25 years (as of 2000). Partly because of its ready access, but also perhaps because of the duration of protection, it has provided considerable information regarding interactions among organisms on coastal reefs. Fish and spiny lobster populations have been investigated in great detail, and there is some indication that neighboring areas benefit from the existence of the reserve. The most interesting development has been the loss of sea urchin-dominated "barren grounds" (areas with no seaweed) at Leigh, and the reduction of their

area in several other northern reserves. It has been suggested that the replacement of barren grounds by kelp forests in reserves reflects the high abundance of predators that consume sea urchins, though that conclusion is contentious. Those studies are of considerable importance internationally, since only in New Zealand have completely protected marine reserves been established for sufficient duration to determine long-term effects.

Scientific investigations targeting blue cod and spiny lobsters have recently been undertaken in more southern marine reserves, and have generally found evidence of increased abundances and/or size similar to those done in North Island marine reserves. As more studies are undertaken in reserves the generality of effects will emerge. It is not surprising that humans have important effects on populations of marine organisms, given the intensity of fishing effort, improved positioning capacity via GPS, and improved quality of access due to changes in both roading and vessel capabilities. Strong gradients of fish abundance and size in relation to access and/or protection status are obvious in many fisheries.

Summary

New Zealand's coast contains most physical environments and biological habitats common in temperate regions worldwide. Knowledge of physical and biological environments is increasing rapidly. Direct and indirect human activities are important influences on marine systems. Coastal developments, introduced species, aquaculture and marine reserves will be important influences on nearshore marine systems in the future.

Cross-References

- [Aquaculture](#)
- [Climate Patterns in the Coastal Zone](#)
- [Estuaries](#)
- [Mangroves, Ecology](#)
- [Marine Parks](#)
- [New Zealand, Coastal Geomorphology and Oceanography](#)
- [Rock Coast Processes](#)
- [Sandy Coasts](#)
- [Vegetated Coasts](#)

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New Zealand, Coastal Geomorphology and Oceanography

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Historical Background

New Zealand's coastal geomorphology exhibits a myriad of coastal landforms and sedimentary deposits, varying from estuarine types of fiords, rias, barrier enclosed estuarine lagoons, eroded calderas, and flooded grabens; to rocky cliffs and shore platforms, raised coastal terraces; to modern sandy Holocene barrier islands and spits, modern long sandy beaches and bayhead pocket beaches; and various types of nearshore sedimentary facies, to Pleistocene paleo-barriers; to various types of mixed sand-gravel beaches, boulder beaches and barriers; to muddy tidal flats and shelly gravel chenier ridges and beaches.

From the early 1900s, the classical geomorphology of New Zealand was widely expounded in the publications of Sir Charles Cotton, and especially in his book "*Geomorphology of New Zealand Part 1: Systematic*" first published in 1922 and revised as "*Geomorphology: An Introduction to the Study of Landforms*" (Cotton 1942). About one-fifth of this classical text was devoted to coastal processes and landforms, including especially submerged and emerged coasts. Cotton (1974) later published the book "*Bold Coasts*" comprising a series of his papers on rias, cliffs, and rocky coasts and drawing widely on New Zealand examples. Cotton's pioneer work was focused primarily on deductive interpretation of erosional geomorphic features such as cliffs, headlands, and "old hat" platforms, from the Davisian geomorphic cycle perspective. Little emphasis was accorded coastal sedimentary deposits or investigation of coastal processes, although naturally at that time there was little technological capability to measure coastal geomorphic processes. Likewise some of the early New Zealand geologists and geographers, viz. P. Marshall, J. Bartrum, and G. Jobberns, wrote deductive papers on aspects of coastal geomorphology, for example, the origin of gravel in the Hawkes Bay beaches, and the origin of shore

Terry R. Healy: deceased.

platform high level benches around the Auckland area. Jobberns was the first specialist “coastal geomorphologist,” and researched the raised strandlines of North Canterbury (McLean 1977). But at that time there was no fundamental research involving quantitative measurement of coastal sedimentary deposits or landforming processes. Healy and Kirk (1982) addressed this issue, and presented a more modern synthesis of New Zealand coastal geomorphology.

Schofield (1960) is credited with undertaking the first substantive surveys combined with shallow cores and dating to obtain quantitative data for New Zealand coastal evolution. His pioneering paper on the origin of a sequence of 13 progradational chenier ridges at the muddy head of the Firth of Thames attempted to relate progradational phases to oscillations of late Holocene sea levels. Later Schofield (1970) also published the first substantive regional beach sedimentological investigations based upon textural and mineralogical analysis of multiple samples from representative sand beaches of the northern half of the North Island.

The first modern systematic beach studies – involving quantitative measurement of form and process of Canterbury mixed sand-gravel beaches (McLean and Kirk 1969) – were established at University of Canterbury in the 1960s by R.F. McLean, who may be regarded as the “father” of modern coastal geomorphology in New Zealand (Fig. 1). McLean’s student, R.M. Kirk, has continued coastal geomorphic studies at Canterbury since the 1970s. In 1973, sandy beach and estuarine sedimentary and process research was initiated at University of Waikato by T.R. Healy, where since 1980, and in collaboration with the pioneering developments in numerical simulation of two- and three-dimensional flows, waves, and sediment transport modeling of K.P. Black, the largest undergraduate and graduate school for coastal geomorphic and oceanographic studies in the country developed in the 1990s. Coastal geomorphic studies were also established at University of Auckland from the late 1970s, initially under the leadership of R.F. McLean and subsequently by K. Parnell. In the 1980s, a coastal unit with emphasis on numerical modeling was established within the then Ministry of Works and Development, but with restructuring of scientific research in the early 1990s, this group became part of the

National Institute of Water and Atmosphere (NIWA), and under the leadership of T.M. Hume, has since grown into the largest group of coastal geomorphic and oceanographic researchers with strong national standing in New Zealand.

Environmental Factors Influencing New Zealand Coastal Geomorphology and Processes

The distinctive diversity of coastal geomorphic types and sedimentary deposits characteristic of the extensive New Zealand coastline evolves from a combination of geological structure, tectonic and seismic history, lithology, a mid-latitude oceanic setting for wave and tidal processes, Pleistocene events, and climatic influences.

Geological structure and lithology. A series of lineal axial ranges comprise the essential structural backbone of New Zealand. In the North Island, these consist of lightly metamorphosed and intensely jointed Mesozoic greywacke, and in the South Island, of greywackes and metamorphosed Paleozoic schists of Otago and the alpine fault zone through to gneiss and granodiorites of south Westland. On-lapping the axial ranges are younger Tertiary rocks, predominantly soft clay-rich siltstones and some limestones, while ancient volcanic mounds punctuate the landscape. In the central North Island the active Taupo Volcanic Zone (TVZ), stretches from Lake Taupo to beyond the Bay of Plenty coast. The northern North Island exhibits evidence of back arc andesitic and acid volcanism, with basaltic volcanism around Auckland and Northland. This diversity of lithology provides a wide variety of topographic forms and sediment for distinctive beach types and coastal landscapes.

Influence of structure and Pleistocene events. New Zealand, like Japan and northern California, sits astride a plate margin where the coastal geomorphology closely reflects the adjacent lithology as well as historical geological events. Mesozoic and Tertiary orogenies created the basic axial ranges, the backbone of the country. Ongoing plate margin evolution saw the development of the transcurrent Alpine Fault stretching from the south of the South Island along the southern Alps and across Cook Strait to Wellington and through to Hawkes Bay. Along this transverse fault system the active earthquakes and tectonic dislocations have influenced the coastal geomorphology by creation of drowned transverse fault aligned valleys (rias) of the Marlborough Sounds and uplifted coastal terraces around Wellington.

An important element of structure is the alignment of the main islands comprising New Zealand along the plate margin and active subduction zone. Subduction of the Pacific plate under the central North Island results in a number of geologic processes including the active TVZ, rapid tectonic uplift, folding, faulting, and stratal slumping with associated high seismic activity. These seismic and tectonic (uplift

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Fig. 1 Professor Roger McLean, who may be regarded as the “father” of modern coastal geomorphology in New Zealand



and downsinking) processes in the geological evolution, which are continuing at present, have occurred on both a regional and local scale. They enhance denudation of the hilly and highland catchments, providing high sediment loads to the coasts. Regional downsinking has helped create the rias of the Marlborough Sounds, while regional uplift in the central North Island has resulted in incised rivers without significant estuaries (e.g., Waitara River) cut into coastal terraces.

Events in the Pleistocene also played a major role in molding the modern coastal landscapes. Severe valley glaciation in the south Westland gneissic province produced glaciated U-shaped valleys, which upon drowning in the Holocene transgression have become classical fiords. The intense physical weathering of the fractured greywackes and schists, associated with the glaciations in the Southern Alps, provided outwash sands and gravels to build up alluvial fan deposits, the most striking example of which is the Canterbury Plains. These subsequently became blanketed in loess. Periglacial weathering action in the ranges of the North Island assisted in delivering a large volume of gravel sediment to the lowland plains and coast, notably in Hawkes Bay. Of course at that time the lowland plains extended 100–120 m below present sea level, so that many sand deposits were reworked over the inner continental shelf as sea level rose with the post-glacial transgression.

Influence of volcanism. Both modern and ancient volcanic influences are evident in the coastal geomorphic landscape. Eroded caldera in the ancient volcanic mound of Banks Peninsula today contain the harbors of Lyttelton and Akaroa. Volcanic deposits have also had an important influence in the central North Island west coast where erosion of lahar flows provides both cobbles and boulders for veneer beaches over eroding shore platforms, as well as the distinctive black titanomagnetite heavy mineral sand for the extensive sand littoral drift systems of the North Island west coast. Propylitized Miocene andesites and Pliocene rhyolites and ignimbrites comprising the Coromandel Ranges are back arc volcanics and have provided distinctive mineralogy for the beach sands of the embayed east Coromandel Peninsula.

Around the Auckland isthmus some of the numerous Pleistocene basalt volcanic cones are now eroded headlands, with the outstanding landscape feature being the recently formed island of Rangitoto, which erupted as Hawaiian-type fluid basalt flows, and is surmounted by a scoria cone crater active as recently as 600 years ago. Pleistocene and modern active volcanism from the TVZ has demonstrably influenced coastal evolution in the Bay of Plenty, where denudation of the acid volcanic tephra has provided large volumes of sand to the littoral zone. This has accumulated in the broad embayments to create extensive Holocene dune ridge progradations. The eruption of Mt. Tarawera (1886), which blanketed the surrounding landscape with pumiceous deposits, provides

detailed evidence of the impact of a large influx of sandy sediment into the coastal littoral system, and the air-fall ash deposit on the coastal dunes allows reliable dating of progradation rates (Pullar and Selby 1971; Richmond et al. 1984). The ability to use the technique of tephrochronology has allowed reliable dating of Pleistocene and Holocene dune and barrier deposits and interpretation of rates of geomorphic evolution along both the west and east coasts of the North Island.

Wave and climate influences. The mid-latitude oceanic islands of New Zealand are dominated by southwesterly waves originating from the Southern Oceans, and driven by the prevailing southwesterly winds. These waves, typically of period ~8–11 s and height ~1–3 m, drive littoral drift systems within regional and local coastal compartments along the west coasts. On the east coast of the South Island and the southeast coast of the North Island the wave climate is likewise predominantly swell from the southern oceans. North of East Cape (North Island), the embayed coast – a “lee coast” from the prevailing westerly winds – is sheltered from the southerly swells and is subject to a mixture of distant Pacific swells and local storm and wind-generated waves, typically with periods of 7–9 s and heights of 0.5–1.5 m from variable onshore directions. Episodic storms, especially decaying tropical cyclones from the north, generate an important component of the wave climate which cause dune erosion problems.

The El Niño–Southern Oscillation (ENSO) plays a major role in the beach geomorphology and coastal oceanography. In El Niño conditions New Zealand experiences continual 15–20 knot winds from the southwest. This causes a surfeit of orographically induced rainfall and flooding particularly in Westland of the South Island, and enhancing sediment supply to the coast, while the constant onshore wind stress induces downwelling on the coast and inner shelf. But on the east coasts, particularly of the North Island, the continual offshore wind stress induces drought conditions, onshore transport of nearshore sands, and inner shelf oceanographic upwelling. As a result the east coast beaches become generally well accreted in El Niño conditions and the coastal upwelling brings nutrient-rich ocean waters to the inner shelf, which enhances conditions for algal and phytoplankton blooms in the nearshore coastal waters and harbors – a condition which severely impacts on fisheries and aquaculture industries. During the La Nina phase the opposite conditions occur, with more frequent, stronger northeasterly winds causing erosional tendencies for the northeast coast beaches, but upwelling induced algal blooms on the west coast. To date it is not clear whether the west coast beaches significantly accrete during La Nina conditions.

Tidal effects. Tides are of the semi-diurnal type. Walters et al. (2002) calculated the tidal amplitudes around the oceanic islands of New Zealand based upon the predominant M2 constituent, to produce an approximate mean tide range.

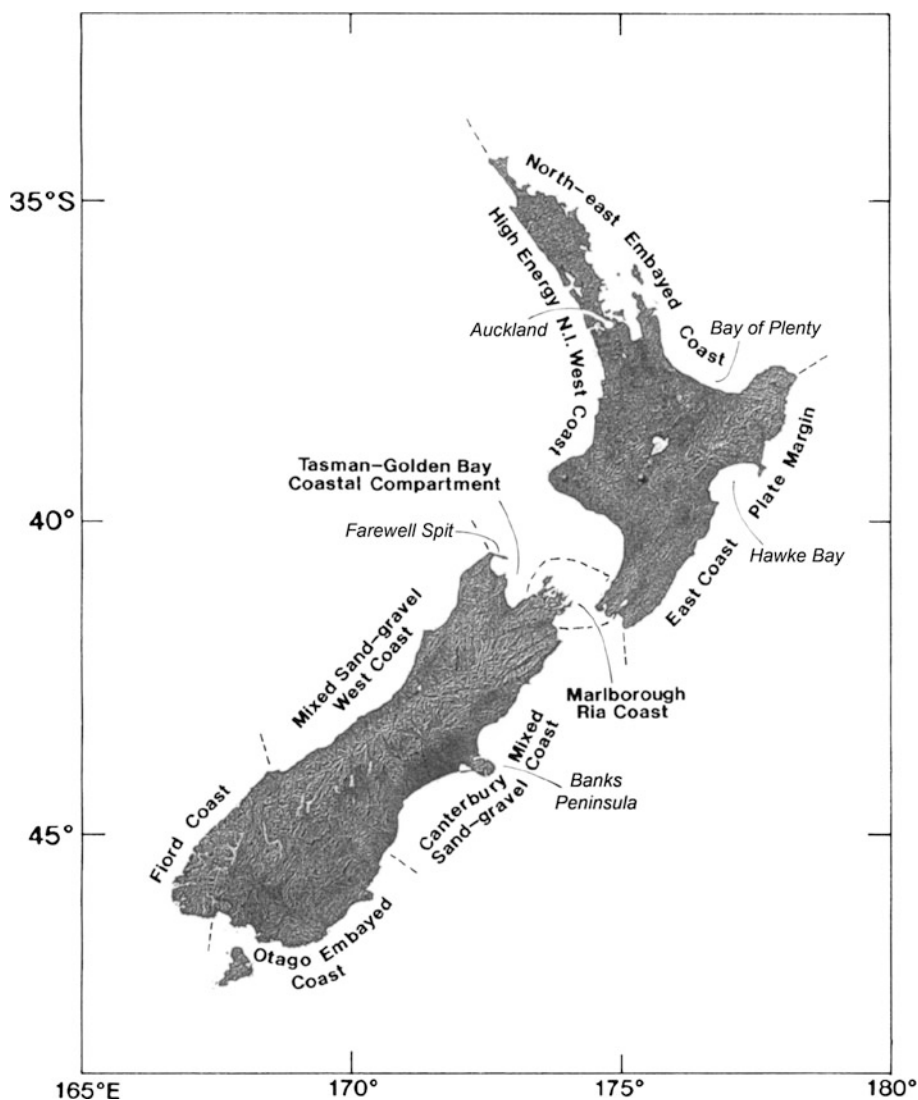
Based upon the tidal gages at the “standard ports,” the average tidal range is 1.8 m with spring means of 2.1 m and neap means of 1.4 m. The tidal ranges are predominantly micro (<2 m mean tide). Largest tidal ranges (>3.5 m mean tide) occur in Tasman and Golden bays in the northern South Island, and along the central west coast (e.g., Onehunga, Port Taranaki) and exhibit a mesotidal (>2 m to <4 m) mean range. However, spring tidal ranges in some areas approach 4 m. Lowest tidal ranges occur along the East Coast region. In most ports and harbors (e.g., Bluff, Lyttelton, Dunedin and Timaru ports and Manukau, Kaipara and Whangarei harbors) the tidal range is amplified relative to the open coast tidal range, indicating a hypersynchronous tidal condition, while two (Wellington, and Picton) are suppressed, indicating a hyposynchronous tidal condition. The difference in tidal phase between the west and east coast results in strong tidal currents through Cook Strait separating the main islands.

Regional Coastal Geomorphology

The New Zealand coastline can be classified into several distinctive geomorphic sectors (Fig. 2).

High-energy North Island west coast. This sector is subjected to high wave energies with prevailing onshore winds, and a divergent littoral drift system, flowing northwards north of Cape Egmont and north of Wellington, and southwards south of Cape Egmont. The coastline is essentially curvilinear in alignment. Around Taranaki, the coastline exhibits low cliffs and rugged terrain, with gravel boulders littering the shore platforms cut into the lahar deposits which flowed from Mount Taranaki. The high wave energies and swell wave refraction around small promontories create numerous locations for good surfing waves, with renown “breaks” at Raglan and other locations. The distinctive fine “black ironsands” comprise titanomagnetite heavy mineral mainly eroded from the Taranaki lahars, which in some Pleistocene dune deposits

New Zealand, Coastal Geomorphology and Oceanography, Fig. 2 New Zealand coastal geomorphic provinces



is of sufficient concentration to mine (Waverly, Taharoa). There is a general reduction of heavy mineral concentrations away from Taranaki, both north and south. South of Cape Egmont, wave refraction induces a southeastward moving drift, but north of Cape Egmont there is a consistent northward drift to North Cape. Beaches along this sector typically exhibit a dissipative morphodynamic regime. Some of the littoral system is cliffed but the high-energy waves also move sand on the submarine cut platforms of the inner shelf. The coastal sector centered on Taranaki is also undergoing rapid tectonic uplift forming coastal terraces, so that rivers, such as the Waitara, are actively downcutting and no large estuaries occur. North of Kawhia Harbor a number of large shallow barrier enclosed estuarine lagoon harbors occur (Aotea, Raglan, Manukau, and Kaipara). Each is a tidal inlet type, but unusual in that the barriers enclosing the harbors comprise Pleistocene dunes up to 300 m high, and depending upon tidal discharge relative to wave exposure, these tidal inlets tend to have well formed fan shaped ebb-tidal deltas. Because of the high tidal ranges the currents in the inlet gorges tend to be strong – as much as 4 m/s.

Northeast embayed “lee” coast. Between North Cape and East Cape the coast is characteristically indented, including several large embayments with many islands (e.g., Bay of Islands, Hauraki Gulf). The embayments possess compartmentalized sandy littoral drift systems, some of which may be quite extensive (e.g., Bay of Plenty, Bream Bay, Doubtless Bay, Pakiri-Mangawhai), and contain extensive Holocene dune ridge sandy barrier progradation systems. The typically hilly catchments and active erosion of the soft Tertiary lithologies means that the drowned river valleys have to a large extent become infilled with sediment, so that the harbors are typically shallow and display extensive intertidal flats. At the mouth of many former drowned river valleys are Holocene sandy dune ridge barrier spits (e.g., Pauanui, Matarangi), which enclose shallow infilling estuarine lagoons. Within the Hauraki Gulf the soft Miocene alternating sandstone–siltstone flysch rocks have undergone rapid erosion in the Holocene and today display active cliffs and extensive shore platforms up to 400 m wide.

The Bay of Plenty contains the longest littoral system of the northeast region, and is dominated by low refracted swell waves of $T = 7\text{--}9$ s and $H = 0.5\text{--}1.0$ m. The bay is notable for containing the largest Holocene progradational extent of dune sands, 9 km in the Rangitaiki Plains, and also the largest barrier island (Matakana Island) in the country. The extensive Tauranga Harbor, geomorphically a barrier enclosed estuarine lagoon, is ~70% exposed intertidal flats at low tide, and possesses a unique enclosing barrier system comprising Holocene barrier tombolos and an intervening barrier island. The surfeit of sand in the Bay of Plenty system is derived from the adjacent TVZ, having been brought to the coast during the Pleistocene, and reworked across the shelf during the

Holocene post-glacial transgression, with continuous additions from some of the larger rivers.

The Firth of Thames graben is not typical of this “lee” coast, but is notable for containing a major area of muddy coast, with deep Holocene mud deposits, broad intertidal mud deposits, and a well-developed shelly chenier ridge plain (Healy 2002).

East Coast Plate Margin coast. Extending from East Cape to the Wairarapa and Wellington is a rocky, eroding high-energy, mainly cliffed coast with extensive shore platforms cut into the Tertiary rocks. The coast is surmounted by veneer sandy pocket beaches in shallow embayments, and often backed by raised marine cut terraces. This coast is seismically active and undergoing rapid uplift due to subduction processes. The general NE–SW alignment is broken by two large embayments, Poverty Bay and Hawke Bay. Poverty Bay lowlands comprise intercollated dune ridges and alluvial deposits. The high discharges from the Waipaoa River eject considerable muddy suspended sediment from the soft Tertiary mudstone catchments into the nearshore and continental shelf. Thus paradoxically, this coast is high energy but possesses very muddy sediment at shallow depths due to the high mud supply. Hawke Bay is likewise subjected to high muddy river discharges to produce muddy sediment at shallow depths, but the rivers here also deliver greywacke from the central axial ranges, so that the beaches are predominantly composed of mixed sand and angular greywacke gravel. Resulting from the Napier earthquake of 1931, a large area of shallow seafloor and harbor was uplifted ~2 m to create new coastal lowland backed by degraded cliffs where marine processes are no longer active.

Marlborough Sounds ria coast. Fault aligned valleys, transverse to the coastal alignment, result in classical rias (*sensu stricto*) known as the Marlborough Sounds. The structural alignment and transcurrent faulting have created long linear rias. The steep hillslopes dip below the waterline with only minimal interruption to the profile and there is relatively little infilling sedimentation except at the valley heads. Thus the rias remain deep. Wellington Harbor on the southern North Island is likewise part of the ria coast.

Canterbury mixed sand–gravel coast. This is a distinctive and unusual littoral drift coastal sector formed from high-energy storm and swell wave processes eroding the alluvial outwash gravel deposits comprising the Canterbury plains (Kirk 1980). The result is a well-formed steep berm beach comprised of mixed sand–gravel sediments which are fed partly by direct erosion of the outwash gravel deposits and partly from episodic inputs from the several large braided gravel river systems. Along sectors of the coast the active synoptic beach berm is surmounted by higher storm berms which reach almost 20 m above sea level adjacent to Banks Peninsula. Erosion of the outwash gravel deposits has resulted in low coastal cliffs behind the beach along the southern

sector. South of Banks Peninsula backing dunes are rare, but they do occur on the sector north of the Peninsula. Banks Peninsula itself comprises a Miocene andesitic volcanic mound with eroded calderas creating present-day harbors. Eroded loess deposits blanketing the hills have provided muddy infill sediments within the caldera harbors.

Southeast Otago cliffed and mildly embayed coast. Low hills of southeast Otago result in a mainly cliffed coast with pocket sand beaches and where rivers reach the sea. The region receives large swells from the southern oceans, and the beaches typically exhibit high-energy morpho-dynamic conditions. Otago Harbor is a ria that has become largely infilled.

Fiord Coast of South Westland. The southwest tip of New Zealand comprises a gneiss and granodiorite massif standing up to 3,000 m above sea level. At the coast precipitous slopes drop down to, and below, sea level. Some boulder beaches occur but the coastline is predominantly steep. Pleistocene valley glaciations have carved a spectacular landscape of classical U-shaped valleys, forming classical fiords where they have become flooded at the coast (e.g., Milford Sound, Doubtful Sound). The location in the zone of “roaring forties” and high topography means that the region receives very high precipitation, which provides fluxes of freshwater to drive a fiord-type estuarine salinity structure and circulation. As the area comprises national park the natural environment of this spectacular forested coastline remains in an essentially pristine condition.

Mixed sand–gravel sector of the South Island west coast. Waves from the southern oceans drive a northward littoral drift with sediment being fed to the littoral system by the numerous braided gravel rivers draining from the Southern Alps. The beaches in the southern sector are mixed sand-gravel type, and the wave climate is high energy. As in the North Island west coast littoral system, sand evidently bypasses rocky cliffs, and ultimately feeds the large sand deposits of the 30 km Farewell Spit at the northern tip of the South Island.

Golden and Tasman bays sandy compartment. Located between the Marlborough Sounds and Farewell Spit are sandy compartments. Two features are remarkable here, namely the existence of a low dune ridge Holocene barrier island (one of only two in New Zealand) and the 13 km long Nelson boulder barrier spit, which encloses a shallow muddy lagoon of Nelson Haven. About half of the enclosed area has been reclaimed, and the region possesses one of the highest tidal ranges in the country.

Major Coastal Geomorphic and Oceanographic Research Achievements in the Modern Era (Post-1970)

In the 1960s, the modern era of quantitative and process-based coastal research in the universities commenced.

Research was initially centered on shore platforms, and beach morphology and sediments based upon the contemporary concepts of the sediment budget and process response model (McLean 1977). Hume et al. (1992) outline the major achievements of coastal geomorphic and oceanographic research until about 1990.

Estuarine sedimentation. In the 1970s, there were rising concerns about the impact of changing land use practices – especially urbanization – on the water quality, biology, and sedimentation in estuaries. Early integrated interdisciplinary studies were initiated on the Avon–Heathcote estuary, the Pauatahanui Inlet, and the Waitemata Harbor. Numerous systematic sedimentological studies followed, often undertaken as thesis projects from the universities, especially from the Waikato University group (Healy et al. 1996). Most studies tended to concentrate on either the ecology or surficial sediments and only limited interdisciplinary investigations occurred.

Tidal inlets. New Zealand tidal inlets tend to have formed where Holocene barrier spits form across the embayment at the entrance to reentrant valley estuaries. They are found predominantly in the northeastern North Island where about 30 such features exist. They possess the typical morphological features of a narrow inlet gorge, and ebb and flood delta sand bodies, and are dynamic equilibrium systems. R.A. Heath investigated the tidal hydraulic equilibrium of the major inlets. The first quantitative hydrodynamics and sediment transport investigations were carried out by R.J. Davies-Colley around the entrance to Tauranga Harbor in 1976, where, prompted by port developments, the most comprehensive and detailed studies in a New Zealand inlet have since been carried out. Hume and Herdendorf (1990) provide a comprehensive compilation of morphological and empirical factors controlling New Zealand tidal inlet stability.

Numerical modeling of coastal hydrodynamics and sediment transport. Hydrodynamic numerical modeling of estuarine and inlet tidal flows was introduced to New Zealand relatively early by the pioneering work of K.P. Black when he developed his prototype one- and two-dimensional models for current flows and sediment transport in the tidal inlet at Whangarei Harbor (Black 1983). For calibration and verification of the model, as well as wider understanding of the inlet physical system, an extensive field research program included hydrographic surveys, water level recorders, side-scan sonar mapping of the bottom sediments, a wide range of bottom photographs and sediment sampling taken by SCUBA diving, continuous recording current meters, drogue tracking of current flows, numerous tidal cycle vertical current profiles, and sediment traps, and underwater video records and sediment threshold experiments, and wave refraction studies. Subsequently modeling of estuarine and inlet hydrodynamics has been widely applied (Black et al. 1999), initially in Tauranga Harbor, but also in most of the ports, and across a

wide range of applications. In more recent years, the models have become three-dimensional (Black et al. 2000) and applied to a range of EIA problems such as pollution dispersion associated with outfall effluents.

Tsunami modeling and research. New Zealand has experienced at least 32 identifiable tsunami events during its recorded history, but to date without the disastrous consequences experienced by Japan. The largest reported tsunami wave elevations were of order 10 m, originating primarily from local sources. The largest pan-Pacific tsunami recorded was the May 1960 Chilean tsunami which produced a wave elevation of 7 m at Whitianga in Mercury Bay. The most comprehensive tsunami research has been carried out by de Lange and Healy (1986) and their students. As part of disaster emergency planning and response, several tsunami scenarios have now been modeled for coastal sectors and harbors identified as potentially susceptible to significant tsunami impacts, for example, the Waitemata Harbor of Auckland city, the Bay of Plenty, Gisborne, Wellington Harbor, Port Lyttelton, and the Canterbury coast. Modern research is focusing on the ability of submarine slumping mechanisms to generate localized tsunami waves as occurred in March and May, 1947, centred around Tatapouri north of Gisborne.

Beach erosion and morphodynamics. As in many countries, severe beach erosion, especially when coastal property is placed at risk, spawns research programs on beach morphodynamics and beach budgets. The first integrated coastal erosion survey was along the Bay of Plenty coast in 1976–77, where episodes of erosion resulted in rapid duneline retreat, loss of houses, and construction of sea walls. The most spectacular example of erosion was along the embayed Holocene barrier spit of Omaha Beach, where in 1978 the seawall fronting a new subdivision was demolished at the height of a storm. This led to several studies seeking the cause and appropriate remediation, as well as litigation. Although the subject of much “expert” debate at the time, most would agree today that the cause was related to a long-term negative sand budget due to historical sand mining from the spit, consequent duneline retreat, and severe wave energy focusing in the storm. Other cases of erosion occurred locally as a result of interruptions to the littoral drift such as at inlets (e.g., Ohiwa Harbor) or downdrift of port developments (e.g., Port Napier and Port Taranaki), or due to sand mining from the beach (Papamoa in the Bay of Plenty). Concern for the sustainability and impact of long-term sand extraction from the nearshore zone along the 30 km Pakiri–Mangawhai coastal compartment resulted in a large and comprehensive study from 1996 to 1999. The study, lead by NIWA in collaboration with the universities of Waikato and Auckland, involved assessment of sand volumes onshore in the Holocene dune fields, investigation of the surficial sediment patterns from side-scan sonar mapping, bottom photographs and sedimentology, and shallow subsurface seismic profiling with calibrating vibrocore

data. Historical beach profile data were analyzed for beach sediment budget. Oceanographic data were measured with recording current meters and wave gauges, and the wider Hauraki Gulf subjected to hydrodynamic and sediment transport modeling.

Beach–shore face–inner shelf sedimentation dynamics. It became evident from a number of studies that strict wave driven alongshore littoral drift is often difficult to ascertain for the compartmentalized New Zealand coast. The earliest studies linking the beach with the inner shelf sediments arose from investigations into severe beach erosion episodes at Omaha and Mangawhai beaches north of Auckland and the east Coromandel beaches, which had suffered during the “decade of erosion” from 1968 to 1978. The most detailed studies during the 1980s were on the east Coromandel “lee” shelf by B. Bradshaw and collaborators. These studies linked closely with the detailed research monitoring the dispersal of sandy dredged material on the inner shelf off Tauranga Harbor by the Waikato school. During the decade of the 1990s, NIWA undertook detailed research on inner shelf and beach sedimentary morphodynamic linkages offshore from the Katikati inlet to Tauranga Harbor.

Port developments and dredging issues. Requirements for port development EIAs have driven considerable research into the impact of dredging and spoil disposal. The most detailed work has been undertaken around the Port of Tauranga, a port which has expanded to become the largest export–import in the country during the last decade. The large Tauranga Harbor Study (1983–85), the most intensive harbor study to date, arose from the need to investigate whether dredging through the tidal delta inlet system had substantially affected the channel hydraulics and morphodynamic changes of the flood tidal delta and channels. Subsequent studies were undertaken on the dredge spoil dispersion on the inner shelf and shelf–beach interaction, as well as further studies on the impact of deepening the channels to take post-panamax sized vessels. Those physical and sedimentological impact studies were matched with ecological impact studies. Disposal of dredged material became a major environmental and litigation issue for the Port of Auckland, and resulted in an inquiry spawning numerous reports from the Dredging Options Advisory Group (DOAG).

Coastal hazard analysis and development setback planning. Assessment of coastal hazard and development setback has been a requirement for coastal local authorities since the 1970s. The first New Zealand application of quantifying the independent components of coastal hazard along a sandy duned coast was evolved by T.R. Healy for the setback of new subdivisions along the Bay of Plenty coast in 1976, and included quantitative assessment of the four independent parameters of longterm erosion (or accretion) trends, short (decadal) term “cut and fill” duneline fluctuations, assessment of expected sea-level rise effects, and the dune topographic

stability factor. Summation of these four independent factors allows determination of an initial setback estimate, measured from the toe of the frontal dune, and based upon a planning time frame of 100 years (Healy and Dean 2000). The methodology has been widely applied and is similar to that later proposed by Gibb (1981). Over recent years the methodology has been refined and the initial setback determination is now subjected to three tests, viz. (1) Is there sufficient reservoir of sand in the frontal dune to allow for the 1: 100 year storm erosion episode? (2) Is the setback sufficient to allow for the 100 year storm surge? and (3) Is the setback sufficient to allow for tsunami washover? If the initial setback estimate is still subject to the above hazards, the setback is extended landwards beyond the zone of hazardous impact. The methodology may also be applied to existing subdivisions, in which the setback zone becomes a zone of non-further development.

Methodology for hazard analysis of development setback for cliffed coasts is given by Moon and Healy (1994) based upon identification of the mechanism of slope failure, and addition of a safety factor. For the cliffs cut into Miocene flysch deposits of the Waitemata Formation around the city of Auckland, the typical 100 year hazard setback is taken as ~16 m plus a safety factor of 7 m.

Innovative coastal environmental solutions and engineering applications. Early coastal engineering solutions to coastal erosion problems followed standard international practice of the early twentieth century, and groins and sea-walls were erected in many locations of localized coastal erosion. Most have subsequently failed or require expensive continuous maintenance. Since the 1970s, however, the environmental philosophy of allowing “nature to function in nature’s way” has tended to hold sway, so that much scientific effort has been expended on understanding the coastal dynamic systems rather than resorting to “hard” engineering solutions. Many innovative applications have been demonstrated. Among the more notable are: (1) The application of *sediment transport modeling* at the design stage by K.P. Black in 1983 to determine the optimum length of bridge span relative to causeway length for the Tauranga Harbor bridge, which would minimize scour and shallow estuary morphodynamic impact. This resulted in a reduction of the steel bridge span and a saving of about \$800,000 to the cost of the bridge construction. This application firmly demonstrated the value of numerical simulation as a major coastal processes tool. (2) The development of the concept of *artificial surfing reefs* by K.P. Black. This initiative originated during the late 1990s (Black et al. 1997) with the dual purposes of coastal protection against beach erosion, and for enhancing breaking waves for surf board riders. The first artificial reef so designed was constructed at Narrowneck on the Queensland Gold Coast in 1999. The reef units themselves are comprised of geotextile bags filled with dredged sand, a low cost construction system. Several such artificial surfing reefs are

planned for New Zealand, including at Mt. Maunganui in the Bay of Plenty, New Plymouth and Opunake on the Taranaki coast, and at Summer Beach near Christchurch. (3) In the late 1990s, considerable work has been undertaken on the *re-design of Port Gisborne* to accommodate the large expansion of timber product for export coming on stream from 2005. An intensive field and modeling program has been undertaken and initial designs subjected to wave and sediment transport modeling (Healy et al. 1998). This innovative application is to include substantial public amenity in the re-designed port-enclosing breakwater so that it acts as an artificial surfing reef to enhance a left hand break for the Pacific swells refracting into Poverty Bay. (4) *Artificial headlands for low-energy estuarine beach geomorphology* have been designed and implemented as a coastal erosion and coastal management option in the Waitemata Harbor, Auckland (2000). The artificial headlands use the principle that wave refraction into embayments at high tide will reduce sediment loss from the upper estuarine beach face by wave driven littoral drift, and induce retention of the sand and shelly gravel within the embayments. (5) The *use of dredge spoil for beach replenishment* has been applied at a few open coastal locations, including near Port Napier, Port of Tauranga, and Port Taranaki. These cases have used the concept of depositing the sandy dredge spoil as a berm in the shallow nearshore, with wave action taking the sediment on to the beach. The most detailed investigation has recently been for Port Taranaki (1997–99) where an extensive field program, including a large-scale tracer experiment, and supplemented by wave and current driven numerical simulation, has identified a suitable disposal site, which in conjunction with a planned artificial surfing reef, aims to re-establish the beaches of New Plymouth which had disappeared as a result of updrift port breakwater construction (McComb et al. 2000).

Issues for the Future

Research in the early years of the 2000 millennium is likely to concentrate on ongoing port developments and environmental impacts of dredging and disposal. Under development is investigation of the nature of mud deposition in ports and marinas, and attempts to induce its removal by “mud re-agitation” techniques, which requires ongoing field studies and numerical simulation of the results. Issues of the impact of sediment extraction from the littoral zone and the importance of closure depth and the links between diabathic and parabathic exchange will come under increasing scrutiny, and continuing developments can be expected in the frontier of morphodynamic numerical simulation of sedimentary bodies. Considerable effort is being undertaken to research beach and frontal dune behavior as part of “Dune-Care” programs in which local communities become stakeholders in the

management and protection of the dunes. With the advent of modern data collection the application of marine GIS linked with dynamic modeling, presently unknown in the New Zealand context, will develop for wide application. Overall New Zealand has a sound basis of research achievements to tackle the coastal morphodynamic issues of the new millennium.

Cross-References

- [Barrier](#)
- [El Niño–Southern Oscillation \(ENSO\)](#)
- [Gravel Barriers](#)
- [Longshore Sediment Transport](#)
- [Submerging Coasts](#)
- [Surfing](#)
- [Tidal Inlets](#)
- [Tsunami Deposits](#)

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North America, Coastal Ecology

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The North American coastal ecosystems range in latitude from tropical Mexico to arctic Canada and have long been identified as among the most productive in the world (Odum 1963). In reality, there is a broad range of annual productivities (ranging from several hundred to several thousand grams carbon) as well as species diversity. In the Arctic few species dominate the coastal ecosystems, while in the south along the Mexican coasts species diversity is especially rich (Lot et al. 1993). Widely varying in geomorphology and chemistry, they comprise: hypersaline lagoons, cheniers, coralline islands, temperate estuaries and barrier islands of the US Atlantic coast; as well as vast stretches of beach and permafrost coastlines of northern Canada and a myriad of rocky intertidal

communities and embayments along the Pacific Coast to Mexico (Odum et al. 1974). As in most ecosystems, solar radiation exerts primary control on productivity and ultimately impacts distributions of the numerous species. Also critical in influencing the structure and function of these coastal systems are organic matter and nutrient exports from upland sources, as well as tidal energy. In contrast to terrestrial systems where precipitation is considered second in importance to input solar radiation, the force of astronomically driven tides is extremely important in structuring coastal systems which range along this coastline from below a centimeter in some marshes of Chesapeake Bay (Stevenson et al. 2001) to over 10 m in the Bay of Fundy (Chmura et al. 2001).

Water, critical in determining the productivity of terrestrial systems, is the key medium for exchange of energy and nutrients in coastal systems. Generally, inflows of freshwater from rivers are responsible for supplying organic materials and nutrients to coastal estuaries and lagoons as well as control of seasonal changes in salinity. Freshwater inflow plus tidal activity (along with bathymetric configuration of coastal water bodies) largely determine rates of mixing and flushing. Thus, the hydrological inputs at varying local and regional scales are critical in modulating responses to increasing nutrient loadings, which cause numerous changes in coastal ecosystems (Malone et al. 1993; Turner and Rabalais 1994; Bricker and Stevenson 1996; Vorosmarty and Petersen 2000).

Drainage Basin and River Discharge

Although the magnitude of freshwater runoff is responsible for structuring estuaries (Pritchard 1967), water quality is critical in determining their productivity. Increasing the flux of relatively few critical elements (e.g., nitrogen, phosphorus, and silica) in freshwater discharges to coastal waters can cause eutrophication (i.e., abundant nutrients promote high phytoplankton biomass which consumes large amounts of oxygen as it decomposes) negatively impacting submersed grasses and traditional fisheries (Twilley et al. 1985; Stevenson et al. 1993). Water balance/runoff models, geographic information systems (GIS), and remote sensing are tools increasingly used to relate changes in land use to fluxes of nitrogen, phosphorus, and organic carbon into coastal systems. The generalized watershed loading function (GWLF) is an example of a widely used model to estimate nitrogen and phosphorus loading from runoff and groundwater in northeastern watersheds of the United States (Haith and Shoemaker 1987). In this model hydrologic fluxes are driven by daily weather data, including rainfall, evapotranspiration, and snowpack melting (the last two inferred from temperatures). Groundwater fluxes are estimated from considering both vadose (unsaturated) and phreatic

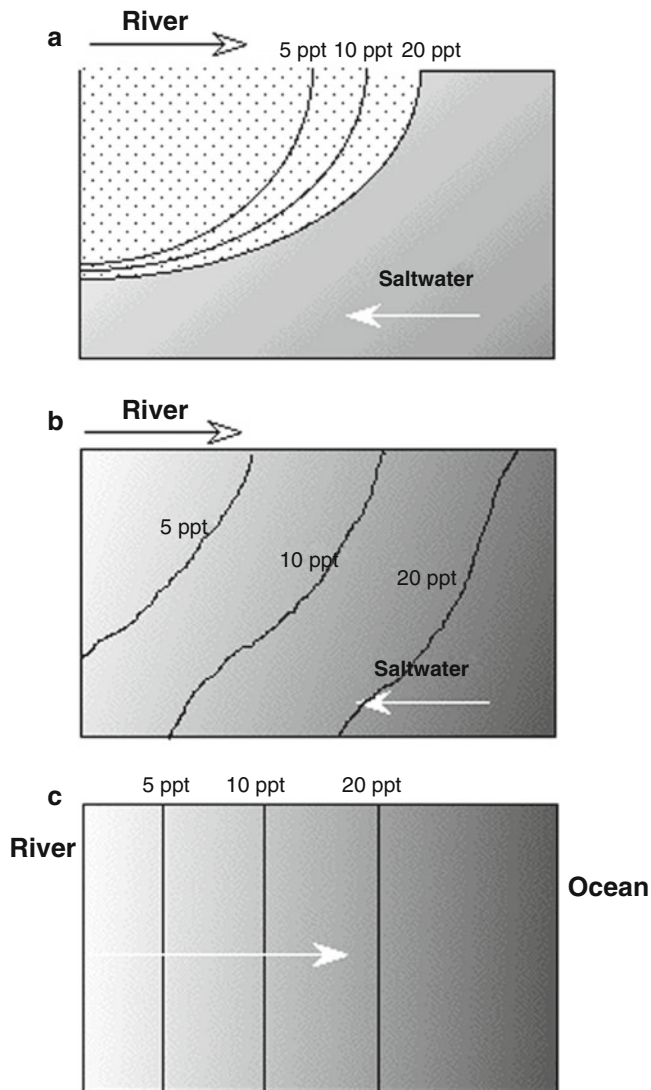
(saturated) contributions in watershed soils. Suspended sediment inputs, important in coastal turbidity and as vectors for trace metals (and other particle-reactive pollutants), have traditionally been estimated from the Universal Soil Loss Equation (Gottschalk 1964).

Once nitrogen, phosphorus, and other materials enter receiving waters they are subject to bioprocessing by various biochemical processes. Remineralization of organic material at the edges of coastal watersheds occurs when bacteria interact with dissolved and particulate phases, in wetlands and rivers. First-order streams, are perhaps the most important elements of coastal watersheds in bioprocessing as the time and volume of water in contact with wetland and the benthos is greater than in higher order streams (Peterson et al. 2001). Nitrogen, the most common element in fertilizer, transported as nitrate in streams or volatilized to ammonium (NH_4^+), often ends up in coastal embayments (Staver and Brinsfield 1996). However, since biogeochemical buffering often occurs, actual nutrient delivery to coastal waters is not as great as watershed yield coefficients of particular land uses might suggest.

Sources of Mixing

Waves, tides, and rivers control mixing in marginal environments. In estuaries, where major rivers meet the ocean, mixing is predominantly the result of tidal and river forces (Pritchard 1967). Although there is an expansion of water surface area in estuaries with the incoming tide, particularly those of the low relief Atlantic coastal plain, the total net volume of water remains approximately the same throughout the tidal cycle; total estuarine volume rarely changes much (Berner and Berner 1996). Essentially, incoming salty oceanic waters are transported landwards along the bottom balancing freshwater loss above the pycnocline in estuaries (Pritchard 1954). Mixing occurs along the salinity gradient resulting in complex circulation patterns which are often estuary specific and impact the dispersion of sediment particles and other pollutants (Nichols and Biggs 1985). These gradients generally depend on whether river discharge dominates the system, or tides (i.e., tidal currents).

A salt-wedge circulation occurs where inflowing river discharge greatly exceeds the volume of the incoming tide (expressed as the tidal prism) (Fig. 1a). In situations where the river velocities are high, freshwater can over-ride the saltwater, yielding a sharp interface where mixing occurs by internal waves at the interface lifting parcels of saltwater into the inflowing river water (Nichols and Biggs 1985). This distinct boundary is revealed by the tight clustering of isohalines (lines connecting water depths of equal salinity) where salinities can change from 5 to 30 ppt over vertical distance as little as 50 cm (Postma 1980).



North America, Coastal Ecology, Fig. 1 Types of general estuarine circulation and mixing. (a) Salt wedge or highly stratified system, (b) Partially mixed system, (c) Fully mixed system

A partially mixed system occurs where stronger tidal currents diminish domination of the circulation by river flow, and the resulting greater tidal mixing produces a less sharply defined boundary between fresh and saltwater (Fig. 1b). The stronger tidal currents produce both upward mixing of the river water and downward mixing of the saltwater. Partially mixed estuaries can be identified by a balance between river discharge and tidal prism. Fully or well-mixed circulation in an estuary occurs where the tidal prism is significantly larger than river discharge, and the tidal currents retard any tendency toward stratification of fresh and saltwater (Fig. 1c). The result is that the estuary becomes well mixed with depth, and salinities vary only laterally, becoming more saline down estuary.

As critical to the fate of nutrients and pollutants in coastal waters is the flushing time, that is, the time it takes for the

existing volume of freshwater to be replaced by river discharge (Aston 1980), or

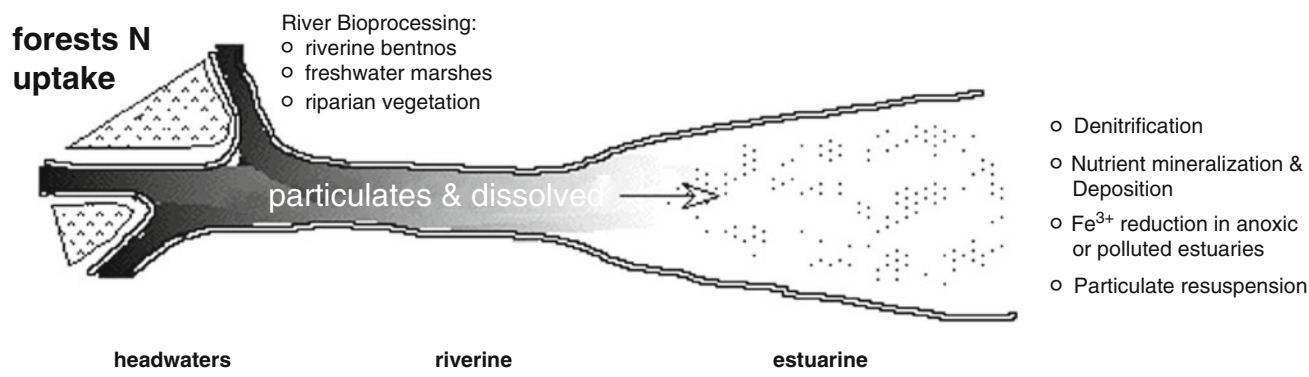
$$\tau = V_f/R,$$

where τ is the flushing time, V_f is the volume of freshwater in the estuary, and R is the river discharge into the estuary. Generally, the flushing times for less stratified and well-mixed estuaries tend to be longer than for more stratified, especially salt-wedge systems, where the fresh river water can flow out the estuary. Longer flushing times translate into longer periods for nutrients to be removed from estuary waters, with the visible evidence of such removal being an overall increase in phytoplankton biomass. In contrast, the faster flushing times for stratified estuaries, phytoplankton nutrient uptake is less, and phytoplankton debris is more likely to be exported from the estuary, slowing nutrient recycling (Berner and Berner 1996). Moreover, stratified estuaries are more likely to become nutrient traps for remineralized nutrients in bottom waters (Redfield et al. 1963).

Waves are important mechanisms of mixing in coastal waters. In salt wedge and stratified estuaries, internal waves produce mixing along the boundary of river discharge and incoming flood waters. In shallow water areas, waves, especially during storms, can fully mix the entire water column. At the same time, such waves may resuspend fine-grained particulates from the bottom. Wave mixing and resuspension also predominate as agents of mixing in shallow coastal lagoons, although considerable mixing from tidal exchange occurs within the vicinity of inlets. A measure of the amount of exchange that occurs in coastal lagoons can be estimated by comparing the volume of the lagoon to the *tidal prism* of the inlet. This is in essence an index of the amount of inlet “water capture” – that is, what volume of the lagoon is exchanged by a particular inlet. In large coastal lagoons, serviced by several inlets, significant variations in inlet tidal prisms probably influence circulation within the lagoon. Net drift in the lagoon is controlled by the tidal prism of the largest inlet.

Biogeochemistry of Estuaries

Spatial and temporal changes in the concentration of nutrients and other chemical constituents of the waters of Atlantic Coast estuaries have had considerable attention, especially as indicators of eutrophication. Studies have focused on sources of chemical constituents, vectors (if associated with particulates), cycling and transformations, and sinks. In simplest terms, the challenge of estuarine biogeochemistry is distinguished between terrestrial, marine and *autochthonous* materials, and their fates (Fig. 2 shows a general model of estuarine biochemical interactions).



North America, Coastal Ecology, Fig. 2 General biochemical interactions for a North American coastal plain estuary. (Based on Berner and Berner 1996)

Riverine Constituents other than Nutrients

Among the common chemical nonorganic constituents of coastal plain estuaries are Fe, Al, and Si. These constituents are commonly regarded as being of fluvial origin and indicative, when in high concentrations, of high rates of weathering within the drainage basin. Si, though naturally abundant in river water as dissolved silica, can also originate from groundwater sources. Various planktonic radiolarians and diatoms also contribute significant percentages. Dissolution of biogenic silica, however, is slowed by high concentrations of Fe^{3+} and Al^{3+} (Aston 1980). Because much of the Al and Fe entering estuaries is often adsorbed on to fine clay particles, rapid flocculation of river clays between 5‰ and 15‰ salinity probably removes most of these metals (Aston 1980). Moreover, this suggests that dissolution rates of biogenic silica are probably greater in the higher salinity (lower) parts of estuaries. Cations, such as Mg^{2+} , Ca^{2+} , K^+ , and Na^+ , estuarine waters are also largely derived riverine sources. In the rivers that flow through the highly weathered saprolitic rocks that compose much of the Piedmont of the Atlantic Seaboard's watersheds, concentrations of these ions tend to be high.

Dissolved organic matter (DOM) is also believed to be largely a function of riverine inputs. There is considerable seasonal as well as year-to-year variability in DOM, depending on river discharge and regional temperatures.

Nutrient constituents: phosphorus. The origin of phosphorus in estuarine waters largely reflects the influx of phosphate minerals in organic particulates and dissolved phosphate. Increasingly, the anthropogenic origin of phosphorus in sewage and other effluents has become a major concern in many North American estuaries. In Chesapeake Bay, Boynton et al. (1995) suggested that the total phosphorus load since European settlement had increased 16.5-fold, presently averaging about $0.065 \text{ t km}^{-2} \text{ yr}^{-1}$. In contrast to nitrogen, annual phosphorus loads largely reflect basin-wide runoff inputs, not groundwater inputs. Thus, the anthropogenic effects tied to land cover and land use changes (especially in regard to

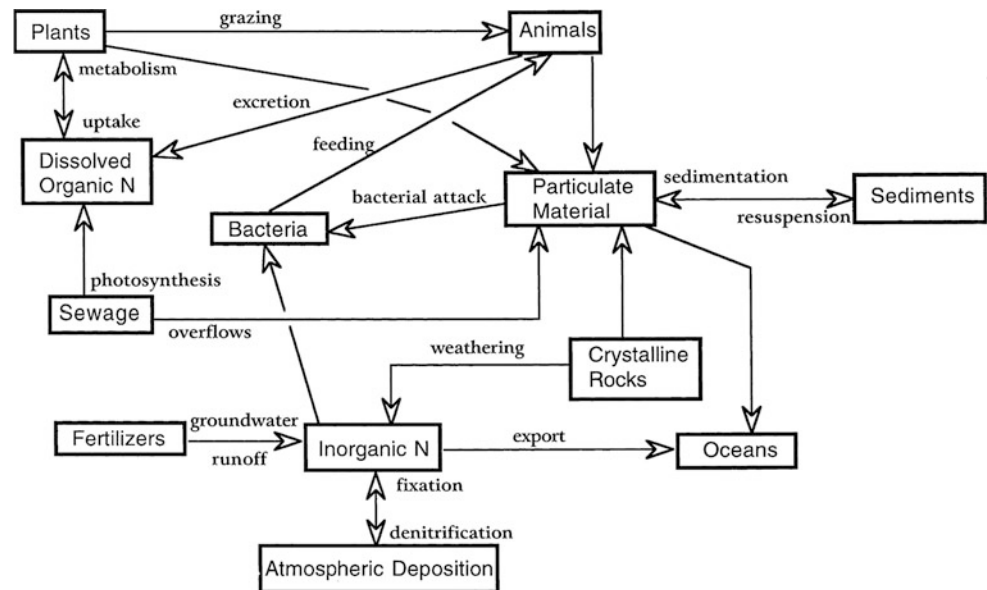
runoff efficiency as well as fertilizer application) are critical to understanding pathways for phosphorus (and nitrogen) export to coastal waters. The decline in forest cover, in particular, has been demonstrated to be a major factor in rising nutrient loads.

Once in coastal waters, particularly estuaries, phosphorus (as phosphate) is subjected to inorganic and biological controls. Phosphate "buffering" by being adsorbed on to mineral surfaces (Pomeroy et al. 1965) is perhaps the major inorganic control. Depending on the phosphate concentration, adsorption surface available (i.e., mineral species and abundance) pH (lower salinities and more acidic pHs than seawater) and, possibly redox (higher oxidation states in estuarine sediments possibly releasing phosphates), phosphate buffering can be a principal means of moderating sewage outfalls in coastal plain estuaries (Aston 1980).

The biological cycling of phosphorus in US Atlantic Coastal Plain estuaries has been shown to exhibit a seasonal cycle. Concentrations of dissolved phosphates in ambient waters are highest in summer (Taft and Taylor 1976). Bacteria and, to a lesser extent, phytoplankton (depending on turbidity and its effects on light penetration and intensity) play the principal roles in uptake of phosphorus in the phosphorus cycle (Berner and Berner 1996). However, the uptake of phosphorus by such organisms may be temporary since consumption by protozoans and other filter feeders with excretion and death can release back to the water column dissolved and particulate organic phosphate compounds (Aston 1980). Ultimate removal may occur upon to adjacent ocean waters.

Nutrient constituents: nitrogen. The impact of increased loadings of nitrogen upon the ecology of coastal waters of North America has been perhaps even greater than that of phosphorus. The input of nitrogen (as nitrate) to Atlantic Coast rivers increased by 30% during the late 1970s (Smith et al. 1987). Coastal waters are generally nitrogen limited due to denitrification by bacteria of nitrate (NO_3^-) under anaerobic

North America, Coastal Ecology, Fig. 3 General nitrogen cycle for North America coasts. (Modified after Aston 1980, with information from Boynton et al. 1995)



conditions, low nitrogen fixing due to low light conditions (Howarth 1988), and the toxicity of sulfides to nitrogen-fixing bacteria (Mitsch and Gosselink 1995). Fluxes of nitrogen into coastal waters from rivers occur both as dissolved nitrate and particulate forms, though the latter probably only predominate during high precipitation events, particularly if these events coincide with the application of fertilizers within local drainage basins (Fig. 3). Considerable research in Chesapeake Bay (Staver and Brinsfield 1996) has underscored the importance of groundwater discharge as the major source of nitrogen loading of coastal estuaries either directly into coastal waters, or by discharges into tributaries as stream base flow. Bachman and Phillips (1996) estimated that 40% of the total nitrogen load of Chesapeake Bay reflects groundwater discharge. The data on groundwater loadings of nitrogen from other areas in North America are limited, but this source is clearly potentially significant in areas with extensive marshes that are cut by numerous tidal creeks (Seagle et al. 1999).

Direct input of nitrogen from the atmosphere has been demonstrated over the last two decades to be a major source of nitrogen to North American coastal waters (Paerl 1985), as it is globally averaging 10–25% of the global atmospheric nitrogen input (Duce 1991). Studies of atmospheric nitrogen loadings in the middle Atlantic coast show that wet deposition accounts for the atmospheric nitrogen, with nitrate being the most common form, though organic nitrogen can be important seasonally (Seagle et al. 1999). Dry deposition, though accounting for less than half the atmospheric loading of nitrogen, nevertheless may yield as much $1.5\text{--}4.1\text{ kgN ha}^{-1}\text{ yr}^{-1}$ in the Chesapeake Bay region (Gardner et al. 1996).

Overall, as Seagle et al. (1999) point out, though figures for atmospheric deposition of nitrogen like those for

groundwater loadings are limited, the phenomenon reinforces the importance of land use in not only controlling runoff and ground-water nutrient inputs into North American coastal waters, but the impact of land use in controlling the fate of atmospheric loadings. In particular, knowing the spatial distribution of upland and riparian forests/shrublands as nutrient buffers clearly holds the key to managing the effects of nutrients on North American Coasts. In this respect, databases like US National Oceanic and Atmospheric Administration's (NOAA), Coastal Change Analysis Program (CCAP; Klemas et al. 1993), which portray changes in land cover and land use, comprise essential tools for wise coastal management.

Biological Interactions

The populations of marine and brackish water organisms that characterize the coastal waters of North America constitute the fundamental components of a variety of ecosystems that range from the mangroves of south Florida to the high-latitude wetlands of Hudson Bay (Fig. 3). The sheer variety of organisms, which vary broadly by latitude, and locally by salinity and season, is clearly beyond the scope of this entry. It is perhaps best to focus on examples of how these organisms comprise basic elements of North American coastal ecosystems, interact at various trophic levels, and influence overall ecosystem functioning.

Phytoplankton

Phytoplankton (diatoms, coccoliths, and dinoflagellates) populations in North American coastal waters, as along all coasts, depend on salinity, nutrient availability, and

illumination. In estuaries, salinity plays an especially important role in the presence of species ranging from freshwater to brackish and fully marine species. Seasonal variations in salinity, particularly those associated with large spring runoff, may allow freshwater and lower salinity brackish species to migrate considerable distances down estuary. This down-estuary displacement, in highly stratified systems (e.g., salt-wedge systems), may ultimately prove disadvantageous, as internal wave mixing along salinity boundaries, may expose freshwater species in particular to osmotic stress. Euryhaline species are less likely to suffer such problems.

Nutrient influences on phytoplankton abundance reflect the similar riverine influxes that control nutrient dynamics. In Chesapeake Bay, low discharges of the Susquehanna River – the defining river of the estuary – and the other major tributaries, like the Potomac and James Rivers, in May through December probably underlie the nutrient limitations on phytoplankton productivity (Carpenter et al. 1969). However, there is likely a greater year-round nutrient demand in most Atlantic Coast estuaries than inputs from external sources (absent significant sewage outfalls), and *in situ* processes in the estuaries must be supplying most of the remaining nutrient demand. These processes have been identified as: (1) remineralized nutrients excreted from herbivorous zooplankton and benthic organisms; (2) releases from sediment resuspension and interstitial sediment water; and (3) exchanges from nutrients in particulate forms and dissolved phases (Smayda 1983).

In coastal lagoons and bays, which lack significant riverine inputs that produce changes in salinity and nutrient concentrations, light intensity in the euphotic zone above the compensation depth (where a balance occurs between rates of phytoplankton photosynthesis and respiration) is probably the major controlling factor. Resuspension of subtidal sediments is the principal mechanism by which turbidity and light attenuation occurs in lagoons, particularly in the open water lagoons that characterize the large, long thin barrier islands like Assateague Island and Atlantic City, New Jersey. Large-scale resuspension in these lagoons – many of which are quite shallow-occurs largely during storms. However, depending on tidal energies (and inlet tidal prisms) appreciable concentrations of subtidal materials may be entrained from the flanks of inlets and in the vicinity of shoals forming flood tidal deltas. Most of these materials probably settle out of suspension within a short distance of the inlets as tidal velocities rapidly slow. But in a lagoon with a large number of inlets, there could be significant diurnal variations in turbidity and illumination affecting phytoplankton. In the marshy lagoons behind drumstick barriers, outwelling of particulates from marshes during ebb is probably an additional source of suspended organic and mineral materials.

Lastly, though salinities in many coastal lagoons and bays do not vary as strongly seasonally as in estuaries, in areas

away from inlets that may be blocked from ready tidal exchange by marshes, hypersaline conditions may prevail in late summer as evaporation rates increase, with concomitant effects on phytoplankton populations.

In estuaries of North America, plankton metabolism serves an important regulator of estuarine chemistry. Plankton photosynthesis is correlated with decreases in the concentration of carbon dioxide, ammonium, phosphate, nitrate, and silicate (Wolff 1980). Moreover, oxygenation of local waters increases. In Long Island Sound, annual primary phytoplankton productivity has been determined to be as much as $380 \text{ gCm}^{-2} \text{ yr}^{-1}$, which is similar to the value of $308 \text{ gCm}^{-2} \text{ yr}^{-1}$ determined for Narragansett Bay (Wolff 1980; Smayda 1983).

Benthic Algae and Macrophytes

The decline of sea grasses and parallel spread of benthic algae estuaries and coastal lagoons of North America have focused concern on the widespread phenomenon of coastal eutrophication. In coastal systems where high anthropogenic inputs of nutrients, coverage of benthic algae can be extensive, ranging from single cell species (like diatoms) to complex, multicellular species, forming mats. Studies of primary production of benthic in estuaries where nutrient concentrations are not yet excessive show comparatively low values, especially compared with phytoplankton (Wolff 1980). Substrate characteristics have been suggested to be perhaps the most important limitation on benthic algae productivity. Single cell species prefer soft substrates like muds, whereas the complex multicellular species prefer hard substrates like shell debris, rocks, or shore construction, such as pilings or bulkheads.

The importance of benthic algae to the ecology of North America coasts has been underestimated, because often this component is measured as part of another system as in salt marshes where its productivity (but not biomass) sometimes rivals macrophytes. No such ambiguity relates to the role played by sea grasses, whose productivities exceed even phytoplankton. Sea grass biomass can reach $500 \text{ gCm}^{-2} \text{ yr}^{-1}$ (Wolff 1980) and Barsdate et al. (1974) indicate that 90% of production in a lagoon in Alaska is due to *Zostera*. The high productivity of sea grasses ensures that they serve as a major food sources for organisms ranging from gastropods to fish, though direct consumption is probably limited and most of their importance to food chains appears to be as detritus (Barsdate et al. 1974; Stevenson 1988).

As plants rooted in the sea bottom, sea grasses cycle a number of compounds from subtidal sediments into ambient waters through their leaves and by decomposition after death. Phosphate, in particular, may be transported up roots and through leaves of *Zostera maritima* at a rate of $0.066 \text{ gm}^{-2} \text{ day}^{-1}$ (McRoy et al. 1972).

Salinity, in lieu of anthropogenic influences, is the major control of the species distribution of sea grasses along North

American coasts. *Zostera* and *Thalassia* can occur where salinities range through 35 ppt, whereas *Ruppia* and *Zanichellia* characterize brackish waters (Wolff 1980).

Coastal Wetlands: Marshes and Mangroves

The marshes – and the mangroves in south Florida – that characterize the North America coast occur in a variety of physiographic/depositional settings, but largely in areas protected from strong wave energies like lagoons and estuaries. In the United States southeast Atlantic Coast and Gulf Coast, coastal wetlands comprise a substantial part of the coastline, and individual marsh systems can easily range in size up to several thousand hectares or more. The contribution of large coastal wetlands to the ecology of coasts is clearly inescapable, but even the much smaller and more isolated marshes of New England up through the Canadian Maritimes are important to local ecosystems. Woodwell et al. (1979) demonstrated that significant interactions occurred between a salt marsh in Long Island Sound and local coastal waters.

Essentially, coastal wetlands serve several roles in North America coastal ecosystems: (1) as important components of several littoral biogeochemical cycles, especially nutrients and carbon; (2) as critical habitats for species ranging from invertebrates to birds; and (3) as sinks for mineral sediments. To these, may be their additional functions in coastline stabilization and erosion control, though rapidly eroding coastlines are seldom areas where marshes persist.

Coastal marshes are generally considered to be net sinks for inorganic nutrients and net exporters of organic materials (including associated nutrients) (Mitsch and Gosselink 1995). Studies of nutrient exchanges in US Atlantic Coast marshes (e.g., Stevenson et al. 1977) leave the general question of whether marshes are net exporters or importer of nutrients unsettled. Nixon (1980) concluded, in his review of available evidence at the time, that marshes were, overall, exporters of dissolved organic nitrogen (DON) and dissolved phosphorus, but importers of nitrite and nitrate. Nonetheless, it is clear that high levels of denitrification can occur in marshes. Gross denitrification at Great Sippewissett Marsh in Massachusetts, particularly from June through October, was almost twice the import of nitrate into the marsh from groundwater (Kaplan et al. 1979).

The high net primary productivity has been cited as underscoring the importance of coastal wetlands as a source of organic detritus and dissolved organic carbon (DOC) in North American estuaries is clear. In eroding marshes the amount of organic detritus per hectare of marsh can be very large, particularly in brackish or estuarine marshes where the percentage of organic carbon by dry weight in marsh sediments is generally 50% (Stevenson et al. 1985). However, whether all of these materials eventually serve as food sources in estuaries is a matter of debate. The concept that “outwelling” of organic materials from marshes buffers

nutrient demand in estuaries has been suggested to be one of the major functions of tidal marshes in estuarine productivity (Odum 1961). Actual support for this hypothesis has proven inconclusive. Haines (1979), using carbon isotope data, showed that carbon from strictly terrestrial plants dominated the tissues of fish and other herbivorous species, suggesting that whatever the quantity of carbon exported from marshes, it was not preferred as a primary food source.

Similar problems surround the functions of marsh as sediment sinks. While it is clear from sedimentological studies (Stevenson et al. 1986) that marshes – and by extension, mangroves – comprise great stores of mineral sediment, the actual functions of coastal wetlands in littoral sediment budgets of North American coasts is not resolved. In Chesapeake Bay, Stevenson et al. (1988) suggested that suspended sediment deposition in marshes accounts for 11% of the total estuary sediment budget. Much of the effectiveness of coastal marshes lies in how tidal velocities vary between flood and ebb tides, with strong ebb domination of the tidal cycle promoting greater export of mineral than import. The few data on sediment fluxes in North American marshes implies that their functions as major sediment sinks are probably limited, with many marshes probably exporting more sediment than importing sediment during the tidal cycle (Stevenson et al. 1988).

The case for providing critical habitat, especially for water birds and commercially valuable species of finfish and shellfish, is much stronger. It has been estimated that 90% of commercially valuable finfish and shellfish spawn and spend their juvenile stages in coastal marshes (Mitsch and Gosselink 1995). More directly important to coastal ecology is the contribution of coastal marshes to sustaining a host of organisms in the marsh food web, particularly larval stages of filter feeding organisms like the oyster (*Crassostrea virginica*), which as adults, contribute significantly to maintenance of estuarine water quality.

Animals

The variety of animals that characterize the coastal waters of North America include organisms that spend their lives ranging more or less freely in the water column like zooplankton and complex vertebrates, organisms that largely inhabit the benthos (mainly invertebrates among the metazoans and many unicellular species like various bacteria), and those that move from the benthos to water column, often depending on life stage. These animals, whether herbivorous or predatory, are often part of complex trophic interactions in littoral food webs, above all in estuaries. The occurrence of both pelagic and benthic species is related to turbidity, salinity, dissolved oxygen, and availability of food sources. The last, not surprisingly, is exemplified by the fact that zooplankton are often found in greatest abundance in North American coastal waters (as elsewhere) where numbers of

phytoplankton, their principal food source, are highest (Miller 1983). However, in the general ecological functioning of North American coastal systems, the role of animals can be seen to relate to aspects of coastal biogeochemical cycles: (1) remineralization of nutrients from grazing on phytoplankton; (2) mixing and resuspension of subtidal sediments, affecting turbidity and reintroduction of biogenic debris, often fecal material; and (3) the filtering of largely organic particulates by filter feeders like oysters and clams.

Grazing on phytoplankton and organic detritus, with the ultimate release of remineralized nutrients in the excreta from death of animals, is a major element of nitrogen and phosphorus cycles in estuaries. Though filter-feeding benthic invertebrates clearly play a part in this, the major regenerators of nutrients are zooplankton. Ingestion of detritus as a source of remineralized nutrients by copepods appears to be seasonally important along the US Atlantic Coast, with the months between March and May being the time of year when algal food sources are insufficient (Heinle and Flemer 1976).

Mixing and resuspension of subtidal sediments, particularly organic particles, mainly reflects the activities of burrowing invertebrates. Bioturbation of subtidal sediments in estuaries not only homogenizes sediments down to depths of often a meter (*Mytilus*), but increases the potential for resuspension of sediments, especially fecal pellets. In Atlantic Coast estuaries like Chesapeake Bay, the annual amount of fecal material excreted by oysters can be as high as 1–2 metric tons per hectare (Nichols and Biggs 1985). The potential resuspension of these materials, as well as other organic particulates, for later heterotrophy or bacterial attack by bacteria is a significant source of inorganic nitrogen (e.g., NO_3^-) and orthophosphates (Aston 1980) in ambient waters.

The decline in the populations of filter-feeding organisms like the oyster has been cited as one of the major reasons for the crises in water quality along North America coasts. In Chesapeake Bay, the population of oysters at the time of initial European contact in 1608 was probably sufficient to filter all the Bay's waters in 2–3 weeks, whereas by the late twentieth century it needed several decades (Newell 1988). This considerable contribution to the littoral waters of North American coasts should not diminish the importance of the removal of remineralized nitrogen passing through the guts of oysters as fecal material which binds the sediment, and makes it less vulnerable to resuspension.

Threats to the Ecology of North American Coasts

Eutrophication

Coastal eutrophication in North America during the latter half of the twentieth century has yielded poorer water quality, diminishing catches of commercially valuable species of finfish and shellfish, loss of habitat for other creatures, and lower

recreational values. A recent report by the National Oceans Service of the US NOAA found that 65% of the total estuarine surface area could be considered as exhibiting moderate to high levels of eutrophic conditions, marked by depleted dissolved oxygen levels, loss of submerged aquatic vegetation, growth of macroalgae, increasing frequency of toxic algal blooms, and increasing levels of chlorophyll *a* (Bricker et al. 1999). The middle Atlantic and Gulf Coasts, in particular, represent the most threatened coastal waters in North America.

Not the least of the many impacts of coastal eutrophication, are the mounting costs for saving what remains of rapidly vanishing coastal resources. The importance of the cost issue can be indicated by estimates of expenditures for controlling nutrients in the largest estuary in North America, Chesapeake Bay. Between 1985 and 1996, over US \$3.5 billion were spent on nutrient control in the Bay's watershed (Butt and Brown 2000). The cost of removing each kilogram of total nitrogen per year in the Chesapeake Bay alone probably ranges upwards of US \$35 (Camacho 1992). This cost, of course, does not reflect the large expenditures for the restoration of resources damaged by coastal eutrophication, encompassing funds for the monitoring and replanting of sea grasses, grants for improving the aquaculture of threatened species like the American oyster (*C. virginica*) in several Atlantic Coast estuaries, subsidies for the funding of alternative farming methods, and the like. But the dimensions of the threat posed by the eutrophication of North American coastal waters, epitomized by the rapidly enlarging "dead zone" in the Gulf of Mexico at the mouth of the Mississippi River (Turner and Rabalais 1994), are such that delaying action may prove to incur costs well beyond the money needed for remediation.

Burgeoning human activities around coasts and in coastal watersheds underlie the phenomenon of coastal eutrophication. A major impediment to effective mitigation of the threat has been readily accessible, up-to-date information on the nature and extent of land cover changes and land use practices that contribute to loading of nutrients in coastal waters. The pace of change is such that growth of populations (regardless of the types of activities these people will be engaged in) could increase by 100% in many already heavily settled areas of the US Atlantic Coast (see Stevenson and Kearney 1996).

Monitoring changes in land cover and land use is crucial to any assessment of the effects of human activities on coastal ecosystems. Satellite remote sensing has been particularly useful in this regard, and the C-CAP program in the United States (Klema et al. 1993) is a good example. This program produced the first, uniform information on land use and land cover changes in the US coastal zone using Landsat Thematic Mapper imagery. Application of remote sensing technologies in combination with Geographic Information Systems (GIS) and various spatial environmental models (e.g., Costanza

et al. 1990) are providing the broad synoptic tools for regional to coast-wide management of ecological impacts that attend population growth and development along North American coasts.

Providing information is not necessarily mitigation of ecological threats and damage, which require political action. Here, the progress has been more piecemeal, particularly in the United States with its combination of federal, state, and local jurisdictions, and overlapping and differing mandates between state and federal agencies. General water quality mandates, regulated by the US Environmental Protection Agency (EPA), are related to issues of direct or indirect risk to human health and seldom effectively address issues like coastal eutrophication that are usually loosely related to any specific human activity and tend to be regional to extra-regional in scope. Regional consortia of local and regional governments are perhaps best positioned to assume the various roles necessary for advocating (especially with local municipalities), enacting, and coordinating the measures for coastal mitigation. The Tri-State Commission – Maryland, Virginia, and Pennsylvania – for cleanup of the Chesapeake Bay is a good example of jurisdictional cooperation in a major effort to limit the influx of nutrients in a major coastal system of North America already burdened by extensive coastal eutrophication.

Land use policies aimed at controlling nutrient inputs span those advocating changes in regional agricultural practices to legislated controls on the types of land use permitted in critical areas of large watersheds feeding coastal systems. Controls on land use or density of development are not always popular, but are effective even where employed in only a limited fashion. In the 1980s, the State of Maryland enacted the Critical Areas Act mandating the creation of 1,000-ft (330 m) buffer zones around the Chesapeake Bay and its major tributaries. This buffer was largely intended to protect from encroaching development, upland forests and riparian zones that can attenuate nutrient runoff into coastal waters. In addition, the density of development in certain areas of the Bay coastline was further controlled, again to limit nutrient influx. To be sure, the Act met with mixed responses from various quarters, but subsequent studies did show that nutrient inputs were significantly diminished to the Bay (Marcus et al. 1993).

Sea-Level Rise

The possibility of rapidly rising sea levels from global warming first became widely recognized almost 20 years ago, and was followed over the next two decades by numerous studies outlining what an acceleration in sea-level rise could mean for coastal systems. Though an accelerating sea-level rise could have a multitude of effects (both direct and indirect) on North American coastal ecosystems, at least two are likely: the potential extensive loss of coastal wetlands and

increasing turbidity in Atlantic Coastal Plain estuaries from higher rates of shore erosion.

Estimates derived from historical surveys suggest that almost half of the coastal wetlands in the United States extant in 1900 had disappeared by the last quarter of the twentieth century (Gosselink and Baumann 1980). Many of these losses reflect development or other human activities, but overall it is safe to assume that a more generic cause, sea-level rise, was responsible. In recent years, regional studies ranging from the Mississippi Delta to the middle Atlantic Coast (cf. Stevenson et al. 1986) indicate that marsh vertical accretion rates in many areas lag local rates of submergence. In areas where there are significant deficits in vertical accretion compared to sea-level rise, inventories using aerial photography and other techniques show high wetland losses. In the Mississippi Delta, annual losses of coastal marshes amount to almost 60 km² (Britsch and Kemp 1990).

Projections for rates of global sea-level rise approaching 1 m per century would be catastrophic for the survival of coastal wetlands generally, in particular coastal marshes (Nicholls et al. 1999). Admittedly, most research on the relations of coastal wetlands to sea-level rise has focused on middle latitude coastal marshes, but it is probably safe to assume high-latitude marshes in Hudson's Bay would fare the same. In addition, though the imminent widespread collapse of mangroves due to sea-level rise heralded almost a decade ago (cf. Ellison 1993) was perhaps premature, the survival of mangroves in south Florida in such a scenario would be doubtful as well.

Despite some continuing disagreement concerning the overall ecological role of wetlands in estuaries and lagoons along North America's coasts, reviews by Nixon (1980), among others, make clear that loss of coastal wetlands would affect aspects of coastal ecology as various as nutrient cycling and fisheries. The economic costs of large-scale wetland decline and loss would be equally significant. Costanza et al. (1989) demonstrated that the size of shrimp harvests in Louisiana correlated well with marsh area in the Mississippi Delta.

The role of sea-level rise as the driver of long-term coastal erosion is well known. Nevertheless, the impact that an accelerated rate of sea-level rise would have on water turbidity in North American estuaries is just becoming evident. Detailed sediment budgets for many North America estuaries are generally not available, but it is clear that sediment inputs from shore erosion will increasingly loom larger as rates of sea-level rise increase. At present, shore erosion adds a volume of sediment into Chesapeake Bay every century equivalent to the District of Columbia (US Army Corps of Engineers 1990). Much of the finer-grained eroded sediments will inevitably end up increasing the already high, suspended particulate load of estuaries, thereby further increasing water turbidity. With effects of high turbidity on estuarine biota

that span damage to sea grasses to even the foraging success of predatory fish species like striped bass, mounting inputs of suspended sediments may offset any hard-won gains in water clarity from controls on upland erosion that have been emplaced in many areas in recent decades.

Baja California: A Unique North American Coast

In 1940, a year after the publication of *The Grapes of Wrath*, John Steinbeck with his friend, Edward Ricketts, spent 5 weeks exploring the Sea of Cortez. Exhausted and troubled by the controversy generated by his great novel, Steinbeck, connecting with his early undergraduate training as a marine biologist, surely could not have chosen a more unique coastal system in North America to visit. Surrounded by deserts, the Sea of Cortez contains some of the deepest basins (over 2,000 m) of any coastal region of the continent. About 4.5 million years old (Atwater 1970), and formed by separation of the Baja Peninsula during the late Pliocene and Pleistocene, the area is seismically active, and characterized by occasional tsunamis, with volcanism occurring as recently as the Holocene (Gastil et al. 1983). One of the most salient features of the Gulf, is its many islands (30–40) that fall into two general groups, northern and southern islands. The islands differ in origin, some being volcanic, others formed by uplift or submergence, and those next to the mainland created by narrow headlands separated by erosion (Gastil et al. 1983).

The upper Gulf is an area characterized by some of the highest tides in the world, with spring tidal ranges in the vicinity of the Colorado River Delta reaching almost 10 m (Matthews 1968). Tidal phase also changes from the delta to the middle Gulf, and though the tidal wave is modified by islands and, in particular, by shoals in the vicinity of the delta (Maluf 1983). However, tidal phases for most of the Gulf appear to be largely semidiurnal in character, although they are regularly diurnal around Bahía Concepción and Guaymas during the month (Maluf 1983).

The ecology of the Sea of Cortez is dominated by the seasonal climate variations and influence of the islands. Because the sea is too small to modify regional climates, water temperatures mirror seasonal changes in regional temperature, especially in the upper Gulf (Maluf 1983). Together with the sometimes extreme range in salinities near the Colorado Delta, produced by changes in discharge of the Colorado, shallow-water biota show a greater interannual variations in composition (Maluf 1983). The influence of the islands is reflected in bird, reptile, and mammal populations that change by location of the islands and their proximity to mainland coasts. Slight terrestrial faunal differences between the Baja Peninsula and the mainland Mexican Coast tend to characterize islands nearest either shore. Phytoplankton densities decline around islands clustered closest to

mainland shores, where water turbidity is higher due to shore erosion and wave resuspension during storms (Thomson and Gilligan 1983). Because water masses in the Sea of Cortez can be broadly partitioned into either being mainland-influenced or Pacific-influenced, there is said to be a pelagic species gradient within the upper and lower Gulf, but the actual gradient may run between the mainland shore and off-shore islands, reflecting the greater seasonal variation in sea surface temperature along mainland coasts (Thomson and Gilligan 1983). Changes in benthic species also show mainland versus island differences, in this case largely due to the reefs that ring many of the islands, especially in the lower Gulf.

In the lower Gulf, a large, stable low oxygen zone, produced by decomposing phytoplankton, limits benthic species where it impinges upon shallower water areas (Maluf 1983). At the mouth of the Gulf, ocean circulation disrupts, reinforcing the essentially confined-sea character of the Gulf of California.

Cross-References

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North America, Coastal Geomorphology

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Definition

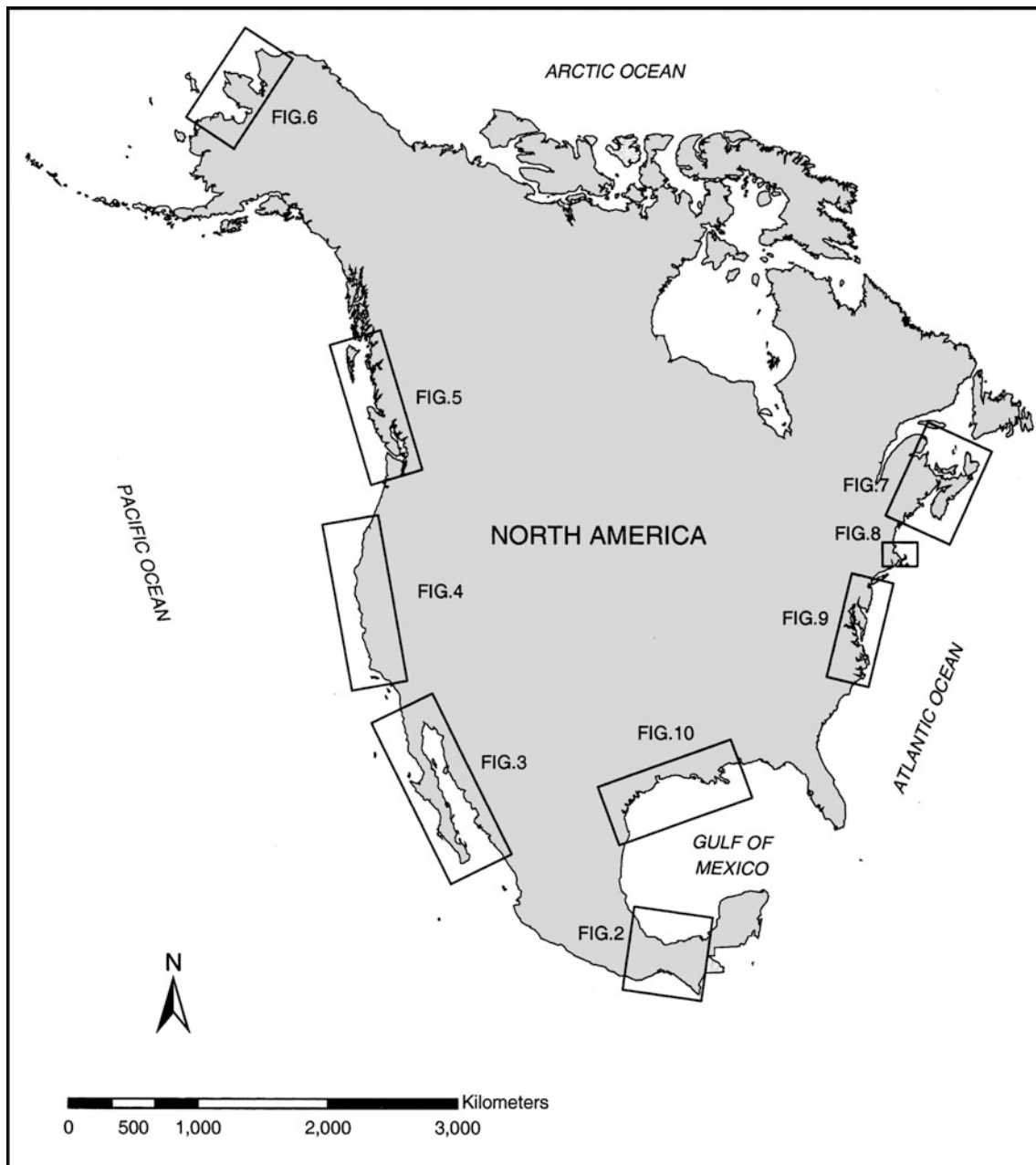
The shores of the North American continent span approximately 55° of latitude and 100° of longitude. The continent is bounded by four major coastal regimes associated with the Atlantic, Arctic, and Pacific Oceans, and the Gulf of Mexico. The length of open ocean coast, excluding nonbarrier islands, exceeds 110,000 km and is much longer if the Great Lakes, embayments, *lagoons*, and islands are included. The coastal morphology includes all of the major coastal types described in Shepard's (1937) classification. There are *mangrove* and coral systems in the southern reaches of the continent, and *ice*-dominated systems in the north. The common occurrence of fjords and rias attests to the delayed response of most of the North American coast to glacial processes and *eustatic* and *isostatic sea-level* changes.

Geomorphic Provinces

The coast of North America (Fig. 1) comprises parts of at least seven of the geomorphic provinces described in Graf (1987). The designations for these provinces, and their characteristics, are taken from Graf (1987), unless otherwise noted, and they are used to organize and introduce the principal features of coastal regimes discussed herein. The North American distribution of *muddy coasts* and coarse clastic coasts are presented separately.

Pacific Rim Province

The west coast of the continent, from the Mexico–Guatemala border to the Alaska Peninsula, is included in the Pacific Rim Province. It is a region that is tectonically active, lying along the convergent boundaries of the Pacific, Juan de Fuca, and Cocos Plates with the North American Plate, although the Gulf of California is a spreading center. Most of this coast, other than Baja California and Southern California, is a collision coast using the terminology of Inman and Nordstrom (1971). The general morphology of the province is characterized by a narrow continental shelf and narrow or absent coastal plains backed immediately by steep and high mountain ranges. Coastal *cliffs* are common and estuarine or

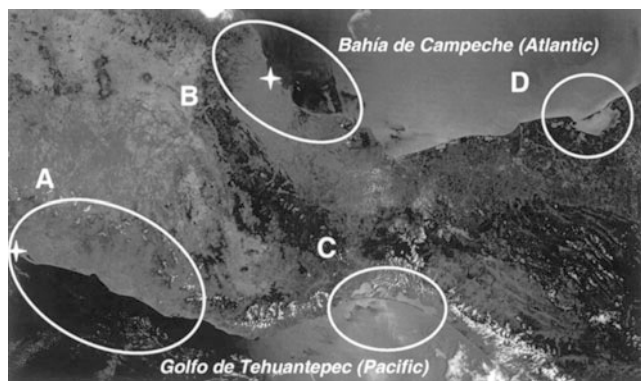


North America, Coastal Geomorphology, Fig. 1 An outline of the North American continent. The insets mark regions represented in Figs. 2, 3, 4, 5, 6, 7, 8, 9, and 10

barrier systems are limited to the vicinity of drowned river mouths. Notable exceptions are the barrier systems along the eastern shore of the Gulf of California. Spring *tide* ranges vary from about 0.5 m along the Mexican coast near Acapulco to 1.0–3.0 m along most of the rest of the Pacific coast to the United States–Canada border. Ranges increase northward to more than 10 m in the vicinity of Anchorage, Alaska. Spring tide ranges reach 7 m in the northern reaches of the Gulf of California (Kellett 1995). The dominant source of wave energy along most of this coast is from the ocean swell,

although tropical *storm*-induced waves are important in Mexico and (occasionally) Southern California.

The Pacific coast of Mexico in the vicinity of Guatemala (Fig. 2) is characterized by barrier/lagoon systems. *Dunes* are common on the barriers, and mangroves are present in many of the lagoons. *Coral reefs* occur sporadically along this coastline. Much of the coast north of the Gulf of Tehuantepec to Nayarit is cliffed where the Sierra Madre del Sur lies adjacent the coast. Along other reaches of this region, barrier/lagoon systems are extensive, especially east of Acapulco

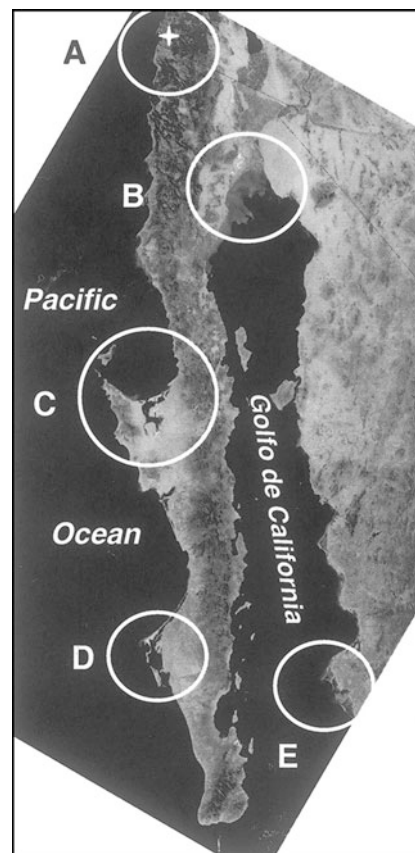


North America, Coastal Geomorphology, Fig. 2 The Isthmus of Tehuantepec in tropical Mexico. This is a MODIS image (courtesy of NASA) from 29/03/02. (A) Sandy coastline extending east from Acapulco (↔) with barrier and lagoon systems. (B) Lowland coast in the vicinity of Veracruz (↔) includes sandy barrier systems and extensive marshes. A coral reef system is located just offshore from Veracruz. (C) The extensive barrier/lagoon/marsh systems of the Gulf of Tehuantepec. (D) The Laguna de Términos is about 50 km wide, fronted by a sandy barrier and backed by extensive marshes

(Fig. 2), in the vicinity of Zihuatanejo, and between Puerto Vallarta and Mazatlan.

There is another barrier/lagoon complex east of Los Mochis (Fig. 3). The *sandy* barriers include coastal dune and *beach ridge* complexes. Many of the lagoons provide mangrove habitat (Kelletat 1995). There are similar systems north from Los Mochis, but they become smaller and occur less frequently along the increasingly arid coastline of the Gulf of California. Wave energy decreases northward into the Gulf, but the tide range increases to a maximum at the mouth of the Colorado River (Fig. 3). The Colorado River *delta* comprises a large network of distributaries, estuaries, marshes, and saline lagunas. Many of these coastal environments are threatened by reduced fluvial discharge and sediment supply caused by large *dams* on the river (Carriquiry and Sánchez 1999). The location of the delta is controlled by the San Andreas *Fault* that separates the North American plate on the east, from the Pacific plate, with Baja California, on the west.

The east coast of Baja California (Fig. 3) is shaped largely by recent *tectonic* activity, with the Sierra de San Pedro Martir and the Sierra de la Giganta (both part of the Peninsular Range that runs north to Los Angeles) abutting the coast in the northern and southern Gulf, respectively. The coast is irregular and frequently cliffed with pocket beaches between headlands. The arid climate produces few perennial streams to deliver sediment to the shore, and the entire coast is low energy, so beaches tend to be relatively narrow. There are only a few, small barrier systems, at Santa Rosalia, for example, or La Paz. Baja California's west coast is also quite rugged. However, drainage basins are larger than those found along the Gulf coast, and sediment supplies are larger.



North America, Coastal Geomorphology, Fig. 3 Baja California Peninsula (Mexico) and the Gulf of California. This is a MODIS image (courtesy of NASA) from 11/11/02. (A) San Diego, California, USA (↔). Note the small size of the bay relative to those at locations C and D in Mexico. (B) The delta of the Colorado River and the river's suspended sediment plume. (C) Laguna Ojo de Liebre and barrier systems in Vizcaino Bay. (D) Bahia Magdalena and its extensive barrier systems. (E) Los Mochis and the northern edges of the barrier/lagoon/marsh systems of Sinaloa states

The most prominent coastal features are Bahia Magdalena and Vizcaino Bay. The former is the largest of a series of bays created by the presence of a chain of *barrier islands* and *spits*. Coastal dunes are common on the barriers and the lagoons are backed by marshes or sand and salt flats. In Vizcaino Bay, Laguna Ojo de Liebre (Scammon's Lagoon) results from structural control on coastal development. Sierra Vizcaíno parallels the Sierra de la Giganta to create a large trough that has been filled with sediments from the mountain ranges. The barrier and lagoon systems are formed in these sediments along the bay. Extensive dune fields and salt flats occur across this complex. Raised marine terraces are found at elevations up to 150 m (Orme 1998).

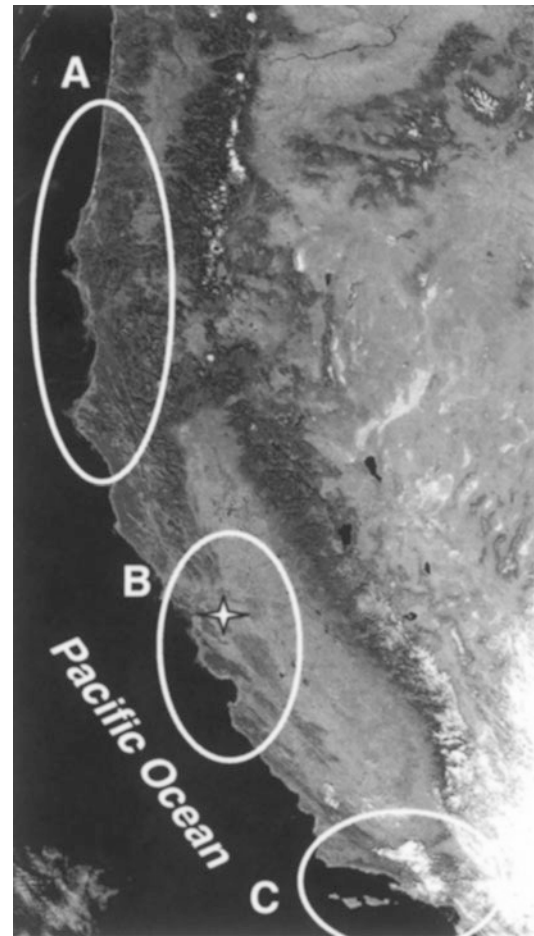
Near the United States–Mexico border, the Tijuana River created a substantial *estuary* marsh system and has provided the sediments that were transported northward to form the barrier spit that protects San Diego Bay from the open ocean

(Fig. 3). Most of the coast in this region has cliffs fronted by narrow sand or cobble beaches. This configuration, often backed by sets of marine terraces, is common to the rest of the Pacific coast, except where structural controls have formed basins or where rivers have formed estuaries. The city of Los Angeles, for example, is built across the structurally controlled Los Angeles Basin that has been filled with thick deposits (locally in excess of 6,000 m) of marine sediments. It is the largest coastal plain in California, not coincidentally occupied by the largest city. Los Angeles beaches are backed by *Holocene* (and older) dunes, although most dune surfaces have been urbanized. The Southern California coastline, with the exception of the military reservation at Camp Pendleton in San Diego County, is also extensively human altered through beach nourishment, cliff stabilization efforts, and seawalls and groin fields (Griggs 1998), and the extent of alteration rivals that of the New Jersey shore.

The Southern California coastline is also characterized by well-defined sets of raised marine terraces. The Palos Verdes Peninsula divides the coast of the Los Angeles Basin. There are 13 terraces on the peninsula, at elevations up to 411 m, the marine limit (Orme 1998). The marine limit rises to about 600 m, with 15 terraces, in the vicinity of Santa Barbara, and falls to about 200 m near the Big Sur Coast of Central California. The Big Sur coastline (Fig. 4) is formed of high marine cliffs carved into the Santa Lucia Mountains. Pocket beaches, of sand or coarser materials, occur in sheltered locations along this coast. Monterey Bay (Fig. 4) marks the separation of the Santa Lucia and Santa Cruz Ranges. There is a coastal, deltaic plain created by the Salinas River flowing between the mountain ranges. Wave energies along this coast are high, and most of the coastline is backed by dune complexes that extend inland, in locations, almost 10 km (Cooper 1967). Coastal cliffs and low marine terraces back the coast from Santa Cruz, at the northern edge of Monterey Bay, north to San Francisco Bay. Noteworthy is the Devil's Slide coastline south of San Francisco, where steep, 250-m-high cliffs have been eroded into the unstable Franciscan formation and where, as the name implies, slope failures are common.

San Francisco Bay (Fig. 4) is formed by the submergence of a Pliocene basin. The Sacramento River system flows into the bay, forming an inland delta and marsh complex. The distributaries are confined between levees and much of the delta surface is currently below sea level. During Pleistocene lowstands of sea level, the river debouched to the Pacific through the channel it had eroded between the Golden Gates. North of San Francisco is the Point Reyes peninsula, an outcrop of the Pacific Plate separated from the mainland by Tomales Bay, formed in a trough along the San Andreas fault line.

The California coast north to Cape Mendocino is extremely rugged, comprising high cliffs cut into the Coastal Range Mountains, and localized sand or gravel beaches.



North America, Coastal Geomorphology, Fig. 4 The coasts of Oregon and California. This is a SeaWiFS image (courtesy of NASA) from 23/10/00. (A) The relatively straight coast of Southern Oregon includes large barrier spit/estuary systems and is backed by extensive coastal dunes. The major promontories are Cape Blanco in the north and Cape Mendocino in California. (B) The Central California coast from San Francisco (✧) to the Big Sur Coast. San Francisco and the Bay are immediately west of the ✧. The large embayment to the south is Monterey Bay. (C) The west-east trending coast of Southern California, including the Channel Islands. This coast is characterized by a series of small barrier/estuary systems separated by headlands and cliffs

Numerous submarine canyons extend close to the shore. There are no large drainage systems reaching this coast and most of the reach remains inaccessible by land. North of the Cape, several large rivers, including the Eel, Mad, Klamath, and Smith, do reach the coast to form estuary and baymouth barrier systems such as the one that forms Humboldt Bay. Large coastal dune complexes have formed from the sediments blown off the beaches near these rivers.

The relatively straight coastlines of Oregon and Washington (Fig. 4) lie adjacent to the subduction zones associated with the Juan de Fuca and Gorda Plates. As is common with collision coasts, rugged terrain and cliffs are characteristic along these shores. There are, however, numerous fluvial/estuarine systems with baymouth barrier spits. Good examples

of these occur at Coos Bay, Newport, and Tillamook Bay in Oregon, and Willapa Bay, and Grays Harbor, Washington. There are also extensive dune systems along this coast (Cooper 1958). Many of the dunes are confined to the barriers, but at several locations, such as near Florence, or Sand Lake, Oregon, they extend inland for kilometers. The Pacific coasts of Canada and Alaska (Fig. 5) show the marked influences of alpine glaciation. The coast of British Columbia shows a strong structural control on morphology, with the Georgia and Hecate depressions separating Vancouver Island and the Queen Charlotte Islands from the coastal mountain ranges on the mainland.

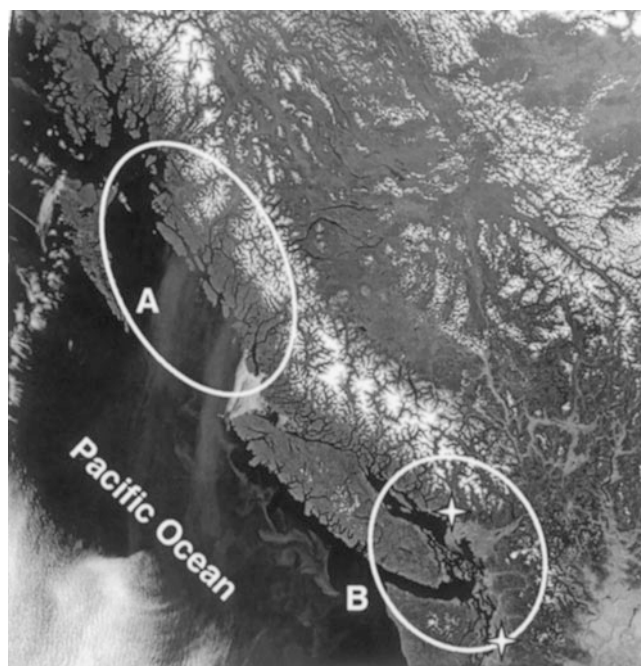
The coast is deeply incised as a result of glacial erosion and the creation of numerous fjords. There are very few beaches along this entire coast, and they tend to be confined to the headwaters of the fjords where rivers deposit sands and gravels. There is a large delta system at the mouth of the Fraser River (Fig. 5), creating a large coastal plain that is currently occupied by the city of Vancouver and its suburbs. Most of this coast, continuing into Alaska, is inaccessible by land.

The Pacific coast of Alaska (Fig. 5) is similar to that of British Columbia, with three important distinctions. First, there is greater tectonic activity, especially along the Aleutian Megathrust, and this has spawned a chain of active *volcanoes*

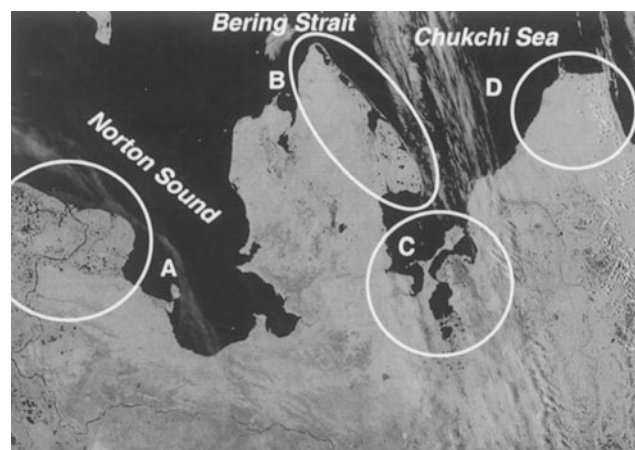
that have created the Aleutian Islands. Second, many of the Alaskan fjords retain tide-water glaciers that preclude any beach formation. Third, tidal ranges increase dramatically toward Anchorage, and this produces swift tidal *currents* that transport much of the fine sediment load from the coastal streams into deepwater. Most of this coastline is accessible only by water or air. About midway between Sitka and Cordova, Alaska, sits Lituya Bay, site of the largest “recorded” wave uprush or swash (just over 500 m) caused by an earthquake and resulting landslide, in 1958.

The Arctic Coastline

The Arctic coasts of North America include portions of four geomorphic provinces: the Interior Mountains and Plateaus, the Rocky Mountains, the Arctic Lowlands, and the Canadian Shield. Overarching characteristics of these coastal regimes are the effects of extensive sea ice, the effects of continental glaciation and isostatic adjustments, and the effects of permafrost. In addition to generating a suite of ice-related coastal landforms, such as ice foot, push ridges, thermokarst lagoons (Fig. 6), or scour features, the presence of sea ice reduces or eliminates fetches, thereby reducing the wave energy available to reshape the coast (e.g., Trenhaile 1990). Tide ranges are generally less than 1 m along the exposed arctic coasts, but increase to more than 4 m in parts of Hudson Bay and more than 6 m along parts of the Ungava Peninsula (Davies 1980).



North America, Coastal Geomorphology, Fig. 5 The west coast of Canada. This is a SeaWiFS image (courtesy of NASA) from 09/08/01. (A) The glaciated coast between Alaska and Vancouver Island. Rocky islands, fjords, and a paucity of sandy beaches are characteristic. (B) Vancouver, British Columbia (northern $\diamond$) is on the northern edge of the Fraser River delta, on the Strait of Georgia. Seattle ($\diamond$) is on Puget Sound, connected to the Pacific through the Strait of Juan de Fuca. Fjords, rocky islands, and deltas are common



North America, Coastal Geomorphology, Fig. 6 Alaska's Seward Peninsula (figure has west at the top). This is a MODIS image (courtesy of NASA) from 04/08/02. (A) The Yukon River delta, characterized by fine sediments and thermokarst erosion. (B) North coast of the Seward Peninsula, with sandy barrier systems backed by lagoons and extensive marshes. Thermokarst erosion is common in the fine sediment deposits. (C) Kotzebue Sound and the Baldwin Peninsula. The southern shore has extensive coastal dunes. Most of the rest of the coast comprises deltaic mudflats with some thermokarst erosion. (D) Point Hope, at the western end of the Brooks Range

Interior Mountain Province

The coast between the Alaskan Peninsula and northern Kotzebue Sound (Fig. 6) is part of the Interior Mountains and Plateaus Province. The regional morphology is dominated by a series of expansive, Quaternary coastal plains associated with a series of deltas. The largest of these, the 3,000 km² Yukon delta, receives about 90% of the fluvial sediment yield in this Province (Walker 1998). Coastal sediments are relatively fine-grained and mudflats (Flemming 2002) and coastal dune fields (Walker 1998) are common in the vicinity of the deltas.

Rocky Mountain Province

The Rocky Mountain Province is represented along a short distance of the Alaskan coast by the northern terminus of the Brooks Range. This coastline is characterized by permafrost cliffs with elevations between about 10 and 300 m (Walker 1998), with the latter elevations occurring where the Brooks Range directly abuts the coast. Many of the cliffs are fronted by narrow, sandy beaches.

Arctic Lowlands Province

The Arctic Lowlands stretch from a point west of Point Barrow to the northern tip of the Boothia Peninsula and include a number of large, low-relief islands. Banks, Victoria, Melville, and Prince of Wales Islands are the largest of these. Permafrost is ubiquitous in the Province, and many of the characteristic landscape elements of the coastal zone reflect this. Thermokarst depressions and pingos are common (Carter et al. 1987). Coastal morphology reflects extensive control by glaciation and sea ice. This coast is ice-bound for much of the year, limiting wave-induced erosion. The eastern reaches of the lowlands are sheltered further by islands. Coastal sediments are typically derived from *bluff* erosion, glacial tills, or deltas, such as those of the Coleville and the MacKenzie Rivers. Much of the drainage of northern Canada debouches across the lowlands, creating extensive delta networks and delivering large sediment loads.

Canadian Shield Province

The Canadian Shield coast stretches from the relatively low relief of the Boothia Peninsula to the high-relief coast of mainland Newfoundland (Labrador) and the northern coast of the Gulf of Saint Lawrence. It includes many islands, most notably Baffin Island. The landscape is dominated by the effects of Pleistocene glaciation (continental and alpine) and subsequent isostatic rebound. There is continuous permafrost

along most of the low-lying coast, with sporadic permafrost found as far east (and south) as Labrador (Shilts et al. 1987). Marine processes are minimized by short fetch distances caused by the presence of islands and by the frequent presence of nearshore or shore-fast sea ice. The coasts of Baffin Island and much of Labrador are characterized by coastal cliff and fjord systems typical of high-relief *glaciated* coasts. The most distinctive large-scale morphological feature of the Shield coast is Hudson Bay, a flooded depression in the shield that has an east-west extent of about 800 km and a north-south extent exceeding 1,000 km (including James Bay). The coast of southern and western Hudson Bay has emergent beach ridge systems that may extend inland up to 75 km (Martini et al. 1980). Many of the beaches along eastern Hudson Bay through Ungava Bay are characterized by boulder strewn *tidal flats*. Rias are common along many reaches of the north coast of the Gulf of Saint Lawrence (Dubois 1980).

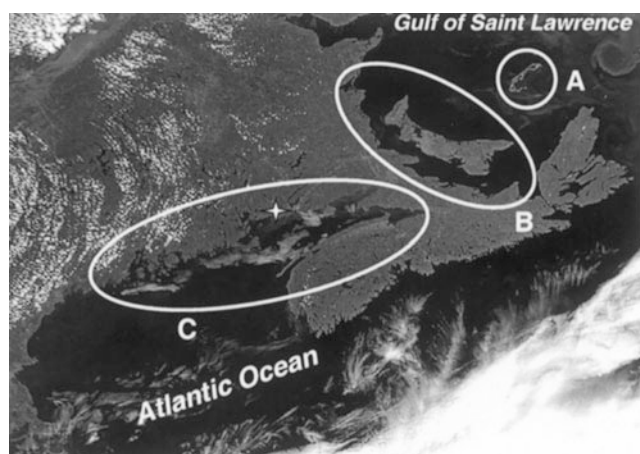
Appalachian Mountains and Plateaus Province

The maritime provinces of Canada, and the coast of Maine, are in the Appalachian Mountains and Plateaus Province. This includes the island of Newfoundland, Prince Edward Island, and Cape Breton Island. Most of the shores of this region display relict glacial landforms from the Pleistocene. The exposed northern coast of Newfoundland and the eastern coast of Nova Scotia are high wave-energy environments, characterized by *rocky* shores with till-derived, sand and gravel beaches in protected coves and estuaries. The coasts surrounding the southern reaches of the Gulf of Saint Lawrence are characterized by barrier bay systems. Sediments are derived mainly from the erosion of glacial tills or from off-shore sources as a result of *transgression*.

The Bay of Fundy (Fig. 7) is a low wave-energy environment that is dominated by tidal processes. Maximum spring tide ranges exceed 15 m in the southern arm of the bay, the Minas Basin, where large intertidal sand bodies are common. Wide, intertidal mudflats are common in the upper reaches of Chignecto Bay, Fundy's northern arm (e.g., Davidson-Arnott et al. 2002).

Atlantic Coastal Plain Province

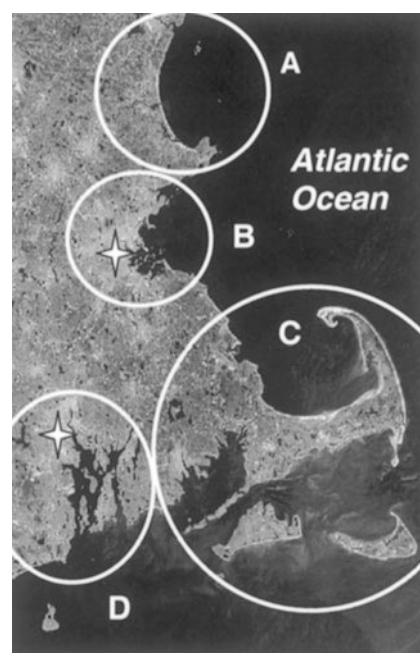
The Atlantic Coastal Plain Province reaches from Maine southward to the United States-Mexico border. Its morphological characteristics have been described by Walker and Coleman (1987). Tectonically, this is a passive, or trailing-edge, coastline (Inman and Nordstrom 1971) of low relief except where glacial or isostatic processes have created coastal cliffs in the northern parts of the Province. Most of the coastline in this Province is characterized by barrier island



North America, Coastal Geomorphology, Fig. 7 Southeastern Canada. This is a SeaWiFS image (courtesy of NASA) from 20/07/01. (A) Îles de la Madeleine, Quebec. The islands are characterized by extensive sandy beaches and coastal dunes, with occasional rocky outcrops. (B) Prince Edward Island, Northumberland Strait, and the coastal plains of New Brunswick and Nova Scotia. The exposed coasts are characterized by barrier/estuary systems typically separated by cliffed headlands. (C) The Bay of Fundy and the glaciated coasts of New Brunswick and Maine. Muddy tidal flats are common in the Bay of Fundy east of Saint John (◇)

and estuary/bay systems, including many of the longest barrier islands on earth. Spring tide ranges on the open Atlantic coasts are typically less than 2 m and less than 1 m along the coast of the Gulf of Mexico. Average annual wave energy decreases southward in this Province.

The coast from Maine south to New York (Figs. 7, 8, and 9) displays the effects of continental glaciation during the Pleistocene, extensive submergence during the Holocene transgression, and a strong geological control on coastline features (Fitzgerald et al. 2002). There are few sandy barriers north of Bigelow Bight (Fig. 8). Drowned river mouths, moraines, and glacial outwash features are common. Examples of the former include Penobscot Bay, Maine, Boston Harbor (that includes drumlins, Fig. 8), Narragansett Bay, Rhode Island (Fig. 8), and the New York Bight (Fig. 9). The dominant glacial sedimentation features include Cape Cod and the islands of Martha's Vineyard and Nantucket (Fig. 8) and Long Island (Fig. 9). Western and northern Cape Cod has a spine of moraine deposits, but the most dramatic coastal features are the large, recurved spits formed from reworked outwash. Barrier systems are common along the Cape Cod shore, and extensive sand dunes occur in the northern reaches. Similarly, Long Island, New York, comprises a morainal spine along its northern coast and an outwash plain along its southern coast. Glacial sands, eroded from the eastern end of the island, have been transported westward by longshore currents to form several barrier islands, including Fire Island. The New York Bight (Fig. 9) is the drowned valley of the Hudson River, bounded to the east by Breezy Point, a barrier

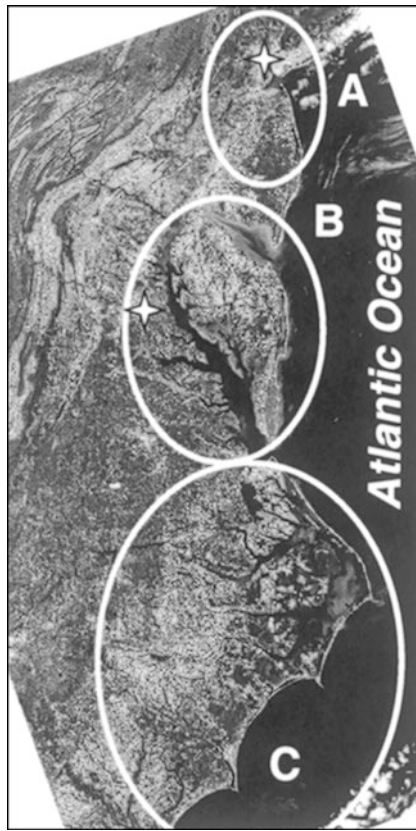


North America, Coastal Geomorphology, Fig. 8 The southern New England coastline. This is a MISR image (courtesy of NASA) from 13/04/00. (A) Bigelow Bight, from Portsmouth, New Hampshire, past the Plum Island barrier, south to Cape Ann, Massachusetts. (B) Massachusetts Bay, including Boston Harbor and the city of Boston (◇). (C) The classic recurved spit of Cape Cod with sandy barrier systems common along the eastern shore and the islands of Martha's Vineyard and Nantucket in the southwest and southeast. (D) The glaciated coastline in the vicinity of Narragansett Bay and Providence, Rhode Island (◇)

spit, and to the south by Sandy Hook, another barrier spit. Staten Island is formed by a remnant moraine and is near the southern limit of glacially affected coastline in North America.

The rest of the coastline in this geomorphic Province is dominated by a series of barrier and lagoon systems. Fisher (1982) identified four barrier island coastlines in the United States, and all are in this Province: the Middle Atlantic States from New York (south shore of Long Island) through North Carolina, South Carolina and Georgia Sea Islands, Florida's east coast, and the Texas and Mexico coasts. The groups are distinguished not only by geography but also by characteristic form assemblages. All of the barrier islands include coastal dunes (although perhaps substantially human-altered), and many of the larger islands have extensive dune systems.

In the Middle Atlantic States, the shores of New Jersey and the Delmarva Peninsula (Fig. 9) are morphologically similar. Both are characterized by barrier island chains often separated from the mainland by wide bays. Both coasts are terminated by large estuaries, Delaware Bay to the north and Chesapeake Bay to the south formed by the submergence of fluvial systems. Along the northern New Jersey coast, the direction of net *alongshore transport* is toward the north, terminating at Sandy Hook. From central New Jersey to Chesapeake Bay,



North America, Coastal Geomorphology, Fig. 9 The mid-Atlantic seaboard of the United States. This is a MODIS image (courtesy of NASA) from 11/10/00. (A) New York Bight, western Long Island, New York City (◇), and the barrier systems of northern and central New Jersey. (B) Chesapeake Bay, a classic ria, and the Delmarva Peninsula. The ◇ indicates the approximate location of the Baltimore/Washington urban region. (C) The Carolina Capes-Hatteras, Lookout, and Fear (from north to south), with their long, sandy barriers, and Pamlico Sound

transport along the Atlantic coast is to the south. The configurations of Cape May and Cape Charles result in part from this process. Kraft (1985) notes that this reach typifies the response of coastal plains to marine transgressions with elongated and often overlapping islands. The extensive use of shore protection structures and beach nourishment projects make the barrier coast of New Jersey one of the most extensively, human-altered shorelines in North America (e.g., Nordstrom 2000).

The Middle Atlantic coast of North Carolina also features nearly continuous barrier island chains (Fig. 9). There are pronounced changes in orientation along this coast, with structural control across the boundary of the Chesapeake-Delaware Basin and the Cape Fear Arch (Walker and Coleman 1987). The Holocene transgression submerged several large, low-relief drainage basins, leaving the elongated barrier chains configured as a set of capes, Capes Hatteras, Lookout, and Fear. The first two are well separated from the

mainland by Albermarle and Pamlico Sounds. The larger barrier islands (e.g., Hatteras) have large, well-developed dune complexes.

The South Carolina and Georgia Sea Islands comprise a set of barriers with a distinctive morphology. Walker and Coleman (1987) describe three island types: Pleistocene remnants, Holocene barriers, and marsh islands (protected by the barriers). These barrier islands are much less elongated than their Middle Atlantic counterparts, with many more *inlets*.

The east coast of Florida between Jacksonville and Palm Beach is straight, other than the foreland caused by Cape Canaveral. There is no fluvial sediment delivery to the coast along this reach. Wave energies along this coast are moderate to low, but the region is subject to high-energy events caused by tropical storms. The barrier islands are elongated, and the bays behind them are also elongated with relatively few inlets. There are small dune systems on the barriers, sabellariid reefs in the intertidal zone (Kirtley and Tanner 1968), and relict beach ridge plains on the mainland. Along the southern reaches of the Florida Peninsula, coral reefs replace the sabellariid colonies, there are low-relief limestone cliffs, and calcium carbonate replaces silica as the dominant constituent of the beach sand. Mangroves occur in sheltered waters of the Florida Keys.

The west coast of Florida is *micro-tidal* with very low wave energy. There are scattered barrier systems in the central sections of this reach, with mainly silica sands. Mangroves become locally important along the northern coasts, as wave energy north of Tampa Bay becomes trivial. In the Florida Panhandle, wave energy increases westward, although tide ranges remain small. There are well-developed barrier, beach ridge, and coastal dune systems along this coast, especially from the Apalachicola River Delta west to Pensacola.

The major morphological features of the US coast of the Gulf of Mexico are the Mississippi River Delta and the barrier islands of Texas (Fig. 10). There is minimal structural control along this coast, and the coastal plain is quite wide, 100 km or more in Texas and Louisiana (Walker and Coleman 1987). Wave energy and tide ranges are low, although this region is also vulnerable to tropical storms. Beach sediments along the northwestern Gulf show strong influences of voluminous, fine sediment delivery by the Mississippi River and subsequent redistribution to the east by Gulf currents. The river drains more than $3 \times 10^6 \text{ km}^2$ and discharges $580 \times 10^9 \text{ m}^3/\text{year}$ of water and $330 \times 10^9 \text{ kg/year}$ of suspended sediment (Meade 1981). Delivery of this sediment load into the low-energy environment of the northern Gulf of Mexico has caused the progradation of the Mississippi River Delta. The Mississippi River Delta is one of the largest, and most studied, deltas on earth (e.g., Walker and Coleman 1987; Wells 1996). Its classic bird-foot shape has grown more pronounced through historic times as the loss of *wetlands* has emphasized the configuration of the natural levees. Wetland loss, at rates averaging in



North America, Coastal Geomorphology, Fig. 10 The Gulf of Mexico coasts of Texas and western Louisiana. This is a MODIS image (courtesy of NASA) from 13/11/02. (A) The coast of Texas from the Rio Grande River to the border with Louisiana. Long sandy barrier islands and extensive lagoon/marsh systems are characteristic of this low-lying coast. Houston (◇) is near the head of Galveston Bay. (B) The Mississippi River Delta, characterized by extensive marshes and fine sediment deposition (note sediment plume from river mouth). New Orleans (◇) is located on the south shore of Lake Pontchartrain, with the latter created by the progradation of the delta

excess of 50 km²/year for more than a half a century (Reed 2002), represents the greatest rates of coastal land loss in North America. These losses are caused by oxidation of organic materials, subsidence of the delta deposits, and sea-level rise; the local, relative rate of sea-level rise exceeds 10 mm/year (Reed 2002). Coastal erosion contributes only slightly to the wetland loss rate, although it has contributed to the formation of several barrier island chains, such as the Timbaliers and Chandeleurs.

The coastline of Texas (Fig. 10) is low energy and microtidal. There is a wide coastal plain of low relief, and the mainland coast is fronted by a series of long barrier islands and spits, including the longest undeveloped barrier in the world, Padre Island. The barriers are transgressive, being driven landward by sea-level rise. Beach sediments are also derived from the Mississippi River and the four Texas streams that discharge through the barriers: the Brazos, San Bernard, Colorado Rivers, and the Rio Grande (Aronow and Kaczorowski 1985). The bays are also elongated; the Laguna Madre, for example, is about 200 km in length. Numerous small estuaries, from submerged drainage systems, form the landward perimeter of the bays. The entire coastal system is vulnerable to erosion, inundation, or washover, caused by tropical storms. Dune systems are common, especially along the southern coasts. Inlets are widely spaced, tend to have small ebb tidal deltas, and sediment *bypassing* is sporadic and event driven (Morton et al. 1995). Galveston Island, once inundated by the most catastrophic hurricane in US history,

is now the most human-altered barrier in Texas, comprising groin fields, landfill, beach nourishment, and a seawall.

The Gulf coast of Mexico has several well-developed barrier/marsh/lagoon systems. South from the Rio Grande Delta is the Laguna Madre and a long barrier spit and island system backed by a series of estuaries. This system extends from the Rio Grande south to the vicinity of La Pesca. The barriers and lagoons then become smaller and discontinuous until the large estuarine delta system of the Rio Panuco, near Tampico. Cabo Rojo is part of the Isla del Idolo barrier complex, including dunes and *beach ridges*, that protects the Laguna Tamiahua, and it is the only pronounced coastal feature in this reach. This coast is also in the range of tropical coral habitat.

Coral is also found off the Veracruz coast, and this reach is also characterized by large expanses of marshes, lagoons, and sandy barriers. Wave energy in this region is low, as is the tidal range. At about this location, the coast begins a pronounced bend toward the northeast and the large Laguna de Términos. Mangroves occur in sheltered locations within the lagoon systems, and dunes are common on the barriers. This coast changes largely in response to tropical storms and changes in water level. The western coast of the Yucatan Peninsula is dominated by sedimentation from large drainage systems, leading to the development of large beach ridge complexes associated with the deltas (Psuty 1966). The eastern shore of the Yucatan is also a low-energy environment, with coral reef forming a barrier, and mangroves dominate in sheltered waters.

Distribution of Muddy Coasts

Few of the open ocean coasts of North America have extensive systems of muddy beaches. Their distribution is described by Flemming (2002), and the locations of the most extensive systems are summarized here. Muddy coasts are usually low-gradient and often provide substrate for deltaic, lagoonal, or estuarine marsh systems. Typically, muddy coastal environments are found adjacent to large deltas, such as the Mississippi River Delta (Fig. 10) or where wave energy is low in estuaries or embayments, such as the Bay of Fundy (Fig. 7). Large, mud-dominated coasts are found in the lagoon systems of tropical Mexico (Fig. 2), the Gulf of California, and the Baja California Peninsula (Fig. 3) and are usually associated with marsh systems. There are few muddy systems along the west coast of the United States, and those are confined to relatively small estuaries, except for the marshes around San Francisco Bay (Fig. 4) and the Columbia River Estuary. The muddy environments of the western Canadian coast are associated with the deltas of the Fraser and Skeena Rivers (Fig. 5), and there are few substantial deposits along the Pacific coastline of Alaska. The rest of the Alaskan coast

has muddy coastlines around Bristol Bay and in the vicinity of Norton Sound and the Yukon River delta (Fig. 6).

Flemming (2002) notes that the major reaches of muddy coast in Arctic and Atlantic Canada are few but include extensive mudflats in Hudson, James, and Ungava Bays, muddy marshes in the St. Lawrence estuary, and fringing mud deposits in the Bay of Fundy (Fig. 7). The Atlantic coast of the United States is characterized by lengthy barrier systems that provide shelter for back barrier bays and marshes such as those in New Jersey and landward of the Carolina Capes (Fig. 9). In all of these cases, the open Atlantic coasts are mainly sand. The estuaries of this coast, including Delaware and Chesapeake Bays (Fig. 9), and Long Island Sound, provide sufficient protection from wave action to accumulate large muddy deposits, usually manifested in *salt marshes*. Similar systems are found around the Gulf of Mexico. Sandy barriers protecting muddy lagoons and marshes, with some estuaries, characterize the coasts of Mississippi, Alabama, and Texas (Fig. 10), and much of the Mexican coast south to Belize. Notable for the extent of the deposits are Laguna de Términos (Fig. 2) and Laguna de Tamiahua. The Mississippi River Delta (Fig. 10) stands alone as the only significant reach of open coast to be characterized by muddy deposits. This reflects the river's extremely large, fine sediment load and the relatively low wave energy in the northern Gulf.

Distribution of Coarse Clastic Coasts

Beaches composed of coarse clastic sediments, also referred to as shingle, gravel, pebble, cobble, or boulder beaches, are most common in *periglacial* or *paraglacial* environments, along cliffed coasts, or near the mouths of steep gradient coastal streams (Orford et al. 2002). These beaches have steep profiles and are relatively immobile compared to sand-dominated systems. Davies (1980) indicates that such deposits are of some degree of relative importance around most of the North American continent. Specifically, he categorizes these beaches as occasionally important along a reach of the Pacific coast roughly bounded by Acapulco (Fig. 2) and San Diego (Fig. 3), in Canada, from northern Newfoundland to northern Nova Scotia (Fig. 7), and along the reach from Maine (Fig. 7) to New York (Fig. 9). Davies classifies coarse clastic systems as important along the coasts of Canada and the United States north of Puget Sound on the west coast and north of Newfoundland on the east coast. They are also important along the coasts of Newfoundland and northern Maine.

Cross-References

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- [Cheniers](#)
- [Cliffed Coasts](#)
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- [Wetlands](#)

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Notches

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Definition

Littoral notches are more or less horizontal erosion features close to sea level into steeper coastal slopes. They prove that processes of destruction around the surf level are more intensive than in subtidal or supratidal positions. Their relation to tidal levels and their shape may differ from one type of notch to the other, because there are several genetic types

of notches around the coastlines of the world (Trenhaile 2014, 2015, 2016).

Notches Caused by Melting

At ice cliffs of shelf ice, or along drifting icebergs, sharply incised notches occur at sea level, caused by melting processes of warm surficial waters in Arctic or Antarctic environments during summer. These forms develop in a very short time (several days), but they are usually rapidly dislocated (uplifted above sea level or tilted) by the moving of ice margins or tilting of icebergs during the melting process, when the point of gravity will shift. Notches caused by melting processes also occur in sediments of permafrost environments, if warm seawater in the surf zone undermines steep slopes. These forms are rarely stable for longer times, because by melting of the cliff soft material will creep or slide into the surf zone and covers the notches.

Notches Caused by Chemical Solution

In the scientific literature, this kind of notch formation is mentioned for carbonate rocks, because the sharp microforms resemble karstic features from terrestrial environments, and the seawater is believed to act as an aggressive agent upon limestones. This conclusion is incorrect, because seawater is oversaturated (up to several 100%) by dissolved carbonates and is not able to destroy carbonates by dissolution. Dissolution may be responsible for notch formation in halites or gypsum rocks, but no descriptions from these environments exist.

Notches Caused by Salt Weathering

In sedimentary or volcanic rocks with clasts of sand or coarser grains, salt weathering may act as a notch-forming process (Trenhaile 2014). It is often restricted to those levels above mean sea level, where salt water rise through capillary action in the rock, transporting salts. By wetting and drying, the grains in the rock loosen, producing honeycombs and tafoni of different sizes. Galleries of these forms may decorate the rocky shores and, coalesced, form an irregular shaped kind of notch. The process dominates other littoral processes only in warm and arid regions with very limited surf action and tides, as in the eastern Mediterranean.

Notches Caused by Mechanical Abrasion

This is a widespread (although only locally developed) notch type in hard rocks, where sand and pebbles or cobbles move

in the surf zone. They abrade the rock and polish it. These notches show different profiles according to lithological characteristics but are always very smooth, showing the rock's own color, because no algae or other organisms will survive the surf attack with abrading fragments. The relation of this notch type to sea level or tidal levels differs according to the availability of loose sediments. If only a limited amount is present, at a steep coast the polishing occurs in the subtidal; and if a large amount is available, only the uppermost strata of the sediment is moved against coastal rocks, and the abrading and notching process is restricted to cliff parts in the upper surf zone. Therefore, a difference of the notch relative to mean sea level does not always mean sea-level variation or tectonic dislocation. Abrasion notches occur with coastal caves in less resistant rocks. If rocky coasts dip into deep water and no loose material is present in the surf zone, the pressure of wave attack may destroy rocks with textural or structural weakness and breaks out parts of the cliff. In regions with drifting sea ice, this (as well as frost action) may form notch-like features in less resistant rocks.

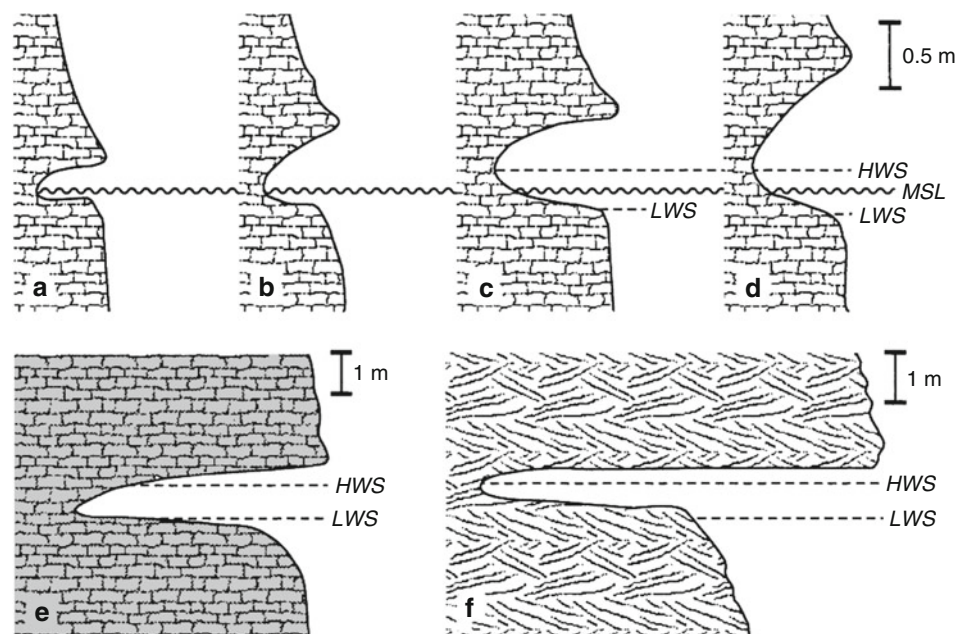
Notches Caused by Bioerosion

This is the most widespread type of notch but is restricted to carbonate rocks including carbonate sandstone (calcarenes) and sometimes dolomites, and to low latitudes (Kellett 1997). Its aspect is a horizontal back cutting of rocky shore faces, often along extensive and continuous stretches of the coastline, with the smaller part under water and a wider above mean sea level. Calling this type “tidal notches” is misleading, as the shape of bioerosive notches differs also according

to exposure (Fig. 1). They are more open and wide with a tidal range larger than 1 m and intensive surf, and narrow with a nearly flat roof in very sheltered and almost tide-less environments. These points are important if the relation of living and fossil notches to varying sea levels is under consideration. In areas of large tidal ranges and extreme surf conditions bioerosional notches may be less developed or absent. Although bioerosive notches will undermine the cliff face rather rapidly (about 1–2 mm/year), collapse of cliffs by this process only occurs in rocks of low resistance like eolianites. Looking closer to the shape of bioerosional notches, they show a sharp microrelief in every part. The normal profile is asymmetrical with a rather flat bottom, an innermost knick-point close to the bottom and a more (exposed) or less (sheltered) inclined notch-“roof” (Fig. 1).

A notch always indicates destructive processes near the surf horizon being more active than below and above. In the case of bioerosive notches, small organisms live in these well-defined belts. We find chitons, limpets (e.g., *Patella* sp.), littorinids, and other gastropods, living and grazing in wetted and well-illuminated zones. Their grazing instrument is a radula, and their food being microscopic organisms like cyanophytes and chlorophytes (and sometimes fungi, see Schneider 1976; Torunski 1979). Where wetting is guaranteed, these organisms live epilithic, showing a greenish color. In the upper and supratidal belt, they escape from drying out and insolation by boring into the rock, thus living as endoliths, dried outer parts showing dark to black colors. The penetration depth (light-compensation depth) depends on the chance for photosynthesis and, therefore, is restricted to parts of a millimeter, depending on grain size and crystal forms in the rock. Under a scanning-electron microscope,

Notches, Fig. 1 Shape of bioerosive notches and their relation to tidal levels (a–d from the Mediterranean): (a) tideless, sheltered; (b) tideless, exposed; (c) microtidal, sheltered; (d) microtidal, exposed; (e) microtidal, sheltered, in Neogene coral rocks of Palau, Micronesia; (f) microtidal, sheltered, in less resistant Pleistocene eolianites of the Bahamas



the pattern of biocorrosion by endoliths exhibits microborings of up to 800,000 per cm², and a three- to ten-fold enlargement of the actively corroded surface. Notch forming depends on the number and activity of limpets and other grazers, which wear down the surface in one grazing event by about 0.01–0.1 mm (bioabrasion). Hence, light can penetrate deeper into the rock, and the endoliths will bore deeper, as well, exhibiting a new sheet of food for the grazers. The restriction of the grazers to well-defined wetting levels results in horizontal bioerosive notches, visible after only some decades of grazing in the same level.

Conclusion

All in all, notches, and in particular bioerosive notches, because of their good and rapid development, are among the best sea-level indicators at the coastlines of the world and are studied well to solve questions of neotectonic dislocation (Antonioli et al. 2015; Boulton and Stewart 2015; Pirazzoli and Evelpidou 2013).

Cross-References

- [Bioerosion](#)
- [Cliffed Coasts](#)
- [Cliffs, Erosion Rates](#)
- [Erosion Processes](#)

- [Microtidal Coasts](#)
- [Sea-Level Indicators, Geomorphic](#)
- [Shore Platforms](#)

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Offshore Sand Banks and Linear Sand Ridges

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Shelf sand banks and linear sand ridges are found on numerous modern and ancient continental shelves where sufficient sand exists and currents are strong enough to transport sand-sized sediment (Off 1963; Snedden and Dalrymple 1999; Dyer and Huntley 1999). Sand banks and linear sand ridges are defined as all elongate coastal to shelf sand bodies that form bathymetric highs on the seafloor and are characterized by a closed bathymetric contour (Fig. 1). Other terms used to refer to these specific bathymetric features include *linear shoals*, *shoreface ridges*, *shoreface-attached* or *detached ridges*, *shoreface-connected* or *disconnected ridges*, *tidal current ridges*, and *banner banks*.

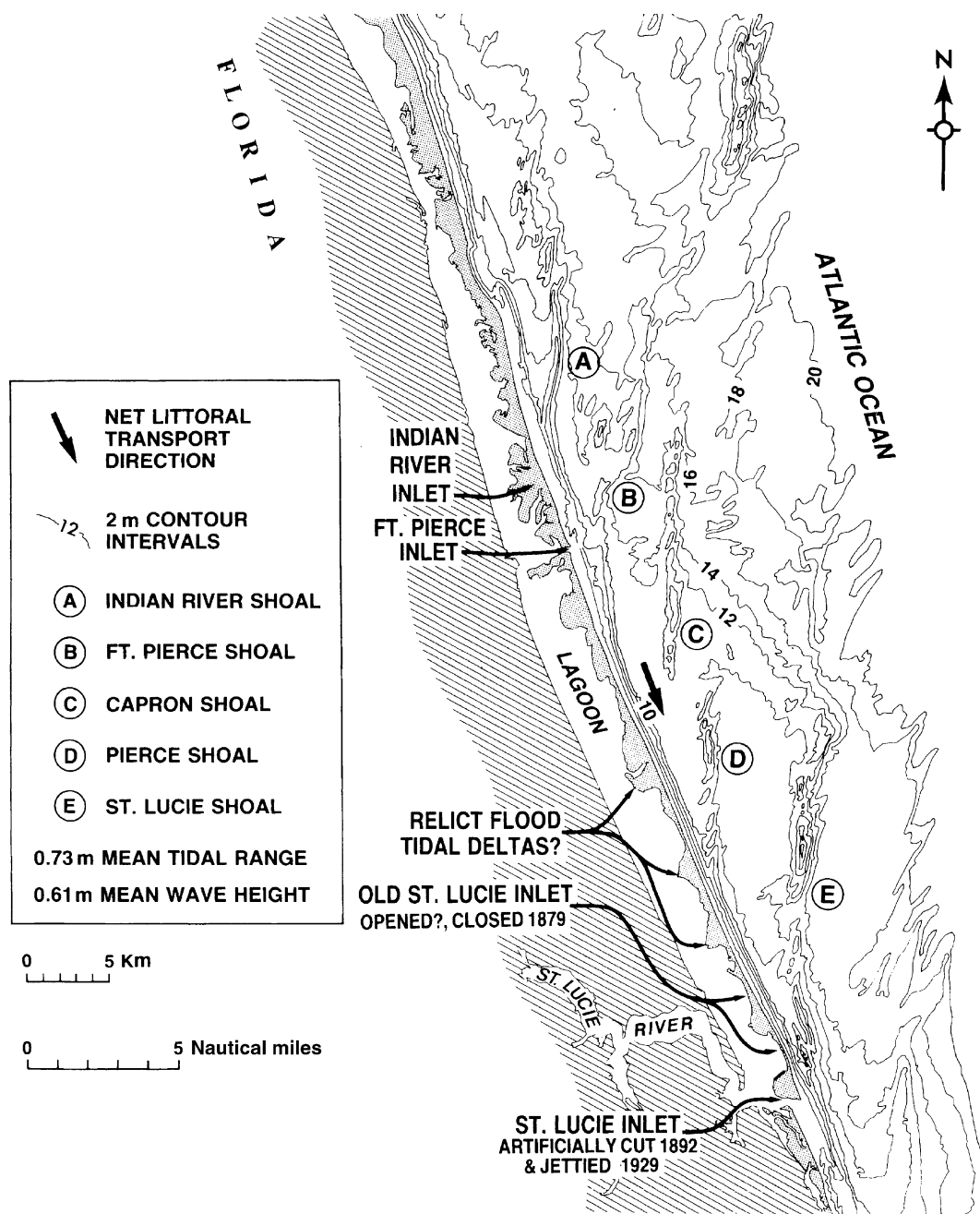
Typically, these linear sand bodies have heights that are more than 20% of the water depth, lengths that range from 5 to 120 km, relief up to 40 m, and side slopes that average $< 1^\circ$. They are 0.5–8 km wide, asymmetrical in profile, and consist of unconsolidated fine-to-coarse sand or even gravel. Axes of these sand bodies are generally oriented shore-parallel, shore-oblique ($10\text{--}50^\circ$), or shore-normal when compared to the adjacent coastline, but any orientation is possible. Sand banks and linear sand ridges are present in a wide range of water depths and found in estuaries, at tidal entrances, adjacent to coastlines (offshore headlands, spits, and barrier islands), on the exposed shelf, and along the continental shelf edge. Moreover, these sand bodies usually occur in groups with spacing of individuals on the order of 250 times the water depth, but solitary ridges do occur (Snedden and Dalrymple 1999). Two or more linear ridges grouped together are referred to as a *sand ridge field*, but other terms like *ridge* and *swale topography* are also used.

Initial investigations regarding bathymetric irregularities on continental shelves date as early as the 1930s (Van Veen 1935; Veatch and Smith 1939). Since that time, investigations into the origin, morphology, and geology of sand ridges may be organized into three groups: (1) morphology and surficial sediment studies (e.g., Off 1963; Uchupi 1968; Duane et al. 1972); (2) genesis and hydrodynamic regime (e.g., Swift and Field 1981; Huthnance 1982; Swift 1985; McBride and Moslow 1991; Snedden and Dalrymple 1999), and (3) internal geology and stratigraphy (e.g., Penland et al. 1989; Snedden et al. 1994; Dalrymple and Hoogendoorn 1997).

Classification

Many different classifications have been proposed for sand banks and linear sand ridges but the three most widely accepted are highlighted here. By the 1980s, leading investigators had classified linear sand bodies on continental shelves into two types – storm or tide built (Amos and King 1984; Swift 1985). In other words, either storm-generated or tide-generated currents were primarily responsible for ridge development and maintenance. Through time, however, researchers also recognized the importance of ridge precursors (i.e., sand body nuclei or initial irregularities, such as ebb-tidal deltas) and the subsequent hydrodynamic reworking of the nuclei (e.g., McBride and Moslow 1991), which eventually led to two generic classifications (Dyer and Huntley 1999; Snedden and Dalrymple 1999).

Dyer and Huntley (1999) proposed a qualitative classification based on a generic relationship between different shelf sand bodies in light of their origin and development. They concluded that the morphodynamic stability models proposed by Huthnance (1982) were the most suitable at explaining the interaction between flow and sand body orientation. Their classification is composed of three primary ridge types and several subtypes as follows:



Offshore Sand Banks and Linear Sand Ridges, Fig. 1 Example of shore-oblique shelf sand ridges, each characterized by a closed bathymetric contour. The Fort Pierce sand ridge field is located south of Cape

Canaveral along the east coast of Florida, USA. (From McBride and Moslow 1991 with permission from Coastal Science). Orientations of other linear sand bodies may be shore-parallel or shore-normal

Type 1 Open shelf ridges

Type 2 Estuary mouth

- (a) Ridges (wide mouth)
- (b) Tidal delta (narrow mouth)
 - (i) Without recession (ebb-tidal deltas)
 - (ii) With recession (shoreface-connected ridges)

Type 3 Headland-associated banks

- (c) Banner banks (nonrecessional headland)
- (d) Alternating ridges (recessional headland)

Recognizing that the Huthnance stability model was applicable to both storm- and tide-built ridges, Snedden and Dalrymple (1999) proposed a unified model for ridge genesis and maintenance. They classified shelf sand ridges according to the amount of reworking of the original sand body (i.e., precursor irregularity) based on sedimentologic and stratigraphic evidence. As such, ridges were subdivided into three classes representing an evolutionary progression of increased ridge reworking and migration (Table 1). Class

Offshore Sand Banks and Linear Sand Ridges, Table 1 Classification and primary characteristics of shelf sand ridges and banks (modified from Snedden and Dalrymple 1999)

	Class I	Class II	Class III
Type	Juvenile/static ridge	Partially evolved	Fully evolved
Precursor	Largely preserved	Partially preserved	Not preserved
Dynamics	Ridge stationary or rapidly buried	Ridge migrates < original ridge width	Ridge migrates $\geq$ original ridge width

I ridges retain all of their original nucleus (i.e., precursor irregularity). Class II ridges are partially evolved sand bodies that have migrated less than their width and thus contain recognizable evidence of their original nucleus, whereas Class III ridges have migrated a distance equal to or more than their original width, thereby eroding virtually all evidence of their original nucleus.

Depositional Setting and global occurrence

Sand ridge genesis is most favorable during transgression on continental shelves that are characterized by the following four conditions: (1) initial irregularities most commonly developed in the nearshore zone, (2) sufficient supply of loose sand, (3) sand-transporting currents (tidal or storm-driven), and (4) sufficient time for the sand to be molded into a ridge or ridge field (Snedden and Dalrymple 1999).

Some of the best developed, *storm-maintained* linear sand ridges and fields are located on continental shelves along North America, South America, and northern Europe. Linear sand ridges on the Atlantic shelf of the United States reside offshore Long Island, New York, New Jersey, Maryland, North Carolina, and the east coast of central Florida (Duane et al. 1972; McBride and Moslow 1991). Numerous linear shoals and banks are also found along the northern Gulf of Mexico of the United States offshore the Florida Panhandle (McBride et al. 1999), Louisiana (Penland et al. 1989), and east Texas (Rodriguez et al. 1999). Other classic localities include eastern Canada offshore Sable Island, Nova Scotia (Dalrymple and Hoogendoorn 1997); northern Europe along the German Frisian Islands (Swift et al. 1978); and South America offshore northern Argentina, Uruguay, and Brazil (Swift et al. 1978). In contrast, well-developed, *tide-maintained* linear sand ridges or fields are located in the North Sea around England (Kenyon et al. 1981; Belderson et al. 1982) and other embayments characterized by >3 m tidal range (e.g., Gulf of Korea; Gulf of Cambay on Indian west coast; northern delta of Amazon River; northern end of Persian Gulf; Prince of Wales Strait just off Cape York, northernmost tip of Australia) as discussed by Off (1963).

Cross-References

- Coastal Currents
- Continental Shelves
- Cross-Shore Sediment Transport
- Estuaries
- Offshore Sand Sheets
- Longshore Sediment Transport
- Shelf Processes
- Storm Surge
- Tidal Inlets
- Tides

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Offshore Sand Sheets

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Sand sheets are common features of modern-day, shallow seafloors. In addition, the ancient rock record contains numerous examples of sheet sandstones in both coastline and shallow marine open shelf associations (e.g., Goldring and Bridges 1973). The Cretaceous epicontinental interior seaway of North America, for example, contains extensive sand sheet deposits and has been the focus of intense research due to their petroleum-bearing nature (Brenner 1978; Walker 1983; Shurr 1984; Kreisa et al. 1986; Nummedal et al. 1989; Walker and Eyles 1990; Winn 1990). Modern sand sheets can also provide societal and economic benefits such as good quality beach nourishment sand, strategic minerals and ores, and productive marine habitat. On the other hand, the mobile sediments and bedforms associated with sand sheets can create navigation hazards and uncover buried cables and pipes. Despite their prevalence and importance, little is known about their geometry, bounding surfaces, dimensions, internal structure, origin, evolution, composition, and texture. In part, the diversity of sedimentary processes that operate in shallow marine environments accounts for the paucity of sand sheet classification schemes and models (e.g., process, facies, etc.). The disparate definitions and descriptions found in the coastal and marine literature attest to this reality.

What Is a Sand Sheet?

Sand sheet definitions range from mass accumulations of sand to a veneer or sheet-like body of surficial sediment (Twichell et al. 1981; Stride et al. 1982; Belderson 1986). They can range in thickness from a few centimeters to tens of meters and have a lateral persistence of a few meters to tens of kilometers. Some genetic definitions include the migration and stacking of large bedforms (i.e., sand waves), especially during transgressions (Nio 1976; Stride et al. 1982; Walker 1984; Belderson 1986; Saito et al. 1989). Sand sheets have also been given:

- lithofacies descriptions such as glauconite facies and calcareous facies
- informal designations such as “facies A” and “A sand”
- descriptive designations such as coarse sand, cross-bedded facies and surficial sand sheet
- environmental interpretations such as tidally dominated sand sheet facies, storm-dominated sand sheet facies, or transgressive sand sheet, and
- combinations of the above such as shallow marine siliciclastic facies.

(Swift et al. 1971; Swift 1976; Middleton 1978; Knebel 1981; Stride et al. 1982; Walker 1983, 1984; Belderson 1986; Johnson and Baldwin 1986).

Some studies of sand sheets on the eastern North American continental shelf have used the term surficial sand sheet to include a variety of (hierarchical) bedform types and sizes. For example, Swift et al. (1973) and Field (1980) delineated three scales of superimposed sand “bodies” on a storm-dominated, epicontinental shelf surficial sand sheet. From largest to smallest by size, these sand bodies include shoal retreat massifs (cape and estuarine; first order features of Johnson and Baldwin 1986); linear sand ridges (second order features of Johnson and Baldwin 1986); and ripple, megaripple, and sand wave bedforms. Stubblefield et al. (1975) showed that, in some locations on the Atlantic shelf, the base of the sand sheet occurs above the topographic base of the ridges due to storm–current deepening of the swales. Since the various components of the surficial sand sheet form as a result of differences in flow regimes, dissimilarities in lithologic and paleontologic characteristics can be expected. Thus, although the massifs and linear ridges are components of the surficial sand sheet, they are discrete sedimentary facies from the sand sheet facies.

Sand Sheet Settings

Surficial “sheets” of sand veneer large portions of marginal (pericontinental) and epeiric (epicontinental) seafloors. In

some regions, sand sheets are found in association with sand ridges, while in others the sheets constitute individual entities. In particular, the sheets have been shown to be interspersed between sand ridges, to underlie the ridges, or to exist some distance from a ridge field (e.g., Swift et al. 1972; Stride et al. 1982).

Four primary physical processes control sediment transport, deposition and erosion on epi- and peri-continental shelves: ocean currents, meteorological (storm) currents, density currents, and tidal currents (Swift et al. 1971; Johnson and Baldwin 1986). Globally, 3% of the shelves are ocean current-dominated, 80% are storm-dominated, and 17% are tide-dominated (Swift et al. 1984). The most extensively studied sand sheets within the storm-, ocean current-, and tide-dominated settings are located on the middle-Atlantic Bight of the North America shelf, the southeast African shelf, and the epicontinental seaways rimming the British Isles, respectively.

Oceanic Current-Dominated Sand Sheets

The primary process controlling the surficial sand sheet characteristics on oceanic current-dominated shelves is the inherently unsteady geostrophic boundary current. The best studied example of an oceanic current-dominated sand sheet is located along the southeast African shelf (Flemming 1978, 1980, 1981). On the narrow central region of the southeast African shelf, a wide variety of bedforms is molded onto a sand sheet that is maintained by the powerful, geostrophic Agulhas Current (surface velocities $>2.5 \text{ m s}^{-1}$), a component of the major Indian Ocean circulation (Pearce et al. 1978). Sediment dispersal along this shelf is also influenced by the morphology of the upper continental margin, the wave regime, wind-driven circulation, and sediment supply. Other oceanic current-dominated sand regimes include the outer Saharan shelf, the South American shelf, the Newfoundland shelf, the Osumi Strait, and the trough region of the Korean Strait (Newton et al. 1973; Johnson and Baldwin 1986; Park and Yoo 1988; Barrie and Collins 1989; Ikehara 1989).

Storm-Dominated Sand Sheets

Storm-dominated sand sheets are typified by their intra- and inter-sheet bathymetric, hydraulic, sedimentologic, and stratigraphic variability. The most comprehensive studies of storm-dominated sand sheets are of those situated along the North American Atlantic shelf between Cape Cod and South Carolina (Swift et al. 1972; Knebel 1981; Pilkey et al. 1981; Swift et al. 1981); the Pacific shelf off northwest United States (Clifton et al. 1971; Sternberg and Larsen, 1976); the Bering Sea (Cacchione and Drake 1979; Field et al. 1981; Nelson et al. 1982); the Gulf of Mexico (Uchupi and Emery 1968; Nelson and Bray 1970; Hobday and Morton 1984); the Bahama Bank (Mullins et al. 1980; Hine et al. 1981); the

southern Brazilian inner continental shelf (Figueiredo et al. 1982); and the Argentina inner shelf (Parker et al. 1982).

The sand sheet located on the United States' east coast continental shelf is one of the largest storm-dominated sheets in existence: Over 85% of the sea floor from Georgia to New England is veneered by sand (Knebel 1981; Swift et al. 1981). In general, the spatial and temporal distribution of processes collectively have had a diverse and widespread effect on the attributes of the sand sheet. The central and southern Atlantic shelf sand sheet is the product of an autochthonous regime formed by erosional shoreface retreat during the Holocene transgression (Swift et al. 1972; Swift 1976). The primary erosional agents have been surf-zone wave action, wave-driven currents, and wind-driven currents seaward of the surf zone. This erosional transgression has resulted in the overprinting of the relict nearshore topography by a palimpsest sand sheet which, in turn, is molded with ridges and swales (Swift et al. 1972; Knebel 1981). The sand sheet is discontinuous, ranging in thickness from 1 m to 10 m. The importance of antecedent topography is indicated by the sand sheet thickness which increases in the vicinity of Pleistocene shelf river valleys, that is, Hudson River, Great Egg River, and Delaware River, and shoal retreat massifs (Swift et al. 1972, 1981; Knebel 1981) and decreases on the paleo-interflaves.

Tide-Dominated Sand Sheets

One of the earliest studied tidally dominated modern sand sheets is located on the epicontinental seaway rimming the British Isles and on the northwest European shelf. Since the 1930s, a plethora of information has been compiled regarding the sand sheet and sand bank facies in the North Sea and Irish Sea. This information includes surface morphological, textural, and process data (e.g., Johnson and Belderson 1969; Stride 1970; McCave 1971; Terwindt 1971). From these data, theoretical (conceptual) models have been constructed regarding large- and small-scale bed-form development, maintenance and internal structure (e.g., Houbolt 1968; Allen 1980; Johnson et al. 1981) and sediment transport paths (e.g., Stride 1963; Johnson et al. 1982).

The thin sand sheet ($< 12 \text{ m}$) flanking the British Isles has formed from the unmixing of Pleistocene-age deposits during the Holocene transgression (Nio 1976; Stride et al. 1982). Materials move along transport paths associated with longitudinal and transverse tidal velocity gradients which produce bedload parting and convergence zones. The net result of tidally dominated sediment transport and deposition is to produce a well-sorted sandy deposit where fine-grained material accumulates at the ends of the bedload transport paths (i.e., grain size is in quasi-equilibrium with peak tidal current velocity; McCave 1970, 1971; Stride et al. 1982; Belderson 1986). Superimposed on this sheet is a wide variety of bedform types ranging from small-scale

Offshore Sand Sheets, Table 1 Classification and attribute list of the tide-dominated sand sheet facies (from Fenster 1995)

Sand sheet environments					
	Estuaries				
Attributes	Coastal plain type (upstream converging funnel)	Semi-enclosed (constricted mouth) (drowned river)	Shelf seas		
			Epicontinental	Pericontinental	Straits
Examples	Oosterschelde Estuary	Long Island Sound	North Sea Northeastern Bering Sea	Georges Bank Nantucket Shoals Irish Sea	Taiwan Strait
	Fraser				
	River Estuary	Chesapeake Bay			Malaca Strait
	Gironde				
	Estuary Ord River	Delaware Bay			Bungo Strait
	Estuary				
Bristol Channel Outer	Korean Strait				
	Thames Estuary			Bass Strait	
	Severn				
	Estuary				
Tidal range (m)	3–15	0.4–1.1 1–1.3	1–4	1–8	1.5–4
Tidal system	Progressive, rectilinear	Standing (resonant co-oscillating) with progressive component: progressive, rectilinear can extend 5–10 km onto shelf	Complex due to interference of more than one amphidromic system: rectilinear to rotary	Rotary	Progressive, rectilinear
Currents (m s ^{−1})					
Peak near-surface	0.6–1.3	0.4–2.2	0.5–1.0	0.3–0.8	0.3–1.0
Peak near-bottom	0.5–0.9	0.3–0.6	0.3–0.6	0.55 times surface velocity	0.25 (ave.)
Dimensions (maximum):					
Length (km)	25	30	400	400	100
Width (km)	1–8	15	50	10–50	40
Area (km ²)	200	450	20,000	4000–20,000	4000
Thickness (m)		Variable, 1–24 m	12	Variable, 1–20 m, up to 80 m	
			(thickness has unknown relationship to original sand waves)		
Depth (m)	6–15 (can reach 35)	33–49	18–44	27–45	70–120
3D geometry		Complex in three-dimensions due to lithofacies zonal variability	Complex in three-dimensions due to bed form sinuosity		
Grain size	Medium-to-coarse sands	0.56–1.07	Fine-to-coarse sand	Fine-to-very fine sand	Fine-to-coarse sand
Mean (mm)	0.32–0.65		0.15–0.40		0.13–0.50
Notes	Distribution influenced by mutual evasive ebb and flood channels, grain size gradient expected	Progressive decrease in mean grain size along velocity gradients (grain size in quasi-equilibrium with peak velocities)	Progressive decrease in mean grain size along velocity gradients (grain size in quasi-equilibrium with peak velocities)		
Sorting composition	Moderate-to-well sorted	Moderate-to-well sorted	Well sorted	Well sorted	Volcaniclastic, consolidated mud clasts, quartz
		Quartzose minor amounts of rock fragments, heavy minerals	Quartzose minor amounts of calcareous material, heavy minerals	Quartzose to subarkosic minor amounts of heavy minerals, rock, and shell fragments	

(continued)

Offshore Sand Sheets, Table 1 (continued)

Sand sheet environments					
Biogenic		Within bedform spatial and temporal density changes	Increase in faunal density and diversity away from zone of large sand waves; within bed form changes		
Surficial elements	Ubiquitous presence of large and/or small bed forms; asymmetry governed by stronger peak ebb or flood current; sand wave $H_t \leq 8$ m 88–300 m	Ubiquitous presence of large and/or small bedforms-asymmetry governed by stronger peak ebb or flood current; sand wave $H_t = 1$ –17 m, $l =$ (most 5–15 m;	Ubiquitous presence of large and/or small bed forms; asymmetry governed by stronger peak ebb or flood current; sand wave $H_t = 1$ –24 m, $l < 400$ m $l = 150$ –750 m	Ubiquitous presence of large and/or small bed forms; asymmetry governed by stronger peak ebb or flood current; sand wave $H_t = 1$ –10 m, most < 3.4 m	Ubiquitous presence of large and/or small bed forms; asymmetry governed by stronger peak ebb or flood current; sand wave $H_t = 0.5$ –15.3 m $l = 100$ –300 m
Lower bounding surface	Erosional truncation due to migrating tidal channels	Erosional truncation due to migrating tidal channels; erosional truncation due to migrating bed forms	Erosional truncation due to migrating bedforms	Erosional truncation due to ravinement surface	Erosional truncation due to ravinement surface
Lateral bounding surface	Controlled by:	Controlled by:	Controlled by:	Controlled by:	Controlled by:
	(1) sand availability	(1) sand availability	(1) sand availability	(1) sand availability	(1) sand availability
	(2) lower estuary physiography	(2) lower estuary/bay geometry and physiography	(2) current velocity	(2) current velocity	(2) current velocity
			(3) wave surge; sand/gravel boundary lies at near-surface mean spring peak tidal current = 1 ms^{-1} ; outer limit of sand (sand/mud) lies at near-surface mean spring peak tidal current = 0.4 ms^{-1} ; decrease in bedform size due to lack of sand	(3) wave surge (4) depth	(3) depth (4) antecedent geology
Internal structure	Erosional features and superposition of tidal channel sequences, megaripple structures, largescale cross bedding with reactivation surfaces, mud couplets, tidal bundles and other rhythmic tidalites of various temporal and spatial scales	Erosional features and superposition of tidal channel sequences, channel fill complexes ranging from complete (mature) to in complete (immature), erosional features associated with bed form trough migration including rising toepoint	Internal structure related to location, 0° , basal surface of sheet; 1° , low angle basal surfaces of sand wave master bedding; 2° , generally pervasive low to moderately dipping sets of crossstrata with high-angle cross-bed dips; 3° , reactivation surfaces		
Vertical sequence	Pleistocene and Holocene sands unconformably overlie channel fill	Glaciolacustrine $\rightarrow$ fluvial $\rightarrow$ low energy estuary $\rightarrow$ high energy estuary with variability due to Holocene sediment supply, cut and fill processes	Successively formed sand wave complexes unconformably overlie Pleistocene gravel lags	Sand wave complexes unconformably overlie Pleistocene glacial diamict; separated by wave cut unconformity	

(continued)

Offshore Sand Sheets, Table 1 (continued)

Sand sheet environments					
Secondary processes	Estuarine mixing and density gradients	Short-term enhancements due to estuarine circulation, storm forcing, flow divergence due to basin geometry and bathymetry	Geostrophic flows, storms, density gradients (water temperature)	Storms	Oceanic circulation
Partial reference list	Langhorne (1973); Nichols and Biggs (1985); Berné et al. (1988, 1993); Harris and Collins (1991); Stride and Belderson (1991)	Ludwick (1972); Bokuniewicz et al. (1977); Bokuniewicz (1980); Bokuniewicz and Gordon (1980); Knebel (1989); Fenster et al. (1990); Fenster (1995)	Harvey (1966); Belderson and Stride (1966, 1969); Johnson and Belderson (1969); Stride (1970); McCave (1971); Caston and Stride (1973); Belderson et al. (1982); Stride et al. (1982); Belderson (1986)	Twichell et al. (1981); Mann et al. (1981); Twichell (1983)	Keller and Richards (1967); Boggs (1974); Park and Yoo (1988); Arita et al. (1988); Malikides et al. (1989)

ripples to large-scale sand waves (e.g., Johnson et al. 1981; Belderson et al. 1982). Johnson et al. (1981) hypothesized that the form, grain size distribution, and internal structure of the sand sheet are related directly to bedform dynamics.

Within the tide-dominated setting, sand sheets can be located in a variety of environments including marine influenced estuary-and bay- mouths, peri- and epi-continental shelf seas, and straits. A review of the sand sheet attributes of each environment reveals that the majority of information known about the tide-dominated sand sheet comes from studies of sand sheets within epicontinental seaways and upstream converging funnel estuaries. More recently, Fenster (1995) conducted a comprehensive study of the tide-dominated sand sheet in Eastern Long Island Sound. This study showed that the evolution of the sand sheet was linked closely to erosion of a marine delta, changes in the geometry and accumulation space of the basin, sea-level changes, attendant nonlinear tidal wave distortion, and common estuarine processes such as the development and migration of tidal channels.

Belderson (1986) postulated that a continuum of sand sheet facies exists between the tide- and storm-dominated sand sheet end members. The surficial and internal characteristics of the sand sheet depend on the degree of storm-versus tidal-influence. In general, tide-dominated sand sheets contain abundant sand waves which face in the direction of the stronger of the peak ebb or flood currents. Thus, the internal configuration of tide-dominated sand sheets primarily consists of pervasive crossbedding, evidence of rhythmic periodicities, large dip angles, and reactivation surfaces. In contrast, storm-dominated sand sheets are devoid of sand waves as oscillatory currents act as a

destructive force on sand waves (Langhorne 1982; Belderson 1986). In these regions, hummocky megaripples develop in response to along-coast geostrophic flows combined with wave orbitals of similar, or weaker intensity and result in hummocky cross-stratification (Swift et al. 1983). Thus, on the tidedominated sand sheet, storm-wave and storm-current activity sporadically increases transport rates while on the storm-dominated sand sheets, sediment transport is controlled primarily by wave- and wind-induced currents (Belderson 1986).

Gaps in Knowledge

The majority of work conducted to date on the sand sheet facies has focused on the depositional environment and geographical context (e.g., Park and Yoo 1988; Saito et al. 1989), modern processes (e.g., Knebel 1981), surface sediment textural characteristics (e.g., Hollister 1973; Schlee 1973), thickness (e.g., Knebel and Spiker 1977), surficial morphology (e.g., Swift et al. 1972; Twichell 1983), formation (e.g., Stahl et al. 1974; Johnson et al. 1981), and expected development (e.g., Johnson et al. 1981). Much less work has focused on the geometry, boundary characteristics, and internal configuration of the sand sheet facies, especially in tidedominated regimes. Although surficial information can be useful for delineating sediment transport paths, a paucity of information regarding the internal structure of the sand sheet facies partially has been responsible for limited (or incorrect) interpretations of the rock record.

The lack of knowledge with respect to the surficial, sub-surface, and facies characteristics of the sand sheets

primarily has resulted from the difficulty involved in sampling the laterally extensive sand sheet. Consequently, most sand sheet analyses have relied on morphological and surficial sedimentological characteristics. Fenster (1995), however, did construct an evolutionary model and classification of tidedominated sand sheets based, in part, on a regional assessment of modern-day surficial, internal, and boundary sand sheet characteristics (Table 1). More work is needed to define:

- the degree to which sand sheets incorporate relict units
- the three-dimensional distribution of subfacies within a sheet
- the controls on sheet shape (thickness and lateral continuity)
- lateral and bottom boundary characteristics, and
- within sheet characteristics (e.g., extent and continuity of unconformities).

Cross-References

- [Coastal Sedimentary Facies](#)
- [Continental Shelves](#)
- [Ingression, Regression, and Transgression](#)
- [Nearshore Sediment Transport Measurement](#)
- [Offshore Sand Banks and Linear Sand Ridges](#)
- [Ripple Marks](#)
- [Sandy Coasts](#)
- [Scour and Burial of Objects in Shallow Water](#)
- [Sequence Stratigraphy](#)
- [Shelf Processes](#)
- [Shoreface](#)

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Oil Spills

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Definition

The term “oil spill” refers to the accidental release of liquid petroleum hydrocarbons to the environment. The visual nature of black oil coming ashore from a spill commonly attracts public interest on a national and even international scale, oftentimes falsely projecting damages far greater than actually occur. Of all areas, however, long-term damage from oil spillage is likely to be greatest within the ecologically rich coastal zone. Additionally, coastal tides and currents

commonly increase the total extent of shoreline impacts as oil is transported far from the originating spill site, e.g., >2,000 km at *Exxon Valdez* in Alaska (Owens 1991) and 1,700 km at the *Deepwater Horizon* in the Gulf of Mexico (Michel et al. 2013).

References and guidance material are provided in the online proceedings of the International Oil Spill Conference (www.ioscproceedings.org) and from NOAA (<http://response.restoration.noaa.gov/oil-and-chemical-spills/oil-spills>), ITOPF (<http://www.itopf.com/knowledge-resources/>), IPIECA (<http://www.ipieca.org/resources/>), and CEDRE (<http://www.cedre.fr/en/>).

Quantity of Oil Entering the Marine Environment

By far the greatest source of oil entering the marine environment is from natural seeps and runoff from land during the consumption of petroleum products (Table 1). The extraction of petroleum (platforms, atmospheric deposition, and produced waters) accounts for approximately 2.2% worldwide, and discharges from transport (coastal facilities, tank vessels, pipelines, and atmospheric) are 11.9% of the worldwide total and less than 4% in North America due to tighter regulations. Oil released to the marine environment has almost been halved since 1990 when GESAMP (1993) estimated 2.3 million metric tons.

Historically, there has been significant improvement in the prevention of accidental hydrocarbon discharges, particularly those from tanker vessels. ITOPF data (<http://www.itopf.com/knowledge-resources/data-statistics/statistics/>) show a decline in the average annual number of large spills (>700 tons) on a decade basis from 24.5 spills per year in 1970–1789 to 1.7 spills from 2010 to 2016. The number of medium-sized spills (7–700 tons) has similarly decreased, from 28.1 in the 1990s to 5.0 spills annually from 2010 to 2016. These reductions are in spite of a doubling in the amount of petroleum being transported since the 1970s.

Oil Spills, Table 1 Sources of oil entering the marine environment, 1990–1999 data (NRC 2003)

	North America		Worldwide	
	Ton × 1,000	% of total	Ton × 1,000	% of total
Natural seeps	160	62.4	600	47.7
Extraction of petroleum	3	1.2	28	2.2
Petroleum transport	9.1	3.6	150	11.9
Petroleum consumption	84	32.8	480	38.2
Total	256.1	100	1,258	100
NAS rounded value	260		1,300	

Among many factors, a strengthening of international conventions (e.g., <http://www.imo.org>) and an increase in financial compensation requirements (<http://www.iopcfunds.org>) have contributed to safer petroleum transport.

Hydrocarbon Chemistry

Oil is composed of a complex mixture of hydrocarbons that varies highly by its source and subsequent refining. Once spilled, it is exposed to additional chemical and biological processes which further alter its composition.

The hydrocarbon component of oil is comprised of hundreds of organic compounds, principally divided into aliphatics and aromatics. In addition to hydrocarbons, oil may also contain small amounts of N, S, and O, as well as widely varying concentrations of trace metals (e.g., V, Ni, Fe, Al, NA, Ca, Cu, and U). These metals and other unique components of the spilled oil are often used as a tracer to follow the distribution and long-term impacts of the spill (oil spill fingerprinting, e.g., Bayona et al. 2015).

As crude oil is refined, the resulting product markedly changes in composition and physical properties (e.g., specific gravity, viscosity, pour point, and flash point), all of which influence the potential impact to the environment should it be released. The distillation process separates crude oil into refined products used for domestic and industrial purposes. These may be referred to as light fractions (gasoline), middle distillates (kerosene, heating oil, jet fuel, and light gas oil), wide-cut gas oil (lube oil and heavy gas oil), and residuum or bunker oil.

For oil spill response planning, oils are commonly categorized into four groups (ITOPF 2011) in terms of increasing specific gravity and viscosity: Group 1, gasoline, naphtha, and kerosene; Group 2, light crude oils and gas oil; Group 3, intermediate crude oils and medium fuel oils; and Group 4, heavy crudes and fuel oils (e.g., bunker oils). Group 4 oils include products that sink or have a buoyancy near that of water, greatly increasing the difficulty of recovery if spilled.

Spill Weathering and Fate

Once released to the environment, spilled oil is exposed to a series of “weathering” processes that alter its original chemistry and physical properties. These include spreading, advection, evaporation, emulsification, dissolution, dispersion, photooxidation, sedimentation, and biodegradation.

Spreading causes the formation of a relatively thin layer of oil on the water’s surface after it is released. Affecting processes include gravity, friction, viscosity, and surface tension. The color of the oil on the water is related to thickness and has been used as a simple measure of the quantity of oil present (Bonn Agreement 2011). For example, a silver sheen has a maximum thickness of 0.0003 mm and represents 300 L spread over 1 km². When slicks are forced to the coastline by onshore winds, the slick may increase to a centimeter or more in thickness.

Advection is the movement of oil on the water’s surface, influenced by currents and winds. Most predictions assume that oil will move in the direction and speed of the vector addition of current speed and direction plus 3% of wind velocity in the direction of the wind forcing. Shifting tides, eddies, and variable winds all add to the difficulty in predicting the real-time movement of the spill.

Evaporation is a major process influencing all aspects of oil chemistry and its physical properties. During evaporation, the light products (gasoline, kerosene, etc.) escape to the atmosphere leaving a lesser amount of oil on the water, but increasing the specific gravity and viscosity of the remaining oil. Evaporation can commonly remove 30–50% of light crude oils and refined products within a relatively short time (hours to days).

While evaporation can substantially reduce the quantity of oil, the process of emulsification (a water-in-oil mixture) may significantly increase and even double the quantity of total oil mass by incorporating water within the oil structure. Emulsification commonly occurs by wave action and mixing within the water column.

Dissolution is the transference of oil from the water’s surface into solution within the water column. It is very minor accounting for 1% or less of the total oil spilled.

Dispersion is the process of forming oil particles or droplets in the water column as occurs during breaking wave conditions. The process of forming particles of increasing smaller size enables the material to be more susceptible to biodegradation.

Photooxidation is the result of sunlight altering the composition of the oil. It accounts for only a small change (<1%) in the overall content of the spill. Direct effects of photooxidation on an actual spill have not been measured.

Sedimentation is the interaction of sedimentary particles and the spilled oil which may increase the density of the oil and carry it to the bottom. However, as most oils float, transport of substantial quantities to bottom sediments is relatively uncommon.

Biodegradation is based on the ability of bacteria and other organisms to degrade and convert the oil from a hydrocarbon to its fundamental elements. Biodegradation is able to act on small oil particles as the material is dispersed into the aquatic environment or when resident on shorelines. Oxygen and an adequate supply of water and nutrients are needed to sustain biodegradation sufficient to act on larger oil quantities.

Impact and Reaction in Coastal Environments

Spilled oil can have a significant impact on the coastal environment, the level of which varies with physical processes (e.g., winds, tides, and currents) and with the geomorphology and ecology (including human habitation) of the coastline and nearshore habitats (Gundlach 2013).

Oil Spills, Fig. 1 Overview of oil-killed mangroves south of Bodo, Rivers State, Nigeria, on 15 August 2015 with insert showing highly contaminated sediments at the site (scale = 15 cm) (Photograph is by the author)



Coasts have been categorized, ranked, and mapped in terms of sensitivity to oil spills by Gundlach and Hayes (1978). This system, called the Environmental Sensitivity Index (ESI), is internationally used for spill contingency planning and to assist the response effort during spills (NOAA 2012; IPIECA 2012). The ESI shoreline ranking indicates that exposed rocky shores are the least sensitive to spilled oil, followed by beaches of varying sediment sizes (sands to gravels), and ending with the most spill-sensitive sheltered tidal flats, marshes, and mangroves. Recent surveys in the Niger Delta have found large-scale losses of mangroves and the incorporation of oil into base sediments (Fig. 1). The ESI and others (e.g., Duke 2016) predict oil impacts lasting 30 years or more in mangrove habitats.

Response Methods

Response activities to contain and recover spilled oil have not changed dramatically over the past 25 years but have increased in greater efficiency. The vast majority of oil spills utilize a combination of mechanical and manual measures. Mechanical refers to the use of oil spill boom for shoreline protection or to divert oil, and the use of equipment such as skimmers or vacuum pumps to recover spilled oil. Manual recovery includes all aspects of physical labor, including shoveling, raking, hand wiping, and collection using sorbent materials.

Alternative response methods are also available. Dispersants are the most commonly used alternative method. Most modern dispersants are essentially nontoxic and under appropriate conditions can enhance the dispersion of the slick into

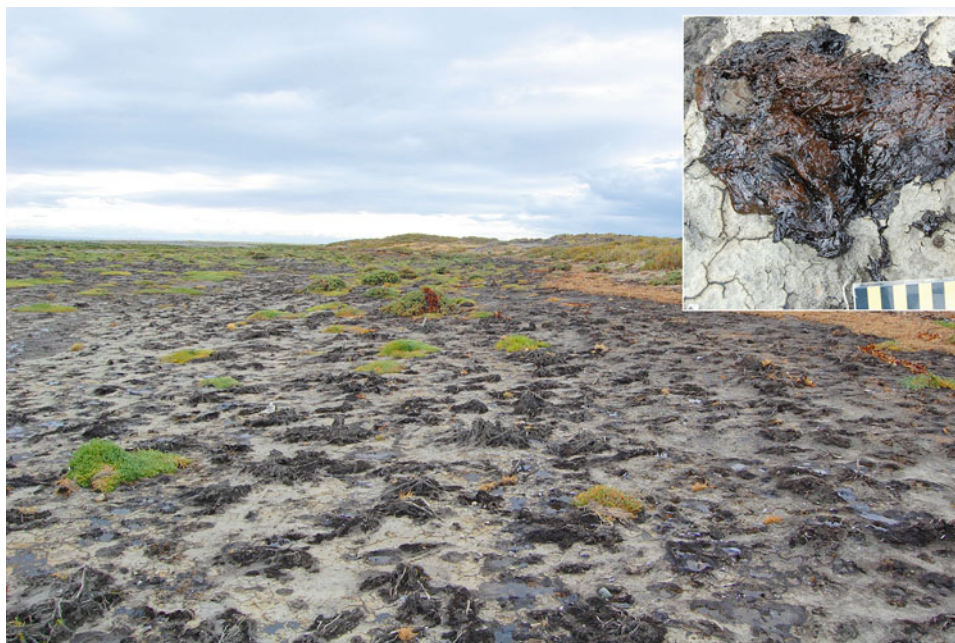
the water column, thereby making it available for biodegradation. Dispersant application is usually limited to deeper, offshore waters. Other agents are available for changing the oil's properties or "lifting" it from solid surfaces. In situ burning, previously rare in use, has gained great attention following its successful and widespread use at the *Deepwater Horizon*. After the gross contamination has been removed by traditional means, the application of fertilizers to "biostimulate" existing bacteria may be used to augment naturally occurring bacterial populations which then degrade the remaining oil. Each alternative method has its place and, when combined with mechanical and manual recovery, provides a range of options to the spill responder. The application of a net environmental benefit analysis (NEBA) assists in scientifically determining a cleanup endpoint and avoids causing more harm than good (IPIECA 2015).

Worst- and best-case calculations from the *Deepwater Horizon* show that 10–29% of the total oil was chemically dispersed and 5–6% burned (FISG 2010). These are very substantial quantities considering that over 700,000 tons of oil was discharged from the well blowout.

Spill Persistence

The longevity of oil in the water's surface is estimated by ITOPF (2011) to extend from hours (light Group 1 oils) to possibly longer than a year (heavy Group 4 oils). Once reaching shore, it has a much greater potential for persistence in spite of weathering and the best intentions of cleanup operations. Page et al. (2013) identified a series of large-scale (headlands, boulder layers sheltering underlying oil residues) and small-scale influences (low permeability sediments inhibiting water flow) on persistence from study at the

Oil Spills, Fig. 2 Overview of East Espora Marsh at the *Metula* spill site 8 February 2015 showing tar and mousse (oil-in-water emulsion) remaining with limited plant regrowth nearly 41 years after the event. Scale in the insert is 8 cm (Photograph is by the author)



Exxon Valdez spill site. Results show buried oil after 20 years remaining in isolated sediments and not being released to the surrounding waters.

Other examples are illustrated by oil remaining at the West Falmouth and *Arrow* spill sites after 35 years (Peacock et al. 2005; Owens et al. 2008) and at the *Metula* incident in southern Chile 41 years later (Fig. 2, Gundlach 2017). In all cases, oil remains only where it is sheltered from physical, chemical, and biological degradation.

Summary

Oil spills, while decreasing in frequency, remain a continuing threat to the marine environment. Coastal habitats, with high species richness and diversity, remain the most at threat. Advances in response techniques, as utilized during the *Deep Water Horizon* incident, have centered on the increased use of in situ burning and dispersants. Spills in sheltered habitats have the potential for lasting several decades. After 41 years, the *Metula* incident in southern Chile still shows large quantities of surface and subsurface oil that potentially will remain for several more decades.

Cross-References

- Beach Sediment Characteristics
- Beaufort Wind Scale
- Environmental Quality
- Human Impact on Coasts

- Marine Debris-Onshore, Offshore, and Seafloor Litter
- Rating Beaches
- Water Quality

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Oil Spills, High-Energy Coasts

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Most oil spills at sea occur in relatively confined and sheltered localities, often from tankers in passage to or

from oil terminals and ports. Natural processes, especially evaporation and other weathering effects quickly reduce the volume of oil at the surface but the more persistent residue requires expensive protection and clean-up procedures. Although normally less than is predicted at the time of the incident, substantial environmental and economic damage can occur along impacted coastlines and in nearshore zones. The Exxon Valdez oil spill in Alaska is the type-example of a large oil spill in a confined, low-energy location.

Some major oil spills occur in deep water more than 100 nautical miles from the nearest coastline. Normally weathering processes and, depending on wind and currents, the substantial time interval before the oil reaches a coastline minimize pollution damage. Chemical dispersants which are designed to breakdown the oil into tiny droplets for easier dilution, transport, and removal by microorganisms can be used but some nations prohibit their use nearshore or in confined sea and estuarine areas.

The *Braer* oil spill off Sumburgh Head in Shetland Islands, in 1993 epitomized a different type of oil spill when a large tanker ran aground on a high energy, rocky coastline (Fig. 1). The *Sea Empress* oil spill off the coast of Wales, in 1996 also occurred in a moderately exposed coastline where strong tidal currents were also of considerable importance in the initial stage. In spite of the volume of oil involved and the initial “disaster” reaction to the grounding of the *Braer*, the true ecological and economic impacts were not substantial. The combination of gale-force winds, the incident energy of massive storm waves on the open Atlantic coastline and the strongly dissipative nature of the irregular cliff coastline led to rapid dispersal into a highly turbulent water column. Tidal and wave currents carried this oil, at depth, considerable distances offshore and along-shore. There was virtually no stranding of oil and the characteristic surface oil slick could not form due to the prevailing storm conditions. An exposed sand beach lies less than 1 km east of the position of grounding but, due to the power of backwash from the steep beach face, the oil and water mixture did not contaminate the intertidal zone. Although less than 1% of oil was involved, there were also local spray and aerosol effects due to the severity of wave action on the coastline.

The *Braer* can be used to indicate that an oil spill on an exposed, high-energy coastline, especially a rocky and cliff coastline is unlikely to cause much environmental damage. Under such conditions, oil particles are likely to be transferred to the water column and, in time, lodged on the sea bed, perhaps at considerable distances offshore as a consequence of local and regional currents and water movements. Thus an oil spill on a high-energy coastline is more likely to create



Oil Spills, High-Energy Coasts, Fig. 1 Wreck of the oil tanker Braer on the south coast of the Shetland Islands in January 1993. (Photograph kindly supplied by Mr. P. Fisher)

offshore, marine problems rather than to produce negative effects on the adjacent shore. In short, the critical difference is the presence or absence of a surface oil slick which is a consequence of local shelter and energy conditions at the time of the accident.

Cross-References

- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Marine Debris-Onshore, Offshore, and Seafloor Litter](#)
- [Rock Coast Processes](#)
- [Water Quality](#)

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Open-Coast Tidal Flat Deposits

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Definitions and Distribution

Although tidal flat deposits have been studied extensively over several decades, most studies were focused on embayment and estuary tidal flats, where wave energy is typically low. Open-coast tidal flats, which differ significantly from the embayment and estuary tidal flats, are not well studied until recently. Here we describe the morphological and sedimentological characteristics of open-coast tidal flats in comparison with the embayment tidal flats.

A tidal flat is generally defined as an extensive, nearly horizontal, marshy, or barren tract of land that is alternately covered and uncovered by the tide, consisting of unconsolidated sediment, mostly mud and sand (Bates and Jackson 1980). Tidal flat is also often referred to as intertidal zone, although in some general discussions, it may include subtidal and supratidal zones. Here, the term tidal flat strictly refers to

the intertidal zone, lying between spring low tide and spring high tide levels, as defined in the *Glossary of Geology* (Bates and Jackson 1980). A tidal flat generally shows a zonation related to the duration of submergence, which is reflected in differences in sedimentary characteristics. Therefore, based on characteristics of sediments and sedimentary structures, a tidal flat is often divided into upper, middle, and lower intertidal zones (Reineck and Singh 1980). Although the transitions between the zones are gradual and somewhat subjective, the division provides convenience in describing sedimentary characteristics.

Tidal currents are generally considered to be the dominant driving force for sediment movement on tidal flats, whereas wave-driven sediment motion is often limited to wave ripples. Eisma (1998) estimated that over 70% of locations classified as tidal flats occur in wave-sheltered areas, such as estuaries, while the remainder occur along open coasts, the majority of which are characterized by low wave energy. In a more recent study, Fan (2012) argued that due to their association with large river deltas, open-coast tidal flats are more widely distributed than semi-enclosed ones. Although high wave energy is often considered to be unfavorable for the development of gentle tidal flats (Boggs 1995), they can nonetheless develop extensively along open coast under the condition of

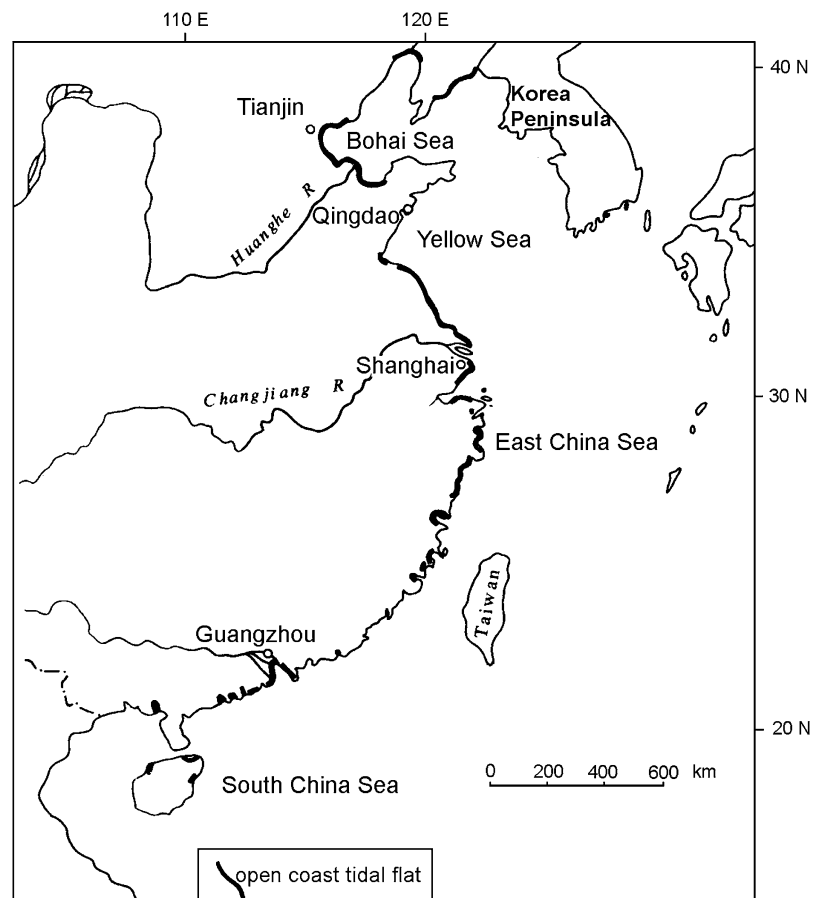
tremendous supply of fine grain sediment (Fan 2012). Examples of open-coast tidal flats are found in the vicinity of the large river mouths along the Chinese coast (Fig. 1; Chen 1998; Shi and Chen 1996).

Hydrodynamics and Sediment Dynamics

Many studies on modern tidal flat deposition have concentrated on wave-sheltered areas in Europe and North America (e.g., Klein 1976; Boersma and Terwindt 1981; Dalrymple et al. 1991; Allen and Duffy 1998; Eisma 1998). Consequently, sedimentary characteristics, such as mega-ripples/dunes, herringbone cross-bedding and reactivation surface, etc., associated with tidal currents flowing through tidal channels have been well documented. In comparison with wave-sheltered tidal flats, open-coast tidal flats are characterized by (1) facing an open ocean or sea and (2) flooding and ebbing tidal currents being not confined and/or regulated by tidal channels (Wang and Cheng 2017). Therefore, the aforementioned sedimentary features are rare to nonexistent.

Wave energy is significantly dissipated over the extensive and gentle muddy flats under normal weather. However, during storms, due to the lack of wave shelter, open-coast

Open-Coast Tidal Flat Deposits, Fig. 1 Distribution of open-coast tidal flats along the coast of China. The tidal flats along the Changjiang River delta discussed here are in the middle of the figure



tidal flats are subject to energetic wave actions, which may induce substantial rework and redeposition of normal weather tidal deposits.

Sediment dynamics along open-coast tidal flats carry strong regional characteristics. The sediment dynamics along the open-coast tidal flats along the Changjiang River delta (Fig. 1), which are used here as an example, are significantly influenced by the tremendous fine sediment supplies from the large rivers. The tidal flats tend to be accretional owing to the abundant fine (silt and clay) riverine sediment supply. Coastal erosion is typically caused by depletion of sediment supply due to switching of river mouths, artificial flood controls, or drainage withdrawal (Fan 2012). Mean sediment grain size ranges from 4 to 8 ϕ (0.063 mm to 0.004 mm). The fine nature of the sediment also contributes to the absence of mega-ripples in addition to the general lack of tidal channels. The width of the tidal flats ranges from 3 to 4 km with a maximum of about 8 km. The average slope of the tidal flat is typically 1:1000 with a maximum of 1:200 and a minimum of 1:5000.

Characteristics of Sedimentary Structures

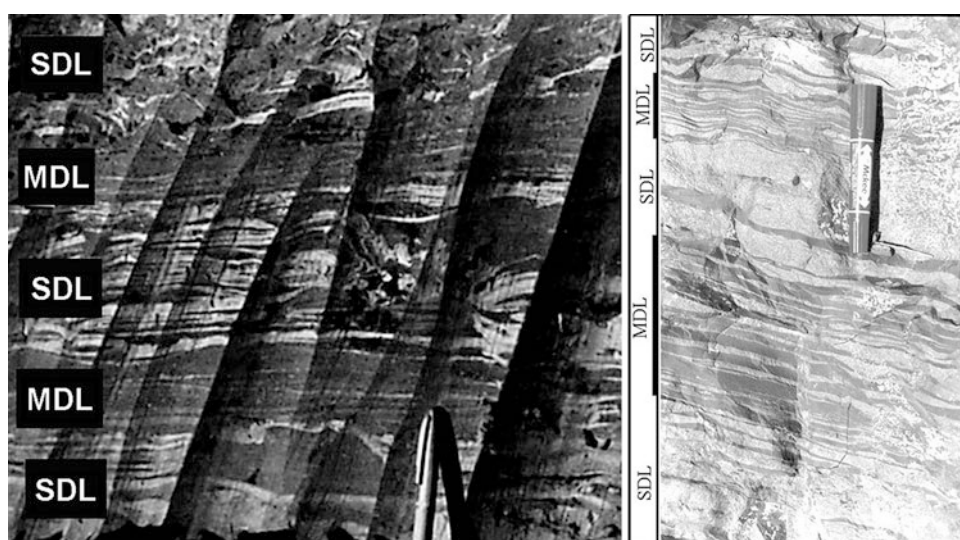
The general sequence of sedimentary structure variation from upper to lower intertidal zones as described by Reineck and Singh (1980) was also observed on open-coast tidal flats. The upper intertidal zone is characterized by relatively finer sediment with thicker muddy laminae. Lenticular bedding is common in the upper intertidal zone. The lower intertidal zone is characteristic of coarser sediment and thicker sandy laminae. Flaser bedding is common in the lower intertidal zone. Wavy bedding is common in the middle intertidal zone.

It is generally accepted that four laminae may theoretically be deposited during one tidal cycle (Allen 1985). Two sandy

laminae can be formed during flooding and ebbing phases and two muddy laminae deposited during high and low tide slack water. It was also found that thicker sand laminae correspond to relatively higher energy events during spring tides, while thinner sand laminae correspond to relatively low energy events during neap tides (Boersma and Terwindt 1981; Allen 1985). Time series analysis of laminae thickness has been applied to quantify the palaeo-tide periodicities and palaeo-sedimentation rates in both modern and ancient tidal deposits (Yang and Nio 1985; Tessier and Gigot 1989; Tessier 1993; Kvale et al. 1989; Kuecher et al. 1990; Kvale and Archer 1990; Miller and Eriksson 1997), and an approximate 14-day periodicity related to neap-spring cycles has been identified in most of these studies. In the above studies, deposition and erosion induced by waves were not considered. Furthermore, 100% preservation of the deposition during the spring-neap cycle was implied. Along open-coast tidal flats, four laminae were rarely observed to have been deposited during one tidal cycle due to their poor preservation potential. Over a continuous daily observation of 8 months, the preservation of all four laminae after one tidal cycle was only recorded twice (Li et al. 1965).

Two different grouping patterns of sandy and muddy laminae were distinguished on open-coast tidal flat deposits (Fig. 2). Groups with generally thicker sandy laminae than adjacent groups are termed as sand-dominated layers (SDL), while groups with generally thinner sandy laminae than adjacent groups are referred to as mud-dominated layers (MDL). Although determination of the exact boundaries between sand- and mud-dominated layers can be somewhat subjective, the overall differences between adjacent SDL and MDL are apparent (Fig. 2). The thickness and number of sandy and muddy laminae in each sand- or mud-dominated layer were not necessarily identical. Detailed description of sand- and mud-dominated layers can be found in Li et al. (2000).

Open-Coast Tidal Flat Deposits, Fig. 2 Different grouping patterns of tidal bedding. SDL denotes a sand-dominated layer, and MDL denotes a mud-dominated layer



Daily, monthly, and yearly sedimentation monitoring along an open-coast tidal flat near the Changjiang River mouth indicated that the MDL described above correspond to calm weather deposition, while the SDL are related to high-energy storm events. These findings are contrary to the neap-spring cycle interpretation of lamina thickness variations. The above interpretation of event-related lamina thickness variation incorporated the significant influence of waves, especially storm waves, on the open-coast tidal flat deposits. High-energy storm events along the east-central Chinese coasts are mainly driven by the typhoon passages typically during the months of August to November and to a lesser extent the passages of winter cold fronts. Therefore, the alternation of SDL and MDL may reflect the variation of storm and calm weather seasons, instead of spring and neap tidal cycles.

Sedimentation Rate and Preservation Potential

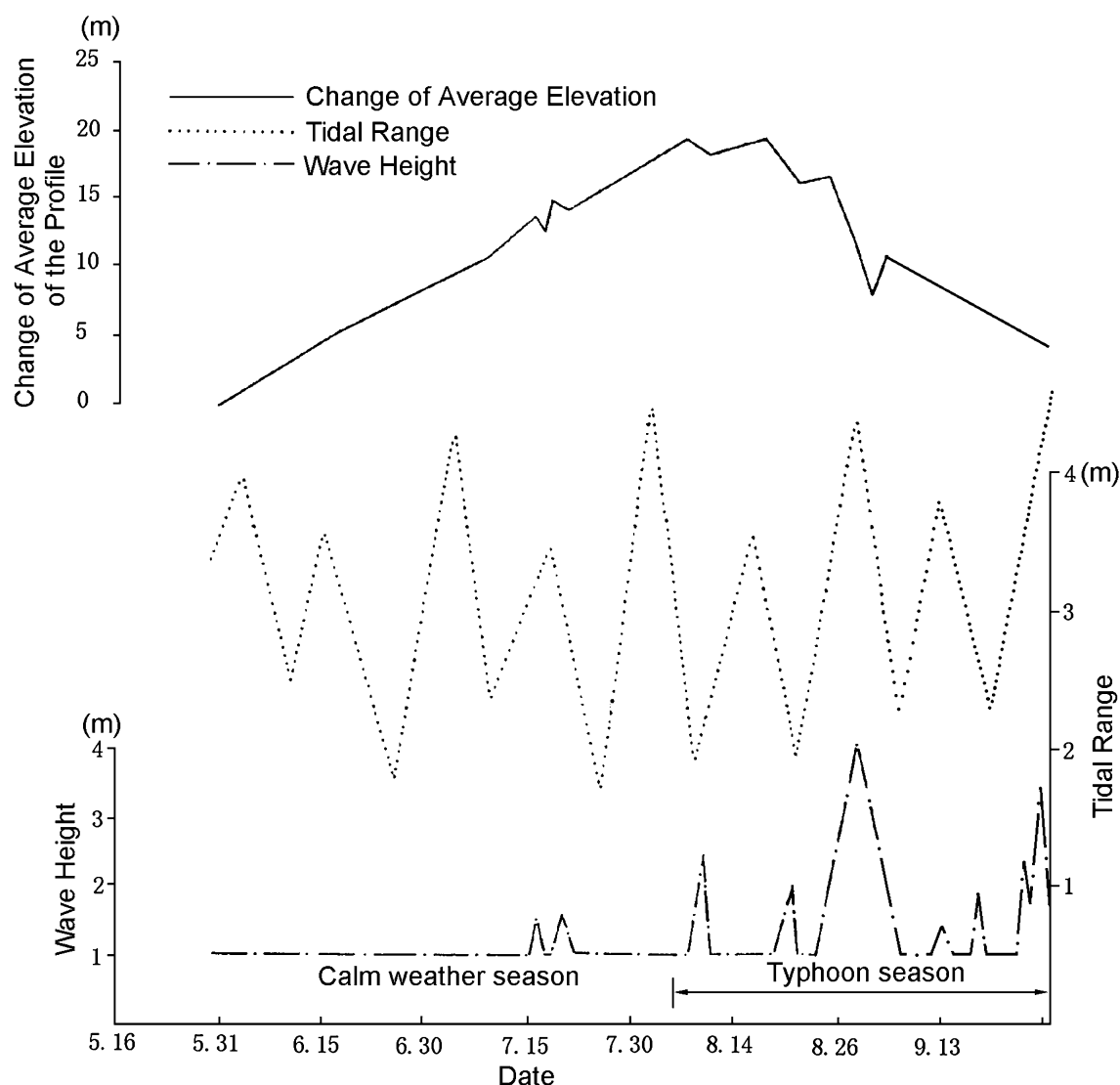
The correspondence of thick-thin variations of tidal laminae with the spring-neap tidal cycles has provided a promising tool to study the paleo-tide periodicities and the paleo-sedimentation rate (Tessier and Gigot 1989; Miller and Eriksson 1997). Time series analysis of the tidal bundle thickness variation has been applied successfully in tidal channel deposits (Yang and Nio 1985). However, direct application of this neap-spring analysis may not be valid along open-coast tidal flats due to the non-negligible impact of wave-induced erosion and sedimentation. Sedimentation rate and preservation potential vary from region to region and are strongly influenced by sediment supplies and regional hydrodynamics. In the following, a case study from Changjiang River delta is discussed as an example.

Realizing that 100% preservation is unrealistic on an open-coast tidal flat, Li et al. (2000) conducted an intensive monitoring of the sedimentation rate and preservation potential. Their methodology included two aspects to understand and quantify the deposition and preservation of tidal bedding. The first aspect emphasized observations on the modern tidal flat. Selected points were visited daily during low tide to examine deposition and erosion of the previous two tidal cycles. This daily observation had been conducted for 17 days over a neap-spring tidal cycle. Seasonal sedimentation and erosion observations across an intertidal profile were conducted over a 4-month period extending from a calm weather season to a storm season. The second aspect of the study focused on examining the vertical characteristics of tidal laminae and bedding, in terms of the lamina numbers and lamina thickness variations. Knowledge gained from the field observations was applied to interpret the vertical distribution of tidal bedding and the preservation of individual sandy and muddy laminae in cores and trenches.

Short-term observation was conducted using two thin plastic plates (40 cm long, 40 cm wide, and 2 mm thick). The plates were placed in the transition area between middle and upper intertidal zones. The thin plates were placed flush with the sediment surface, and their surfaces were sanded to increase the roughness. Plate 1 was left in place for a period of 17 days, and one observation was made at the end of the experiment. The objective was to obtain short-term information without daily disturbance. Deposits on plate 2 were measured daily or every 2 days in cases of poor weather conditions. The objective was to quantify daily sedimentation rate and the number of laminae formed.

The daily monitoring spanned one neap-spring tidal cycle. No apparent trend of lamina thickness variation was observed from the daily plate experiments. The daily sedimentation rates were rather uniform ranging from 17 to 22 mm/day, except for two abnormal values of 12 and 45 mm/day, both measured during spring tides. The calculated cumulative 17-day deposition was 378 mm with 50 sandy and muddy laminae, while 12 laminae with a total thickness of 75 mm were measured on the 17-day plate. Thus, the uninterrupted 17-day sedimentation was 20% of the cumulative daily deposition in terms of thickness (75 mm vs. 378 mm) or 24% in terms of total number of laminae (12 vs. 50). Furthermore, comparison between the results from the 17-day monitoring (12 laminae) and the theoretical estimate (4 laminae per tidal cycle for 33 tidal cycles over 17 days) of 132 laminae indicates that approximately 9% of the theoretical laminae were preserved, during a period of one neap-spring tidal cycle. Continuous daily observation of tidal lamina number and thickness, conducted over three neap-spring cycles in 1999, yielded similar preservation potential.

Seasonal monitoring spanned 4 months with 2 months at the end of a calm weather season and 2 months at the beginning of the following storm season. Sedimentation and/or erosion were measured at 35 locations relative to a series of graduated rods. A total of 22 measurements were conducted during the 4-month period. Overall, the intertidal zone was accretionary during the calm weather season (Fig. 3), as indicated by the increasing average elevation at all the rods. During the storm season, net erosion (elevation decrease) was measured, and the flat was covered by a sandy lamina that is much thicker than those deposited during the calm weather season. Sharp elevation decrease was usually measured directly after the storms, indicating the erosion caused by storm waves. For the convenience of discussion, the calm and storm weather season was divided subjectively by the first significant typhoon impact in the study area. The field observations throughout the 4-month period indicated that the tidal flat was generally muddier in the calm weather season than during the stormy season. The changes of average elevation were related not to the neap-spring tidal cycles, but to high wave events (Fig. 3). Deposition of a relatively thicker



Open-Coast Tidal Flat Deposits, Fig. 3 Elevation change at Donghai Farm tidal flat on the southern flank of the Changjiang River delta and corresponding tide and wave conditions. Wave heights lower than 1 m were not included. The study was conducted in 1992

sandy lamina was directly related to a high-energy wave event induced by the passage of a typhoon, instead of during spring tide conditions.

Long-term sedimentation rate and preservation potential were determined through counting the number of laminae and SDL-MDL in core sections that were deposited over 100 years. Details are described in Li et al. (2000). The centennial sedimentation rate was found to be on the order of 4 cm per year. It is worth emphasizing that the high sedimentation rate is related to the tremendous sediment supply from the adjacent river. Such high sedimentation rate should not be expected at locations without huge sediment supply. Over the 100-year period, the preservation potential of individual lamina, including both calm weather and storm deposits, was found to be on the order of 0.2%, which was 45 times smaller than the 9% estimated for a short term of a

neap-spring cycle. It is expected that the preservation potential decreases as temporal interval increases. The 100-year preservation potential of storm-induced SDL was estimated to be of the order of 10%, 50 times higher than the overall potential of 0.2%.

Summary

Waves, especially high storm waves, have a significant influence on sedimentation and preservation of intertidal deposits along the open-coast tidal flats. The thickness variation of sandy laminae on an open-coast tidal flat is related not to neap-spring tidal cycles, but directly to storm activities. The mud-dominated layers containing thinner sandy laminae were deposited during calm weather conditions, while the

sand-dominated layers containing thicker sandy laminae were deposited during storm seasons. The thick-thin variation of sandy laminae may reflect a much longer cycle of calm weather and storm seasons, instead of the fortnightly spring-neap tidal cycles as suggested from the study on wave-sheltered tidal flats.

One hundred percent preservation of both the number and thickness of individual laminae in tidal flat deposits, which has been assumed in the interpretation of time series analysis of laminae-thickness variation, was found to be unrealistic along open-coast tidal flats. Preservation potential decreases as time scale increases. During one neap-spring tidal cycle under calm weather conditions, the preservation potential of individual lamina was approximately 9%. However, over a period of 100 years, the preservation of individual lamina decreased to about 0.2%. The preservation rate of storm-induced, sand-dominated layers during the 100-year period was found to be on the order of 10%, much greater than the 0.2% of the individual sandy and muddy lamina. Storm deposits have a much greater potential for being preserved due to the higher energy level at which they were deposited.

Cross-References

- [Asia, Eastern, Coastal Geomorphology](#)
- [Deltas](#)
- [Tidal Environments](#)
- [Tidal Flats](#)
- [Tides](#)
- [Waves](#)

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Pacific Ocean Islands, Coastal Ecology

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The ecology of the tropics is an incredibly complex subject, probably far too complex to be grasped in its entirety by the human mind. Yet it is of the greatest urgency that it should be sufficiently understood so as to help man utilize tropical lands without utterly destroying the region eventually as a human habitat.

Marie-Helen Sachet, 1967, Botanist, Smithsonian Institute (cf. National Biodiversity Team of the Republic of the Marshall Islands, 2000, inside front cover)

Adequate comprehension of the complexity of the vast number of relationships between organisms and their environment should be extended to both land and sea, especially where they are in relatively close proximity near the coastline.

Coastal ecology of isolated tropical Pacific islands offers many microcosmic examples of the forms and functions of species. It also focuses attention on the adaptation to a series of biogeochemical conditions presented by natural phenomena, and more recently by human activities. The ecosystems addressed in this selective review include coral reefs, marine, and freshwater wetlands, and coastal strand and forest. All of these can be found at, or relatively near tropical island coastlines within the world's largest single feature, the Pacific Ocean.

Ecological relationships along the coasts of Pacific Islands have been undergoing dynamic adjustment to tectonic and climatic changes at least throughout the last 1.8 million years (the Quaternary period). This synchronous and time-transgressive environmental change has continued into the Holocene Epoch, or recent postglacial warming period that began approximately 12,000 years BP. It is during this period that humans have had an increasing impact on the abiotic and biotic components of coastal ecosystems.

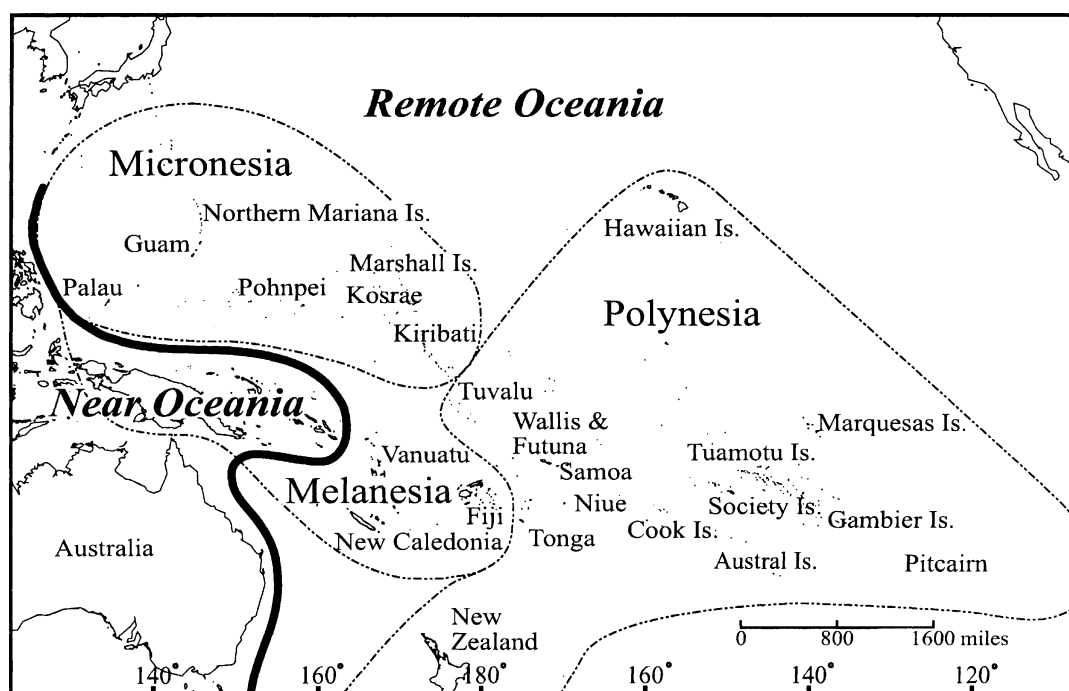
Here we focus on coastal ecology of Remote Oceania, which covers a vast area of the Pacific Ocean, and includes all those tropical Pacific Islands so isolated that humans were only able to find them within the past 4000 years (see Fig. 1). Within this huge region, the cultural impact of people locally, regionally, and perhaps globally must be, more or less, added to those natural factors which affect the changing nature of Pacific coastal ecology, such as volcanic eruptions, earthquakes, tsunami, and hurricanes (typhoons).

The tropical Pacific Islands of Remote Oceania are basically the subaerial portions of hot-spot volcanoes (produced by melting anomalies), or limestone caps of reef development that are near or above the surface of the sea. The range of Remote Oceanic Islands include:

1. High islands in various stages of volcanic development, geomorphological erosion, and subsidence or tectonic uplift (e.g., O'ahu, Nuku Hiva, Mo'orea, and Mangaia Islands);
2. Atolls, which normally consist of an annular-shaped series of coral reefs with few to many low-lying islets comprised of wave-washed, unconsolidated reef, and other fauna or calcareous algal debris, more or less, surrounding a lagoon with all of the volcanic formation sunk below the sea level (e.g., Ulithi, Tarawa, and Takapoto Atolls);
3. Raised Atolls or isolated reefs that have been uplifted or formed at higher stands of the sea, which are commonly referred to in the Pacific region as *makatea*, or raised reef islands (e.g., Fais, Nauru, Makatea, and Henderson Islands).

There are variants that, more or less, bridge the gaps between these islands categories (e.g., almost atolls such as Chuuk, Aitutaki, and Bora Bora Islands).

Thus the range of coastal ecosystems located near coastlines vary from those situated on the slopes of active volcanoes (e.g., some Hawaiian and Galapagos Islands), to those



Pacific Ocean Islands, Coastal Ecology, Fig. 1 Map of Near and Remote Oceania. (After Merlin 2000)

on older dormant or extinct volcanic high islands with well developed fringing and/or barrier reefs and few to many sandy beaches (e.g., Pohnpei, Kaua'i and Tahiti Islands), and those close to or along raised reefs (Atiu and Mitiaro Islands) and atoll shores (e.g., Ailinglaplap and Oeno Atolls).

Above and below the sea, a series of environments, or ecosystems, are distributed along tropical coastlines where inundation by sea water is continual, periodic, seasonal, or episodic in nature. These general environments include coral reefs, mangroves, freshwater wetlands, and strand communities. Their myriad inorganic and organic components interact with one another in wide variety of environmental relationships.

As volcanoes have subsided or been uplifted in concordance with climatically and/or tectonically controlled rising and lowering of the ocean level, those coastal species populations that have adapted to these environmental changes survived by effectively migrating in relationship to changing sea levels. In addition, environmental stresses produced by human use on land and in ocean have, more recently (during the past 1000–4000 years BP), also affected the distribution of species. It is generally assumed that these human impacts have accelerated in modern times; anthropogenic changes will be discussed below. Origins of the native species in the coastal ecosystems of the tropical islands in Remote Oceania are discussed first, followed by a description of the basic features of the ecosystems in which these species inhabit.

Origins of the Species: Biogeography and the Filter Effect

The ecosystem concept is a useful scientific form of reductionism designed to bring order to the diversity of life, including that found in the environments of the tropical coastal Pacific region. Whether or not ecosystems form and function in discrete, interrelated units is still an unresolved question for biologists. It has been argued that plants and animals in terrestrial ecosystems have independent ranges of distribution that often overlap in their adaptations to environmental parameters. This may also apply to marine species. Nevertheless, if only for systematic convenience, the ecosystem concept remains useful for scientific description and analysis.

It is clear that variable dispersal ability helps explain the distancedecay factor ("filter effect") determining which organisms can colonize remote geographic locations. Although many species adapted to tropical marine and nearshore land environments have long-distance dispersal mechanisms and are characteristically found over a vast region, there is consistent decline in the number of many species that become established in marine, and especially inland terrestrial environments over time in Remote Oceania (Carlquist 1974; see Fig. 2). For example, the total number of land birds on islands compared to the totals on neighboring islands and chains of islands in the South Pacific Region declines with distance east from the major source of immigrants in the New Guinea region, where there are 520 species, to 127 in the Solomons, 54 in Fiji, 33 in Samoa, and 17 in the Society Islands.

Pacific Ocean Islands, Coastal Ecology, Fig. 2 Buoyant fruits of *Barringtonia asiatica* float long distances and eventually wash ashore viable in coastal areas such as shown here in the Solomon Islands. (Photograph by Mark Merlin)



Indeed, a significant dropoff in the number of successful colonizing taxa can be verified for many groups of organisms found in tropical Pacific ecosystems, including those located at or near coastlines. For example, in the western tropical Pacific Islands of the Republic of Palau there are over 300 reef building coral species, while in remote Hawaii there are approximately 40 species (Grigg 1983). And the distribution of damselfish (Pomacentridae) in the tropical Pacific drops off from 140 in the New Guinea region to 104 species in Fiji, 98 in Samoa, 81 in Tahiti, and 50 in Easter Island (Stoddart 1992). Some of this oceanic barrier effect has been disrupted by humans who have brought many species (purposefully and unwittingly) with them over time, especially in the terrestrial island environments within the last two centuries (Mueller-Dombois and Fosberg 1998; Kirch 2000).

When distance from continents and other islands is combined with latitudinal proximity to the equator and rainfall distribution, the relative biodiversity of coastal ecosystems in Remote Oceania can be explained. The further a coastal habitat is from the biological diverse continents and other islands, along with its relative distance from the equator in combination with declining annual rainfall, generally the fewer the number of naturally occurring species will be found. And this is much more pronounced for inland, high island environments (Stoddart 1992).

Disturbance and Succession

Environmental conditions in the ocean, atmosphere and on land, in themselves and in combination create periodic, sometimes catastrophic changes. Volcanic activity, including

eruptions of both pyroclastics and lava, play a role in the periodic change of many ecosystems including those in the lowland coastal and nearshore environments. For example, basaltic lava flows erupting out of the active Kilauea volcano on the large island of Hawaii, often destroy coral reef formations as they enter the sea. However, the newly formed igneous surfaces provide the base for pioneer species to recolonize and initiate the successional processes. Over time and with limited disturbance, the ecosystem typically returns to a fully mature community of interacting organisms (Grigg and Maragos 1974).

The erosional forces of mass wasting, along with stream, wind, and wave action, shape and reshape the coastal environments, both gradually and episodically. The relatively slow pace of disturbance and ecological recovery through biological succession, is occasionally punctuated by very intense environmental perturbations. Great waves of water generated in the Pacific Ocean can greatly alter the coral reefs and coastal lands over relatively short periods of time. These huge swells are often induced by high-energy land movements. Large earthquakes and/or giant landslides normally occur along subduction zones, and sometimes near or within archipelagoes of hot-spot volcanoes, particularly those in the youthful subaerial shield building stage (Moore et al. 1994). Given these determinants of biodiversity according to geographical and climatic distribution, as well as ecological disturbance and succession, we can now review the coastal ecosystems or environments of Remote Oceania. These include coral reefs, seagrass beds, mangrove forests, coastal freshwater wetlands, swamps and marshes, strand communities, and coastal lowland forest.

Coral Reefs

Coral reefs are biologically diverse oases of high-energy flow and nutrient cycling. Major organism groups found on coral reefs include algae, seagrasses, corals, and other animals. Coralline algae are often equal or greater contributors to the calcium carbonate structure of the reefs than coral organisms. The zonation and profusion of life in coral reef communities is determined by environmental parameters such as distance and depth from the coastline. Wave exposure, gradients in sedimentation, degree of salinity, and temperature are the primary factors causing this variation in abundance and composition in coral reef communities in Remote Oceania. Corals can be viewed as organisms with specific behavior, physiology, and structure, as well as colonies that serve as habitat for a broad diversity of other organisms. Furthermore, coral reefs themselves may be seen as ecosystems with a great variety of interactions among the organic and inorganic components that function in these environments (see Fig. 3).

Over much of the tropical Pacific Ocean, the composition of coral reefs is frequently similar, with the genus *Acropora* often the major component of the bottom cover (Gulko 1998). Isolated and younger coral reefs in Remote Oceania such as those in the Hawaiian Islands, however, differ significantly from those in other tropical areas. In the case of the Hawaiian reefs, they are not as well developed because of their relatively youthful geological age. Therefore, most reefs in the younger islands are small fringing reefs. Wave exposure is also important. The leeward coasts of the Hawaiian islands, for example, which are generally more sheltered, have reefs with more coral cover than the wave-pounded windward coasts (Grigg 1983; Jokiel 1987).

The younger coral reefs, which lack barrier reefs (e.g., those in the windward Hawaiian Islands and some of the southern Cooks Islands) are normally less productive than other reefs associated with many other older high islands or atolls in Remote Oceania. Less reef area and the lack of lagoons that collect coastal and terrestrial runoff in the very remote, and younger, windward Hawaiian reefs results in relatively poor supplies of nutrients; this in turn helps explain the comparatively low numbers of soft corals, sponges, tunicates, bivalves, and other filterfeeding animals. Unlike many other reefs in Remote Oceania, especially those associated with older high islands or atolls, the reefs of Hawaii are noticeably dominated by corals, even though there are relatively few species of corals on these reefs.

As noted above, because of the remote location of the Hawaiian Islands the coral reefs have considerably less biodiversity than other reefs, particularly those located much closer to the Indo-Pacific region to which they belong. One effect of this lower diversity is reduced specialization of reef-building corals in the Hawaiian Islands, which as a result have more extensive distributions of individual species than in less-isolated regions. Although the biodiversity of marine species is much less in the more isolated coral reefs of Remote Oceania, these reefs, such as those in Hawaii have particularly high levels of endemic marine species.

Seagrass Beds

Seagrasses are flowering plants, and are dissimilar to algae in that they produce true roots, stems, leaves, flowers (often inconspicuous), and seeds. Seagrass beds or meadows are

Pacific Ocean Islands, Coastal Ecology, Fig. 3 Coral reef in Chuuk lagoon with reef islets in the background. (Photograph by Mark Merlin)



found in marine or estuarine waters continuously flooded by saltwater. Most seagrass species are rooted in silty or sandy sediments located in shallow water up to about 7 m deep. Below this depth there is not enough sunlight for the seagrasses to survive.

Worldwide there are approximately 49 species of seagrasses in 12 genera. They are classified as belonging to at least two monocotyledonous families, Hydrocharitaceae and Potamogetonaceae, which should not be confused with members of Poaceae, the grass family. Sixteen species of seagrasses are found in the Pacific Island region. However, little research has been done on seagrasses in this region, and some scientists suspect that new species remain to be described in the tropical Pacific (den Hartog 1970; Mathieson and Nienhuis 1991). Many species of marine algae, as well as numerous animals, are also found in seagrass areas.

Seagrass beds are generally located in a mixed species zone running parallel to shore, but normally separated from the mangrove vegetation by a narrow band of nonvegetated sand or silt. In some areas, the seagrass beds extend out over large areas of the fringing reef flat and often cover interior areas such as embayments.

Like the mangrove forests (described below), the seagrasses, with their dense rhizome and root systems, serve as traps for silts and sediments washed into the sea from land areas. Seagrass beds are extremely productive, providing food as well as shelter for a number of animal species, including sea cucumbers, clams, some fish such as siganids, dugongs, and some reptiles such as the green sea turtle (*Chelonia mydas*). Throughout its huge geographical range in the tropical oceans of the world, the green sea turtle commonly consumes species of seagrasses in the genera of *Thalassia*, *Cymodocea*, and *Halophila*, all of which are found along many coastal areas of the tropical Pacific Ocean.

Seagrass beds are very significant in coastal food webs; in fact they typically sustain more invertebrates and fish than nearby areas lacking seagrasses. A multitude of epiphytic algae and associated small animals live on seagrasses, and several kinds of mollusks, and some fish obtain most of their nutrition by feeding on these epiphytes. Substantial amounts of detritus form as microorganisms colonize dying seagrass tissue; and because of this most animals found in seagrass beds are detritivores (Bortone 2000).

Mangrove Forests

Mangrove forests are unique and successful adaptations to harsh environmental restraints. They are found in saltwater-influenced swamps, and are composed of woody plants (trees and shrubs) adapted to areas affected by ocean tides (Chapman 1976). These distinctive forest communities are located in many subtropical and tropical areas around the world, covering approximately 60–70% of all tropical coasts

(Por and Dor 1984). They are typically found on muddy reef flats of coastal areas.

Mangrove ecosystems are found between the latitudes of 32°N and 38°S, along the tropical coasts of Africa, Australia, Asia, the Americas, and many tropical oceanic islands. There are varying scientific classifications of what constitutes a mangrove plant depending upon how strictly the term “mangrove” is defined. The worldwide number of species classified as mangrove plants varies considerably (e.g., 54 according to Tomlinson 1986, and 75 according to Field 1995). The greatest diversity of mangrove species exists in Southeast Asia.

The biogeographic filter effect discussed above also applies to mangroves. For example, in the coastal saltwater environments of the Republic of Palau there are nine known species; further east in similar coastal environments of Yap there are only seven known species; in the Marshall Islands, much further to the east in the Pacific Ocean, there may be only three species, all of which are not very common; and much further east in the Hawaiian Islands there are no native species of mangroves.

The plants of the mangrove forests are very important for the environment (or ecosystem). They stabilize coastal areas by trapping and holding sediment washed down from inland areas. In some areas, where rainfall washes the thick mud from the interior down to the mangrove swamps, the mass of tangled roots of the woody plants prevent much of this sediment from washing out and smothering the coral reefs and seagrass beds. In those areas where the mangroves are located on the muddy reef flats, the plants of these forests also offer some coastal protection from normal wave action and strong storm surges (Lugo and Snedaker 1974, see Fig. 4).

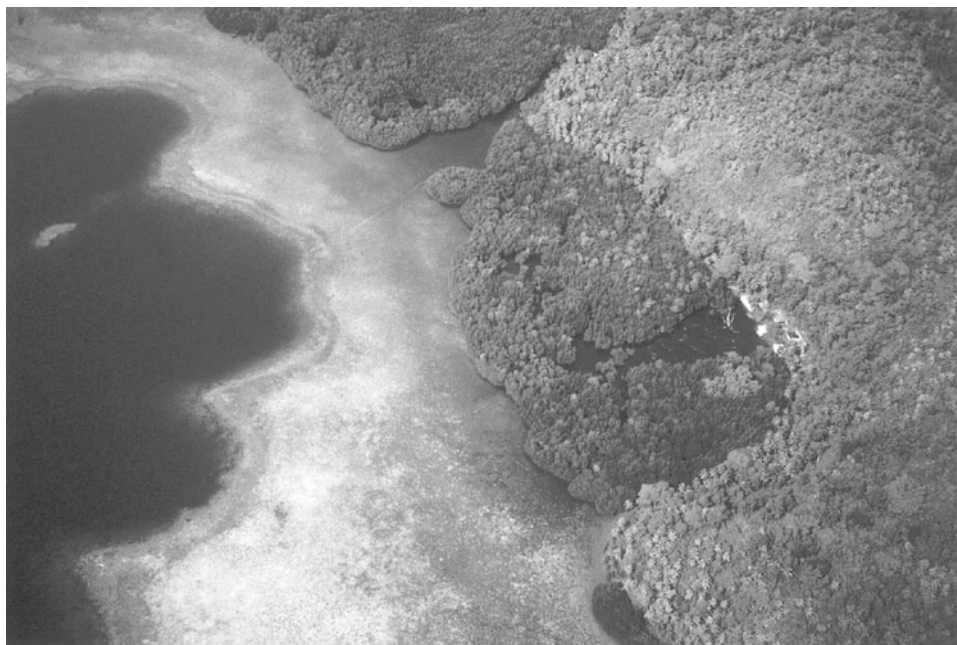
As “producers,” the mangrove plants convert energy from the sun, through the process of photosynthesis, into useful proteins and carbohydrates in their leaves and other parts of the plants. Eventually these plant parts drop off into the water and are carried out with the tides into nearshore waters forming a rich “nutrient soup.” This supplies food for nearshore fisheries and the reef animals. Mangroves also provide an important source of nourishment for many animal species that live within these marine swamps, including various types of shellfish, finfish, skinks, geckos, insects, and birds. In addition, several of these species use the mangroves as nurseries for their young.

All trees that live in the mangrove must have a means of obtaining oxygen since the tide comes in and water regularly covers the roots. Some woody mangrove plants have roots above the ground. These aerial prop roots have special air canals that allow the trees and shrubs to absorb oxygen, which is hard to get from the thick mud where the mangroves grow.

Rhizophora apiculata and *R. mucronata* are common mangrove trees in many tropical Pacific coastal areas; they can be identified quite easily by their hanging, aerial, prop roots.

Pacific Ocean Islands, Coastal Ecology, Fig. 4

Mangrove swamp bordered by fringing coral reef on the ocean side and coastal forest on the land side of Pohnpei Island. (Photograph by Mark Merlin)



Other relatively common mangroves, such as *Sonneratia alba* and *Lumnitzera littorea* can be recognized by the exposed, cone (or pencil) shaped roots that stick up out of the muddy soil from extensive, cable-like roots. These exposed cone shaped roots, known as pneumatophores, serve as breathing organs in the swampy areas. Still other mangrove trees, such as *Xylocarpus granatum* can be identified by their very distinctive buttressed (or arched) roots that spread out like cables (or ribbons). The extraordinary wavy roots stretch out, allowing this large mangrove tree to grow in the loose soil or thick mud. Another plant often associated with the mangrove forest is *Nypa fruticans* (Nipa Palm), which has many uses (Merlin and Juvik 1996).

The mangrove vegetation can be divided into zones from the edge of the land out into the lagoon, or from the shore out onto the muddy fringing reef flat. First, there is the intertidal, border zone where lowland forest and mangrove plants meet. It is a narrow zone along the high tide line with little or no vegetation. This zone usually has shallow mud deposits with patches of shrubs and trees. Some muddy areas in this zone emerge at low tide. This inner margin of the mangrove forest is usually quiet and protected (Teas 1983). The mangrove forest vegetation at the edge of the intertidal border often has more species of mangrove plants than other zones in the mangrove forest.

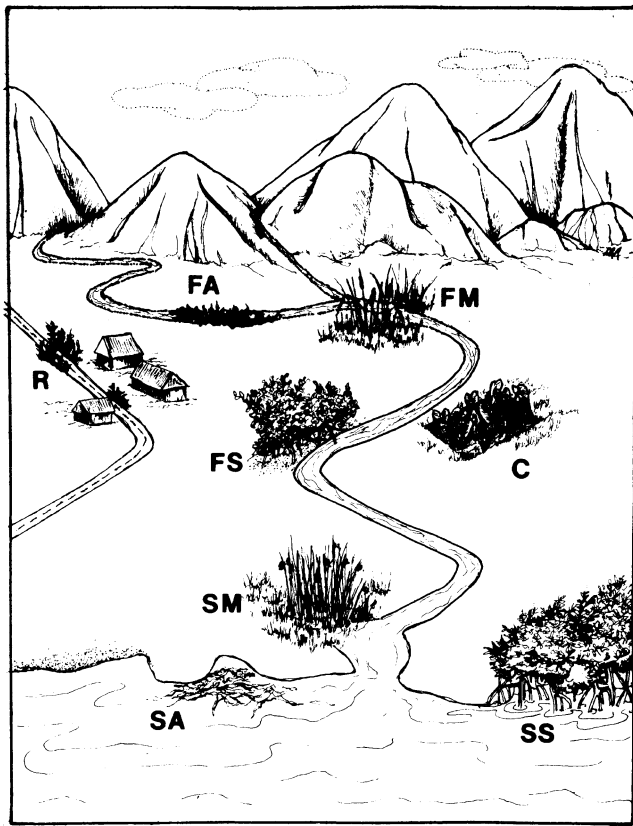
On the outer seaward fringe or channels in the mangrove forest where mangrove plants meet the sea, there is often an overhang of branches above the water. The overhanging branches have two basic forms, depending upon the dominant mangrove tree species fronting the reef flat or lagoon waters. The more common form of overhanging branches is dominated by the *Rhizophora* species. These trees (5–10 m,

15–30 ft), with a green bushy canopy and reddish brown, interlocking, aerial prop roots, often form a dense wall of vegetation. Less common is the form of overhanging branches dominated by groves of *S. alba*, a mangrove tree with the single thick trunk, open understory, and conical-shaped, breathing roots (pneumatophores) sticking out of the mud. Some species of clams and finfish are also taken from the mangrove forest areas.

Coastal Freshwater Wetlands, Swamps, and Marshes

Coastal freshwater wetlands, swamps, and marshes consist of those environments often flooded or saturated by ground or surface freshwater for periods of time long enough to support only plants that can grow in water-logged soils. Tropical freshwater wetland communities in Remote Oceania can be divided into five types: freshwater aquatic vegetation, freshwater marsh vegetation, freshwater swamp forest, cultivated wetland, and ruderal wetland (Stemmerman 1981; Merlin et al. 1992, 1993, see Fig. 5).

Freshwater aquatic vegetation develops in areas permanently flooded with freshwater. This vegetation type is found in slowmoving streams, reservoirs, irrigation ditches, natural and artificial ponds, and open bodies of water surrounded by freshwater marshes. The plants in this aquatic vegetation grow submerged, partially submerged, or floating in the freshwater. They lack well-developed structural support provided by the more or less rigid trunks, stems and branches of plants growing in soil on land. Most of the species in this type of vegetation are annuals. Some of the species found in



Pacific Ocean Islands, Coastal Ecology, Fig. 5 Schematic of wetland types: SA % seagrass beds, SM coastal saltwater marsh, SS mangrove swamp, FA freshwater aquatic, FM freshwater marsh, FS freshwater swamp, C cultivated wetlands, R ruderal wetlands (US Army Corps of Engineers)

the freshwater aquatic vegetation can become troublesome weeds that multiply rapidly and choke waterways. Two examples of aquatic plants that may produce this kind of problem in Remote Oceania are “water lilies” (*Nymphaea* spp.) and “water hyacinth” (*Eichhornia crassipes*), both alien species introduced into Remote Oceania during the twentieth century.

Freshwater marsh vegetation contains mostly nonwoody species, especially grasses and sedges that have roots in the water-logged soil, and are typically difficult to move through. Crops cannot be cultivated in the marshy areas without drainage. The vegetation of marshy areas is dominated by large sedges (Cyperaceae), grasses (Poaceae), and other herbs.

Freshwater swamp forests that have not been heavily exploited by people are mainly dominated by woody plants, such as *Terminalia* spp. and *Barringtonia racemosa*, which grow on soils that are often, if not always, water-logged. The swamp forests are found in depressions or other poorly drained places, usually behind the beach strand. In the past many freshwater swamp and marsh areas were cleared or severely altered in Remote Oceania for farming purposes. This often took place in areas with rich muddy soil where taro plants, such as *Colocasia esculenta* and *Cyrtosperma*

chamissonis could be cultivated. Many of these former freshwater swamps are presently covered with secondary vegetation containing plants that recolonize this kind of wetland after the original vegetation has been heavily disturbed. In those wetlands where considerable disturbance has taken place, and no root crops are being cultivated, successional plants such as *Hibiscus tiliaceus* and *Phragmites karka* are commonly found. The environmental protection of those areas with remaining stands of swamp forests, such as those with *Terminalia carolinensis* found on Pohnpei and Kosrae Islands in the Eastern Caroline Islands of Micronesia, should be seriously considered because of their unique and interesting vegetation, as well as other aspects of resource conservation.

Cultivated wetland exists in areas where traditional “root crops” are grown. In many wetland areas, cultivated plants have replaced natural freshwater marsh and swamp plants. As noted above, traditional, edible taro species are planted in this largely artificial or man-made vegetation (see Fig. 6).

Ruderal wetland vegetation is comprised mostly of weedy or successional species that are often aliens now common throughout the tropical Pacific. It occurs frequently in “waste” areas or terrain subject to flooding and interference by humans. This includes places such as rubbish dumps, roadside verges, and bombed sites.

Strand and Lowland Coastal Forest

The flora and fauna of the coastal land ecosystems of Remote Oceania represent, more or less, an attenuated Indo-Malayan biota, containing relatively few plant and animal species, and even fewer endemics. The size and composition of the biota in these ecosystems is largely a function of the distance from the Indo-Malayan region, size of the environment, hurricane (typhoon) frequency, droughts, salinity, native and introduced plants and animals, wars, and other human-induced disturbances (Manner 1987). Strand and lowland coastal forest includes ecosystems comprised of terrestrial organisms occurring near the coastline. The coastal or littoral vegetation in the tropical region of Remote Oceania is found in a narrow zone near the edge of the sea on open, sandy, and rocky shores. It also typically includes all of the native vegetation on islets comprising atolls.

Plants with different life forms such as small trees, shrubs, vines, and herbs, are found in the coastal or littoral vegetation. Although they can be very different, in some ways, these plants share several things in common. Many produce seeds that float and are able to survive for long periods of time immersed in saltwater. The seeds of these plants are often moved long distances by ocean currents. The plants of the coastal vegetation also have to be able to survive under very sunny, often windy and salty conditions, near the seashore.

At the edge of the sandy and rocky shores, the nonwoody plants commonly have fleshy leaves and sap that is salty. Most of these specially adapted coastal plants, as well as animals such as seabirds, are found along many, if not most, tropical Pacific shores (Merlin and Juvik 1996; Merlin et al. 1997).

The plants that are found in and make up the coastal vegetation can be divided into more or less four separate types of littoral plant communities, including the herbaceous strand, littoral shrub land, pandanus scrub, and littoral forest (Whistler 1994). Herbaceous strand or nonwoody vegetation is located above the high-tide mark of the ocean on both sandy

and rocky or coral rubble shores (see Fig. 7). Most strand species are perennials with deep taproots. Typical plants in the strand include creeping vines such as *Ipomoea pes-caprae* and *Vigna marina*. Littoral shrub land is typically located on windy coastal ridges and slopes, or on the seaward edges of coastal forests. Normally the dominant plant of this community is *Scaevola taccada*. Pandanus scrub is found usually on rocky, often exposed, windswept shores. As indicated in its name, this community is dominated by *Pandanus tectorius*, also known as “screwpine.” *Pemphis acidula* shrubs or small trees are common on rocky, limestone coasts. Littoral forest is

Pacific Ocean Islands, Coastal Ecology, Fig. 6 Cultivated wetland on Peleliu Island with *C. esculenta* taro. (Photograph by Mark Merlin)



Pacific Ocean Islands, Coastal Ecology, Fig. 7 Coastal forest on Islet of Majuro Atoll with coral reef at low tide in the foreground. (Photograph by Mark Merlin)



the most common kind of vegetation found along the tropical shores of Remote Oceania, typically just behind the strand vegetation. This forest is usually dense and is sometimes dominated by a single tree species. Common trees in this community include *Barringtonia asiatica*, *Hernandia sonora*, and *Callophyllum inophyllum*. The typical closed lowland wet forest is dominated by the widespread *Pisonia grandis*.

A variety of native animals reside in the coastal vegetation, especially where human disturbance has been low or non-existent and the lessremote islands. Among these animals are large numbers of breeding seabirds, some landbirds such as Pacific pigeons (*Ducula pacifica*) and fruit-doves (*Ptilinopus* spp.), some reptiles, and coconut crabs (*Birgus latro*).

Human Impact

Human encroachment on marine coastal environments and adjacent areas, has produced some dramatic effects. The discussion that follows describes some of the more important ecological impacts that people have had on coral reefs, seagrass beds, mangrove forests, coastal wetlands, and coastal land communities.

Natural disturbances are certainly major, if not the most important factors affecting coral reefs (Grigg and Dollar 1990). Nevertheless, the possible threats of human-induced global warming, including changes in sea level and increased frequency of intense low pressure system storms and El Niño Southern Oscillation events, are among many serious problems associated with human impact on coral reef ecosystems, including those in Remote Oceania. Among the more important problems are anchor damage, coral bleaching, coastal development, coral diseases, fish feeding, effects of fishing, tropical fish collecting, introduced species, marine tourism, sewage and eutrophication, effects of oil spills and heavy metals, effects of sedimentation, turtle tumors, effects of ultraviolet radiation on coral species, and effects of human divers (Richmond 1993; Gulko 1998).

The development and implementation of effective management plans to maintain the health and productivity of coral reefs requires a solid basis of scientific information and interpretation (Birkeland 1997). Unfortunately, much of this information is still inadequate. For example, in the Hawaiian Islands, invasive alien algae and excessive fishing for food and the aquarium trade are present and dangerous threats to the nearshore coral reefs. However, a basic ecological understanding of how these pestiferous algae have become successfully established is still undetermined. In this case, more data is needed to understand the inherent factors such as growth parameters and reproductive strategies, as well as the anthropogenic impact on factors such as nutrient regimes and

herbivore grazing pressure. These intrinsic and human-exacerbated factors interact ecologically to control the abundance of the alien algae (Hunter and Evans 1995).

As noted above, alien invasions are further complicated if additional human-induced stresses are imposed on ecosystems. Vigorous coral reef ecosystems are normally dominated by reef-building corals and coralline algae, with macroalgae and algal turfs typically limited to places on the reefs that are comparatively more difficult for herbivores to reach. In the case of Hawaii, algal growth (native and non-native species) has been favored in some areas because of excessive fishing and eutrophication, and this has resulted in severe and perhaps permanent changes in reef community structure (Borowitzka 1981; Steven and Larkum 1993; Maragos et al. 1996; McClanahan 1997).

Seagrass meadows form very productive ecosystems and provide nutrition and protection for many kinds of fish and invertebrates. They also help fasten the seafloor and maintain or improve water quality. Some natural disturbances have major effects on seagrasses. These include storms, floods, and droughts. For example, heavy rainfalls or extended droughts can severely affect the normal inflows of freshwater in the marine seagrass beds. Acute changes in salinity may result, and consequently have very adverse effects on the seagrasses. Intense storms may also redistribute coastal sediments through increased water current flows and larger influxes of fresh water. Increased turbidity may result, reducing the light available for seagrass photosynthesis. In addition, in extreme cases, huge quantities of sediment rearrangement can bury some seagrasses.

Anthropogenic disturbances on seagrasses can also be quite severe. Seagrass communities, especially those located in estuaries, are adapted to tolerate cyclic natural disturbances. Continuous heavy disturbances induced by human activities, however, may have severe impact on seagrasses. Water pollution, foreshore development, and recreational use of waterways are some of the major problems affecting the health of seagrass beds.

Intense terrestrial soil erosion can lead to excessive runoff that may limit the sunlight available to seagrasses. This tends to stunt their growth, or even kill the plants. Reductions in the area of seagrass growth can reduce number and extent of animals dependent upon these plant species. Nutrient inputs from agricultural runoff and urban areas can increase to the point that abnormal algal blooms develop. Such blooms decrease light penetration and limit the range and productivity of seagrasses. Reductions in available light may also result from dredging and fish trawling activities that increase turbidity through sediment resuspension. In several areas, inappropriate ship movement in shallow waters where seagrasses are found is another critical problem.

Extensive damage to marine and estuarine habitats due to the unintentional destruction of the seagrasses has been

Pacific Ocean Islands, Coastal Ecology, Fig. 8 Alien mangrove, *R. mangle* in coastal habitat on O'ahu, Hawai'i. (Photograph by Mark Merlin)



documented in several areas. Among the few animals that feed, more or less, exclusively on seagrasses, are green turtles and dugongs, and since these species are listed as threatened animals, their continued existence is dependent upon preservation of seagrass beds.

Mangroves, the “rainforests of the sea,” dominate a large portion of the world’s subtropical and tropical coastlines, covering an estimated area of 22 million hectares. People in tropical areas throughout the world, including Remote Oceania, utilize the raw materials of trees of the mangrove forests FOR numerous purposes (Baines 1981). Unfortunately, during last several decades, the global area in mangroves has diminished at an increasing rate due to a variety of human activities. These disturbances include harvesting for firewood, charcoal, resins, and building materials, as well as freshwater diversion, dredging for land reclamation projects, and drainage to facilitate road construction. To some extent these impacts are also occurring in Remote Oceania (Hamilton and Snedaker 1991).

Mangrove habitat has a special role in maintaining the health of the marine environment and human industries that depend on its abundant life. The increasing removal of mangrove forests for a variety of reasons is having significant impact on fishing industries and tourism. The natural hatchery function of mangrove forest helps restock many marine populations, and mangrove forests have become popular attractions for ecotourism. Like other plants, mangrove trees and shrubs require oxygen. However, a coat of oil, even a thin one, can block the breathing pores on the aerial roots of mangroves and may kill them. Therefore, oil spills can be very dangerous for mangroves.

The introduction of alien species in some cases is also causing profound changes in the ecology of coastal environments in Remote Oceania. For example, as noted above, in the Hawaiian Islands, the introduction of invasive algal species (e.g., *Gracilaria salicornia*) has had significant impact in some coral reef regions. In these same islands, the introduction of mangrove trees (particularly *Rhizophora mangle*, see Fig. 8) to the coastal zone of Hawaii, which previously had no native mangrove species, has created a series of alien mangrove forests, especially on O’ahu Island (Juvik and Juvik 1998). Today, these mangroves intrude upon historical sites and critical habitat for a number of endangered shorebirds.

Low lying, freshwater swamp forests, just inland of mangroves and above tidal influence, are greatly disturbed in many areas of Remote Oceania. Most native vegetation of these freshwater swamp forests has been cleared or modified by human activity. Traditional and modern expansion of taro cultivation, as well as the commercial plantation development and the negative impacts of road development, have greatly affected the freshwater swamp forest habitat in many tropical Pacific Islands. There is an urgent need to protect the widely threatened, native freshwater swamp ecosystems in many areas of the tropical world, including those in Remote Oceania.

The majority of people in Remote Oceania have traditionally lived relatively close to the coastlines of the isolated islands of this region, and this settlement plan has continued into modern times. People have thus utilized the coastal zone for many purposes including the construction of housing. Therefore, the coastal flora and fauna in many areas has been modified significantly by human activity. Over the years, a great many native trees, including the often dominant

trees of *P. grandis* and others, have been replaced with useful introduced, subsistence-oriented food species such as trees of coconut (*Cocos nucifera*), breadfruit (*Artocarpus altilis*), and Tahitian chestnut (*Inocarpus fagifer*).

During the last two centuries or so, much of the area of the traditional cultivated coastal forests of Remote Oceania has been converted to, more or less, single stands of commercial coconut and other plantations. Traditionally, native peoples have left much of the native strand vegetation and forest immediately adjacent to the coastline for protection from strong winds and storm surge. However, modern urban, industrial, military, and tourist development has resulted in the removal of all or most of the natural components of the coastal land ecosystems in many areas of the islands of Remote Oceania. Even on remote atolls, because of traditional farming practices, little remains of the native forests, except on those that remain uninhabited.

Cross-References

- Atolls
- Changing Sea Levels
- Climate Patterns in the Coastal Zone
- Coastal Subsidence
- Coral Reef Islands
- Coral Reefs
- Coral Reefs, Emerged
- Estuaries
- Holocene Coastal Geomorphology
- Human Impact on Coasts
- Mangroves, Ecology
- Natural Hazards
- Pacific Ocean Islands, Coastal Geomorphology
- Salt Marsh
- Sea-Level and Climate Change
- Small Islands
- Storm Surge
- Uplift Coasts
- Volcanic Coasts
- Wetlands

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Pacific Ocean Islands, Coastal Geomorphology

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Scattered across one-third of the earth's surface, Pacific Islands have some of the world's most fascinating landscapes (Menard 1986; Nunn 1994, 1998a). Yet notwithstanding their common attributes (comparatively small size, remoteness, young geology, topographic and structural simplicity, simple climate, and soil patterns) it would be wrong to suppose that Pacific Island coastal landforms are uniform within this vast area.

Much variation stems from latitude. Tropical Pacific Islands within the belt of tropical cyclones (hurricanes, typhoons) generally have coastal landforms which manifest the occasional impact of high-energy waves and winds while those outside this belt may not.

The primary cause of coastal variations in geomorphology lies with island type, of which three major kinds can be recognized in the Pacific;

1. volcanic islands, often slowly subsiding, generally high, well vegetated with broad coastal plains around river mouths;
2. high limestone islands, often rising, commonly cliffed, dense vegetation often only locally, with little lowland close to the shore; and
3. atoll islands, usually rising no more than 3 m above mean sea level, made largely from unconsolidated materials accumulated on reef flats.

Most Pacific Islands are concentrated in the ocean's south-west quadrant because this is where the processes largely responsible for island formation dominate. Most Pacific Islands also occur within the tropics where coral reefs grow.

Within this broad picture, volcanic islands generally occur either in arcs parallel to convergent plate boundaries or in chains in intraplate (midplate) locations. Most high limestone islands are either associated with (former) lines of plate convergence or places where flexure of the intraplate lithosphere has taken place. Most atolls occur in tropical intraplate locations marking the places where volcanic islands have sunk.

Coastal Geomorphology

Owing to the approximately 120-m (postglacial) rise in sea level between the Last Glacial Maximum some 17,000–21,000 years ago and 5,000–4,000 years ago, all Pacific Island coasts exhibit signs of drowning. For many years, it was considered by some that the “Micronesia Curve” of sea-level rise, which supposed sea level to have been rising continuously since the Last Glacial Maximum (Bloom 1970), was applicable to the Pacific Islands. More recent work has shown that instead, throughout this region, sea level reached a maximum around 5,000–4,000 years ago and has since fallen 1–2 m (Fig. 1; Nunn 1995). Thus, superimposed on the drowned nature of many coasts are the effects of a slight recent emergence. Superimposed on this in turn are the effects of tectonics (land-level changes) which affect many islands, particularly those along convergent plate boundaries.

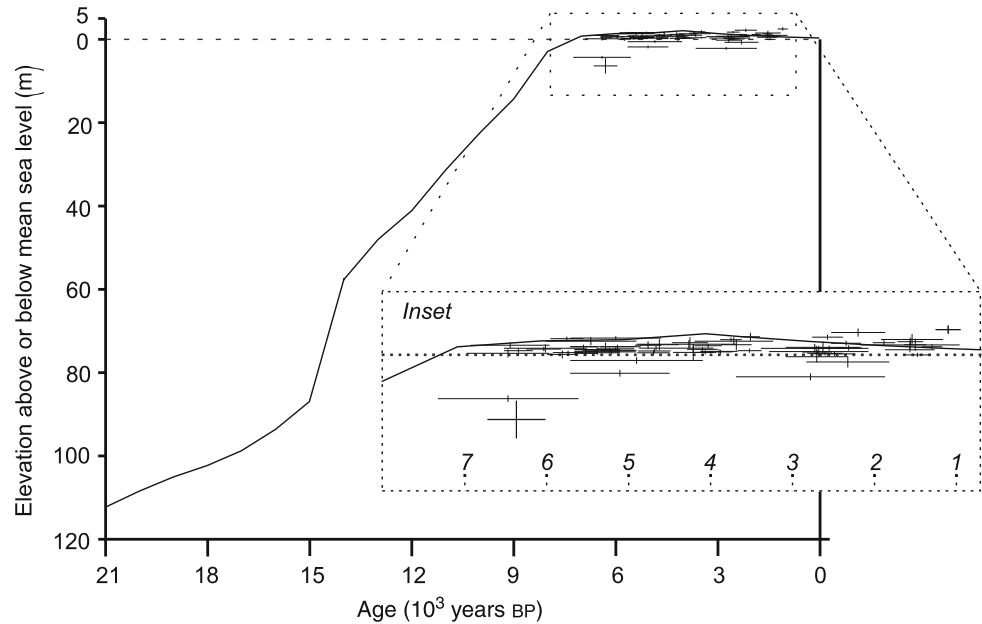
Postglacial warming of ocean surface waters led to the recolonization of many nearshore and shallow areas by corals and associated reef organisms. As sea level continued to rise, corals grew upwards so that the reef surface remained within the photic zone. In some places, such as Tarawa in Kiribati and parts of French Polynesia, coral reefs apparently managed to “keep up” with rising sea level (Nunn 1994). Elsewhere, as in most parts of the tropical southwest Pacific, for example, such reefs grew upwards at slower rates yet managed to “catch up” to sea level when its postglacial rise slowed around 6,000–5,000 years ago. In a few places, notably to the west of the islands of Samoa, reefs believed to have been established at sea level during the early postglacial were unable to grow upwards fast enough and are not visible at the surface today. These are characterized as “give up” reefs in the nomenclature of Neumann and McIntyre (1985).

The importance of coral reefs in the dynamics of modern coastlines in the tropical Pacific Islands cannot be overstated. Many beaches and sand islands are made solely from sediment created on reefs and, where those reefs become degraded and cease to be productive, beaches and sand islands can become severely eroded.

There has been a regrettable tendency to explain all environmental changes within the postsettlement history of Pacific Island coasts by human actions. While this may be true in large part of the last 100 years or so, there is clear evidence that at earlier times natural changes overwhelmed

Pacific Ocean Islands, Coastal Geomorphology, Fig. 1

The course of postglacial sea-level changes in the Fiji Islands (after Nunn and Peltier 2001). The solid line is the predicted sea-level change from the ICE-4G model. The data points are paleosea-level indicators from Fiji, enlarged in the inset to show the general concordance between empirical and predicted data



human endeavors. One of the most significant such occasions was the “AD 1300 event” in which a slight cooling, sea-level fall and increased El Niño frequency, caused widespread reef-surface death and infilling of coastal embayments, impeded lagoon-water circulation, and invoked major societal responses in Pacific Island societies (Nunn 2000a; Nunn and Britton 2001).

The following sections look at Pacific Island coastal geomorphology for each of the three major island types identified above.

Coastal Geomorphology of Volcanic Islands

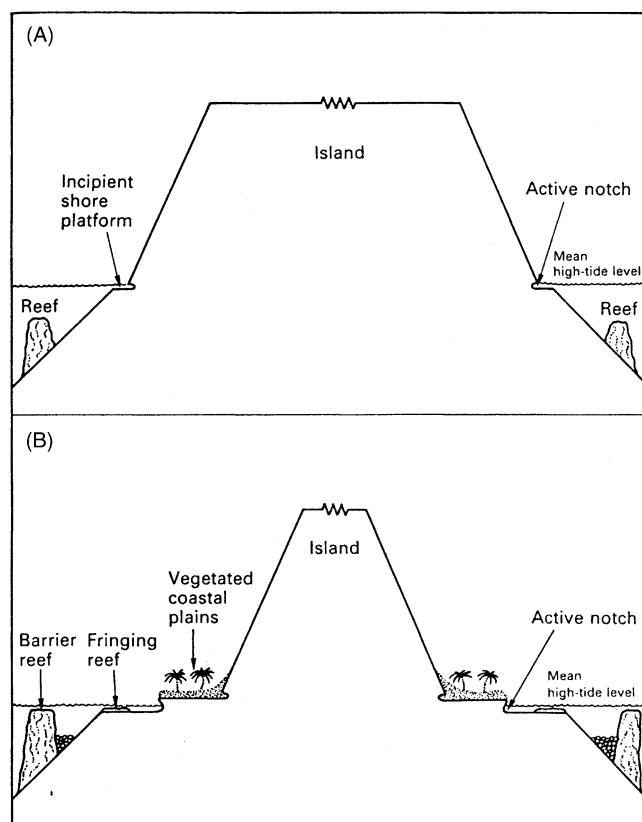
All oceanic islands in the Pacific began life as ocean-floor volcanoes but not all retain a recognizable volcanic form. Those which do are commonly young (<5 million years old) and/or in intraplate locations. Except around the mouths of large rivers, such islands typically have narrow coastal plains and, beneath their fringing reef platforms (where these exist), plunge steeply offshore. Most such coastal plains were formed only after the sea reached its present level 5,000–4,000 years ago and stabilized long enough (after its long postglacial rise) to erode platforms along island coasts. When sea level fell subsequently, these platforms emerged and became draped with sediment to form the coastal plains which have been the favored sites for human habitation in the Pacific Islands throughout the late Holocene (Fig. 2).

Late Holocene sea-level fall also led to the outgrowth of both coral reefs and those coastlines composed of unconsolidated sediments, the infilling of coastal lagoons with sediment (both terrigenous and marine), and the subsequent establishment of characteristic ecosystems such as mangrove swamps (see below).

There is considerable controversy about the precise impact of early humans on Pacific Island landscapes, some regarding it as considerable (Kirch and Hunt 1997), others as generally slight (Nunn 1999a, 2001). Typical of the former is the scenario constructed for Aneityum Island in Vanuatu, where the earliest settlers are believed to have settled in upland areas (the valley floors then being too swampy), burned the forest which existed there, thus releasing large quantities of soil and sediment into the valleys which, after about 1,000 years of occupation, filled them up sufficiently for them to be settled by the people. The latter explanation sees humans as the passive agents in a landscape which was being changed solely by emergence of the land, in this case as a result of sea-level fall and tectonic uplift.

Coastal Geomorphology of High Limestone Islands

High limestone islands are commonest in the Pacific along convergent plate boundaries where tectonic processes are generally most active and most variable in time and space. Typically such islands were formerly subsiding volcanic islands which developed a reefal capping before being uplifted. A good example are the Lau Islands of eastern Fiji which rise from a volcanic arc which became inactive and began subsiding some 5 million years ago. Subsequent uplift in the Plio–Pleistocene, perhaps due to the heating-up of a detached, partly subducted slab of lithosphere, elevated these islands as much as 315 m above sea level (Nunn 1998b). Another example is the island Niue, which has been gradually rising up a flexure in the lithosphere east of the Tonga Trench for around 500,000 years, and is a fine example of an emerged atoll with the former ring reef and atoll lagoon well preserved 70 m above sea level (Hill 1996).



Pacific Ocean Islands, Coastal Geomorphology, Fig. 2 Formation of coastal plains around volcanic islands during the Holocene (from Nunn 1994). (a) The time of the Holocene sea-level maximum. Sea level had been rising for the previous 11,000–13,000 years so little lateral erosion of the coastline had been accomplished. Reefs struggled to catch up with rising sea level; until they did so, a high-energy window was open. Potential settlers would have found this island type unattractive. (b) Following a fall of sea level in the late Holocene, the shore platforms cut at the higher sea level would have emerged and become covered with terrigenous and marine detritus. Barrier reefs would have caught up with sea level, the high-energy window would have been closed so that coastline erosion would have become relatively subdued. Fringing reefs would have developed and the lagoon become shallower. Potential settlers would have found this island type comparatively attractive

Owing to a lack of running water and thus only comparatively little subaerial fluvial development, most high limestone island coasts are cliffed, often riddled with long sub-horizontal epiphreatic caves marking former stands of the water table (controlled by sea level). Coastal embayments on such islands tend to be few and formed from the drowning of karstic hollows or collapsed caves.

A variant on high islands made solely from limestone is the *makatea* islands in which a volcanic island is fringed by high limestone, the products of fringing-reef uplift (Nunn 1994). Examples abound in the southern Cook Islands. Their coasts are similar to those of other high islands, one difference being that because of the fertility of their inland areas, they tend to have been comparatively densely populated so

that human modifications to their coasts are sometimes considerable.

Coastal Geomorphology of Atoll Islands

Atoll (and barrier-reef) islands—narrow strips of land formed on broad atoll (or barrier) reef flats from the accumulation of associated sediments—are entirely coastal. Rarely rising more than 3 m above sea level, atoll islands generally develop on the windward sides of atoll reefs. In the tropical Pacific, most atolls lie in the belt of easterlies and thus have islands exclusively along their eastern sides; good examples are Kapingamarangi Atoll in Micronesia and Tarawa Atoll in Kiribati. Some atolls are occasionally affected by storm waves coming from the opposite direction to windward, and thus display islands of coarser sediment along their leeward sides than those to windward; examples include Nanumaga Atoll in Tuvalu (McLean and Hosking 1991). Most atoll reefs enclose a shallow subcircular lagoon and in those cases where the lagoon is small relative to sediment supply, it may become infilled; such is the case with Nui and Vaitupu Atolls in Tuvalu (McLean and Hosking 1991).

Atoll coasts are among the most vulnerable in the Pacific because of their often/largely unconsolidated character. Ephemeral atoll islands named cays can be distinguished from those which are more permanent named *motu* (Nunn 1994). *Motu* are atoll islands which owe their durability to the formation of various rock armors such as beachrock, conglomerate platforms (*pakakota*), or phosphate rock (Fig. 3). *Pakakota* occur widely in the Tuamotus. Even more enduring among such islands are those which have experienced a degree of emergence (uplift and/or sea-level fall) and thus have a core of fossil reef within the sediment pile. Many of the larger atoll islands, such as several in Kiribati and the Marshall Islands, may have such emerged-reef cores.

Tectonic Controls on Island Coastal Geomorphology

Islands which are sinking or rising comparatively rapidly exhibit distinctive coastal landforms. Subsiding islands generally show an unusual degree of coastal embayment, as recognized first by Dana (1872). Where such islands are composed of a single volcano, drained radially, a characteristic stellate island may form; the island Ono in the Kadavu group of southern Fiji is a classic example. Faulted coasts which are subsided may well be quite straight; they are particularly common along the sides of young high subaerial volcanic islands, as in parts of Samoa. Although large deltas are understandably rare on Pacific Islands, where they occur, they are usually associated with local subsidence. In such instances, sediment supply to the delta front usually exceeds the effects of subsidence, and characteristic delta forms result. Examples include the Rewa Delta on Vitilevu Island in Fiji.

Pacific Ocean Islands, Coastal Geomorphology, Fig. 3

Beachrock exposed along the northern (leeward) coast of Eluvuka (Treasure) Island in the Mamanuca Group of western Fiji. Exposure of this beachrock is thought to be due to erosion of the beach in successive tropical cyclones during the 1990s followed by persistent northwesterly winds which have prevented the return of sand cover. [Digital photo 26—P. Nunn]



Uplifting islands in the tropical Pacific are commonly marked by staircases of emerged coral reefs, comprising fossil reef surfaces separated by terrace risers. Dating of reef staircases on the Huon Peninsula in Papua New Guinea (Chappell 1974) and elsewhere in the southwest Pacific have given rise to precise chronologies of sea-level and tectonic change during the Quaternary. Most coasts of uplifting islands are cliffed or have only very slightly developed coastal plains. Many of these appear to have been formed only during the period of comparative (land–sea) stability between bursts of uplift, commonly associated with large earthquakes and are thus termed coseismic. Coseismic uplift is known to affect many islands along convergent plate boundaries in the southwest and western Pacific. Coseismic uplifts with average magnitudes of 0.74 m affect Tongatapu Island in Tonga approximately every 870 years (Nunn and Finau 1995). The Fiji island Vatulele is thought to experience coseismic uplift of some 1.8 m approximately every 1,400 years (Nunn 1998a).

Challenges for the Twenty–First Century

Pacific Islands and the submarine ridges from which many rise are among the steepest structures on earth and, while the ocean contributes to their stability, it cannot prevent occasional structural failures. Owing to their infrequency, the nature and (potential) effects of such failures have become

clear only recently. A groundbreaking study of the structural failure of Johnston Atoll (Keating 1987) was followed by others, summarized by Keating and McGuire (2000). It has been estimated that a major oceanic island flank failure occurs once every 25 years (Siebert 1992). Flank collapses along the Hawaiian Ridge have been studied and dated using deeply submerged fossil reefs (Moore and Moore 1984; Moore et al. 1994). Such failures/collapses are clearly capable of causing the disappearance of parts of islands, perhaps even whole islands, and producing giant waves which could wrought major changes to Pacific coasts. Their study is of considerable applied interest.

Pacific Island coasts composed of unconsolidated sediments are particularly vulnerable to change as sea level changes. While short-term changes, particularly during El Niño events, can have severe consequences for reef health, it is longer-term sea-level rise, particularly within the past 200 years or so, that has brought about significant changes to many Pacific Island coasts. Based on largely anecdotal evidence (more formal sources generally being absent from this region), it is clear that some coasts have been inundated and/or eroded laterally by as much as 200 m within the last 100 years. Recent warnings have been issued about the disappearing beaches of the Pacific Islands (Coyne et al. 1996; Nunn 1999b).

Such problems have been compounded by the human inhabitants of Pacific Island coasts. Since humans came to occupy most Pacific Islands in larger numbers and to make

more demands on their coasts, so their landscapes have been modified in consequence (Nunn et al. 1999). Much coastal vegetation has been removed and/or replaced by less suitable species. Mangroves in particular have been a major casualty of twentieth century “development” in the Pacific Islands; the case studies of Ovalau and Moturiki Islands in Fiji show how those settlements which have removed their fringing mangrove forests now have severe problems of coastal erosion and inundation compared to those which preserved them (Nunn 2000b).

Cross-References

- Atolls
- Beachrock
- Cays
- Changing Sea Levels
- Clified Coasts
- Cliffs, Erosion Rates
- Coral Reef Coasts
- Coral Reef Islands
- Coral Reefs
- Coral Reefs, Emerged
- Faulted Coasts
- Mangroves, Ecology
- Mangroves, Geomorphology
- Pacific Ocean Islands, Coastal Ecology
- Submerged Coasts
- Submerging Coasts
- Uplift Coasts
- Volcanic Coasts

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Paired Baymouth Spits

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Synonyms

Baymouth barriers; Double spits; Paired spits

Definition

A system of two convergent spits that partially or almost completely encloses a bay or river mouth.

Keys for Identification

Paired baymouth spits are a particular case of spits, which are one of the three process-oriented functional categories of coastal barriers (Otvos 2012). Then, similarly to other spits, each one of the two spits forming a paired baymouth spits system is an emerging, elongated, narrow, sand or gravel body. Both spits are attached to and prograded from each margin of the bay, river mouth, or tidal inlet, and very important, they are converging to the other one since their formation and during their geomorphological evolution, despite in some periods one of the spits can be partially or almost completely eroded.

Bays and river mouths can be partially or near-completely closed by another kind of coastal barriers. In fact, it is relatively usual to observe a coastal barrier with an inlet that encloses a bay, forming a coastal lagoon landward. Therefore, the morphological configuration of paired baymouth spits is similar to some other coastal barriers closing a bay. So, only when the geomorphological evolution of both sedimentary bodies closing the bay or river mouth show a convergent trend, they can be defined as paired baymouth spits.

The term baymouth barriers (Otvos 2012) can be used for several single barriers closing different bays; the terms double spits and paired spits are also used for two convergent spits formed during the closure process of coastal lagoons (Zenkovich 1967). So, the term paired baymouth spits should be preferred to refer to two convergent spits in the mouth of a bay or river. On the other hand, coastal barriers, either attached to the mainland or forming barrier islands, usually present tidal inlets. So, these inlets are limited by two emerged, elongated, narrow sand or gravel bodies. The sedimentary longshore drift determines that the down-drift body shows an alongshore progradation while the counter-drift body shows an alongshore erosion. Therefore, the inlet migrates according to the dominant direction of the longshore drift, and both sand or gravel bodies do not demonstrate a convergent behavior. Exceptionally, they have been identifying some tidal inlets limited by two convergent spits, and consequently, they can be considered paired baymouth spits (Lynch-Blosse and Kumar 1976).

Formation and Degradation of Paired Baymouth Spits

General Conditions

An efficient longshore sediment transport is necessary for the formation of spits. Single baymouth spits close bays forming

sand barrier – lagoon systems, with inlets maintained by tidal currents or river flows. Similarly, longshore drift also forms detached barrier spits in some estuarine river mouths; these baymouth spits can be broken during events of high fluvial discharge (Nienhuis et al. 2016).

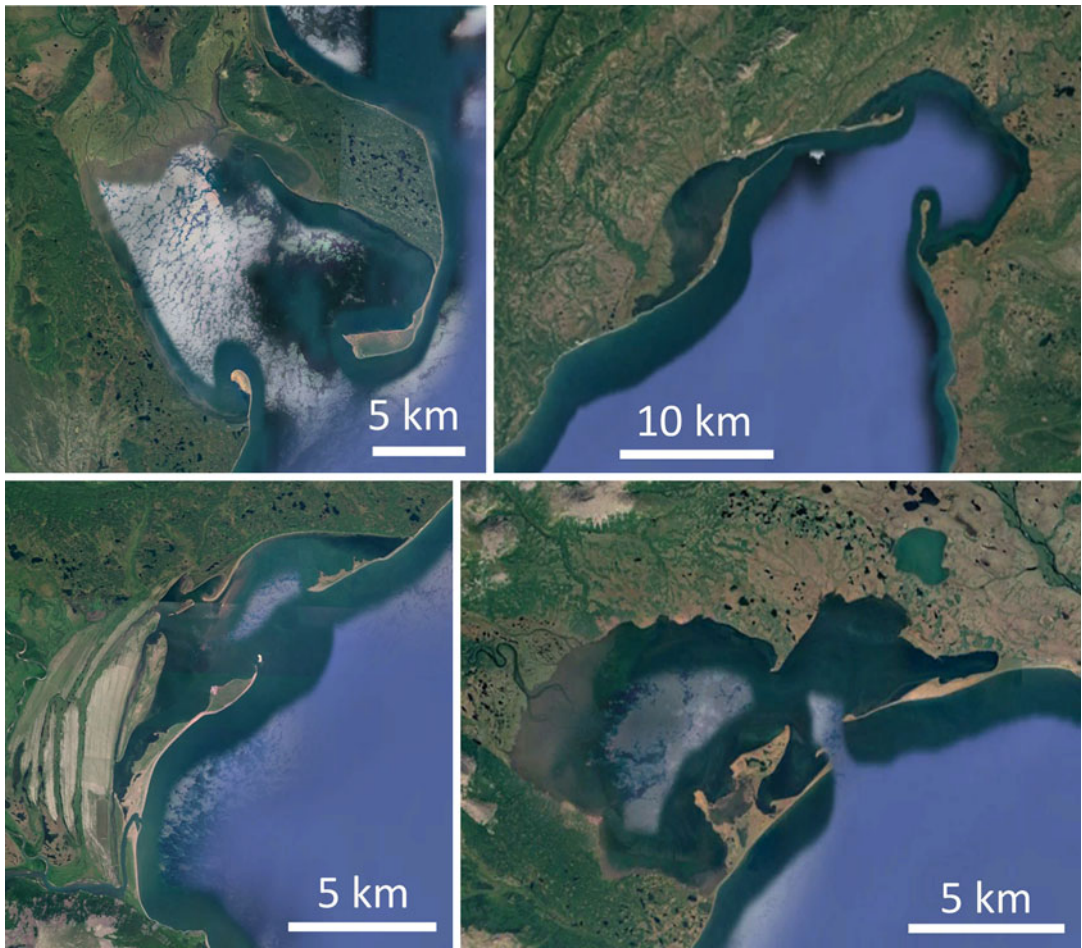
Paired baymouth spits are a particular case of spits that only take place under more special conditions. In narrow bays, the convergent progradation of both spits can be due to frontal waves, generating convergent longshore drifts from the headlands on both sides of the bay (Zenkovich 1967). On the other hand, in open coasts, the convergent longshore drift is related to a seasonal alternation of two wave approach directions. The dominant waves, with a most efficient longshore drift, favor the progradation of the down-drift spit, while the waves of the second component favor the progradation of the counter-drift. Therefore, each spit is typically prograding in a different period of the year, resulting in a convergent evolution. However, the convergence of the paired baymouth spits cannot be only explained by a bidirectional wave pattern, because the net sediment budget for the counter-drift spit would result in an erosional trend. Therefore, it is necessary a hydraulic blockage, which can be due to tidal currents (Lynch-Blosse and Kumar 1976; Aubrey and Gaines 1982) or a fluvial discharge with high competence (Kunte and Wagle 1991; Avinash et al. 2013; Alcántara-Carrió et al. 2018). This cross-shore water current favors the net progradation of the counter-drift spit because it avoids or at least hampers its erosion by the dominant waves.

A significant concentration of paired baymouth spits is located on the eastern flank of the Kamchatka Peninsula (Fig. 1), as well as on the microtidal tropical regions of Western India and Southeastern Brazil, where high fluvial discharge favors the hydraulic blockage of the bidirectional longshore drift. Most of them correspond to microtidal coasts. It is likely due to coastal barriers are better developed in mixture conditions between wave-dominated and river-dominated coasts than in tide-dominated coasts. Nevertheless, some paired baymouth spits have also been identified in mesotidal and macrotidal coasts, while any study case has been reported for megatidal coasts (Table 1).

They have been described in the literature five models for the formation of paired baymouth spits. All of them are based on a bidirectional longshore drift by waves, and most of them include a hydraulic blockage.

Model 1: Break of a Coastal Barrier

A coastal barrier, not necessarily a spit, can be broken for the development of a new fluvial outlet (Robinson 1955; Carr 1986), a new tidal inlet, or extreme events as hurricanes (Aubrey and Gaines 1982) and tsunamis (Liu 2013). The resulting configuration, where the outlet remained relatively stable or migrated due to the progradation of the down-drift spit, was defined as double spits for several study cases in the



Paired Baymouth Spits, Fig. 1 Examples of paired baymouth spits on the eastern flank of Kamchatka Peninsula (E Russia) (Source of LANDSAT images: Gougale Earth Pro)

Paired Baymouth Spits, Table 1 Distribution of paired baymouth spits described in the literature among microtidal, mesotidal and macrotidal coasts

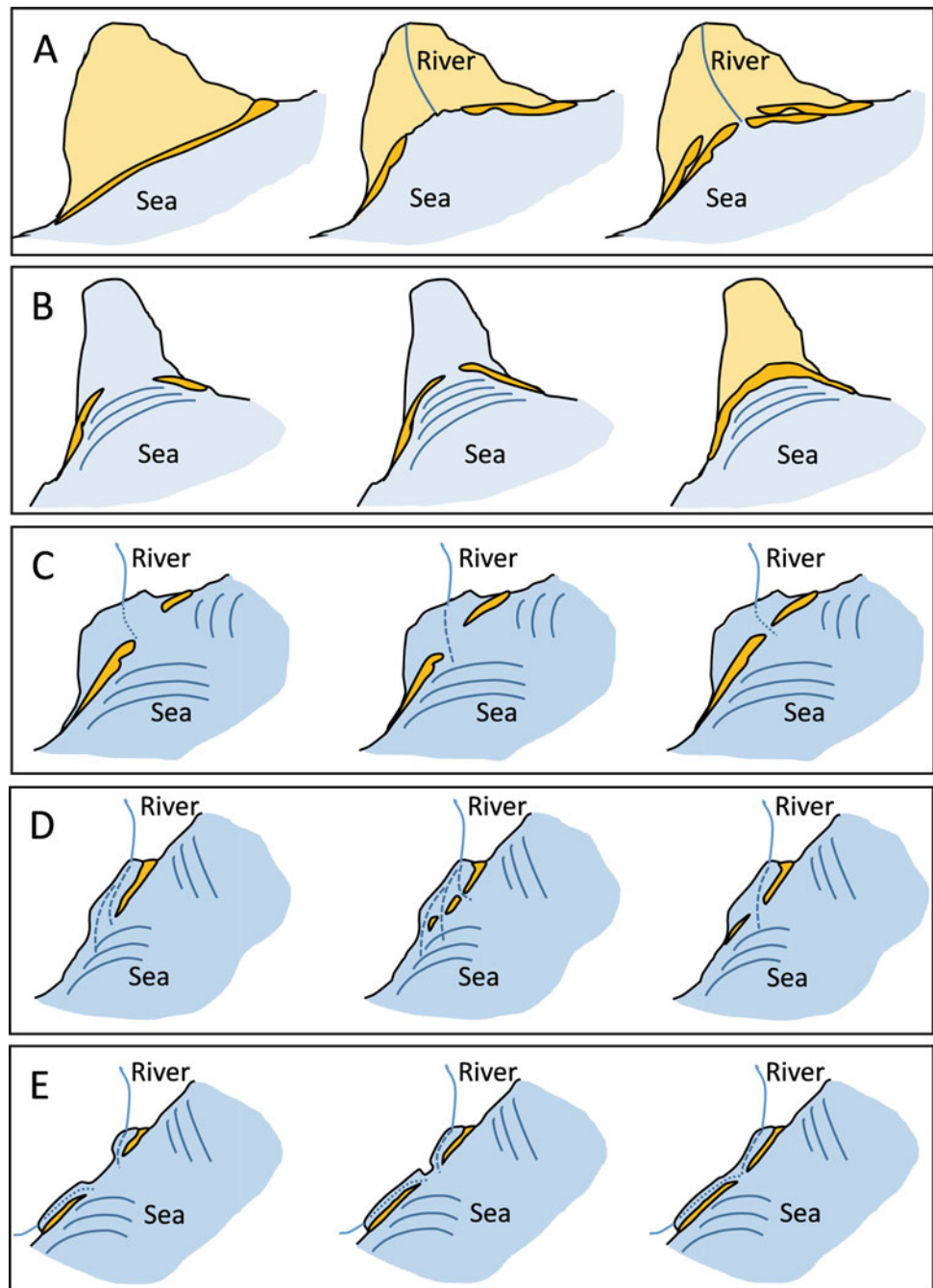
	Microtidal	Mesotidal	Macrotidal
Ward (1922), Lovegrove (1953)			Rye Bay (England)
Robinson (1955)	Poole and Christchurch (England)		Pagham (England)
Zenkovich (1967)		Kamchatka (Russia)	
Lynch-Blosse and Kumar (1976)	Brigantine (USA)		
Aubrey and Gaines (1982)	Popponeset (USA)		
Hume and Herdendorf (1988)	Ohiwa (New Zealand)		
Kunte and Wagle (1991), Avinash et al. (2013)	Karnataka (India)		
Alcántara-Carrió et al. (2018)	Iguape (Brazil)		

South of England (Robinson 1955; Carr 1986). Similarly, after the break of the Popponeset Spit, a double spit configuration was formed, but the counter-drift was progressively eroded until it was completely degraded in few decades (Aubrey and Gaines 1982). Therefore, the break of a coastal barrier or spit can generate a double spit morphology, which

remains or is ephemeral, but actually, they are not paired baymouth spits because both spits did not show a convergent evolution. On the other hand, two recurved spits were formed after the break of a barrier beach at Rye Bay. During four centuries, both spits showed a convergent progradation, until they

Paired Baymouth Spits,

Fig. 2 Conceptual models for the formation of paired baymouth spits: (a) break of a coastal barrier (modified from Lovegrove 1953), (b) convergent longshore drift in a narrow bay (modified from Zenkovich 1967), (c) bidirectional longshore drift with hydraulic blockage, (d) high hydraulic blockage cutting a detached spit, and (e) convergence of two outlets and the associated spits



practically closed the bay, which currently is maintained open by the estuarine flow of Rover River (Ward 1922; Lovegrove 1953). Therefore, this is the best and perhaps the only study case of baymouth spits formed after the break of a previous coastal barrier (Fig. 2a).

Model 2: Convergent Longshore Drift in a Narrow Bay

In narrow bays, fjords, and rías, wave incidence can generate convergent longshore sediment transport from both sides, and therefore, it forms two convergent spits

(Fig. 2b). In this model, a hydraulic blockage is not necessary. In fact, sometimes the baymouth spits merge forming a baymouth barrier closing the bay, with the wider section in the center of the barrier (Zenkovich 1967).

Model 3: One Outlet with High Longshore Drift and Hydraulic Blockage

The longshore drift due to the dominant wave component determines the formation and development of the down-drift



Paired Baymouth Spits, Fig. 3 Rapid formation and degradation (2002–2016) of baymouth spits at the microtidal Itabapoana River mouth (SE Brazil) (Source of LANDSAT images: Google Earth Pro)

spit, which progressively is closing the bay or river mouth during decades, centuries, or longer periods. In addition, the hydraulic blockage of the dominant longshore drift by tidal currents or river flow can favor the formation and progradation of a counter-drift spit from the opposite margin of the tidal inlet (Lynch-Blosse and Kumar 1976) or the opposite headland of the bay (Kunte and Wagle 1991) (Fig. 2c).

Model 4: One Outlet with High Hydraulic Blockage of Longshore Drift

In a river mouth with high fluvial competence, in contrast with a lower longshore drift, the counter-drift spit can be developed during several decades, while the down-drift spit in the opposite margin is not formed. So, extreme rainfall and fluvial discharge, which can be associated with climatic oscillations like El Niño events, can erode the fronting counter-drift spit, forming sandbanks that later emerge attached to the other margin of the mouth, forming the down-drift spit (Figs. 2d, 3, and 4b to d). In this model, the down-drift spit shows an ephemeral duration, being again degraded in the scale of a decade (Alcantara-Carrió et al. 2018).

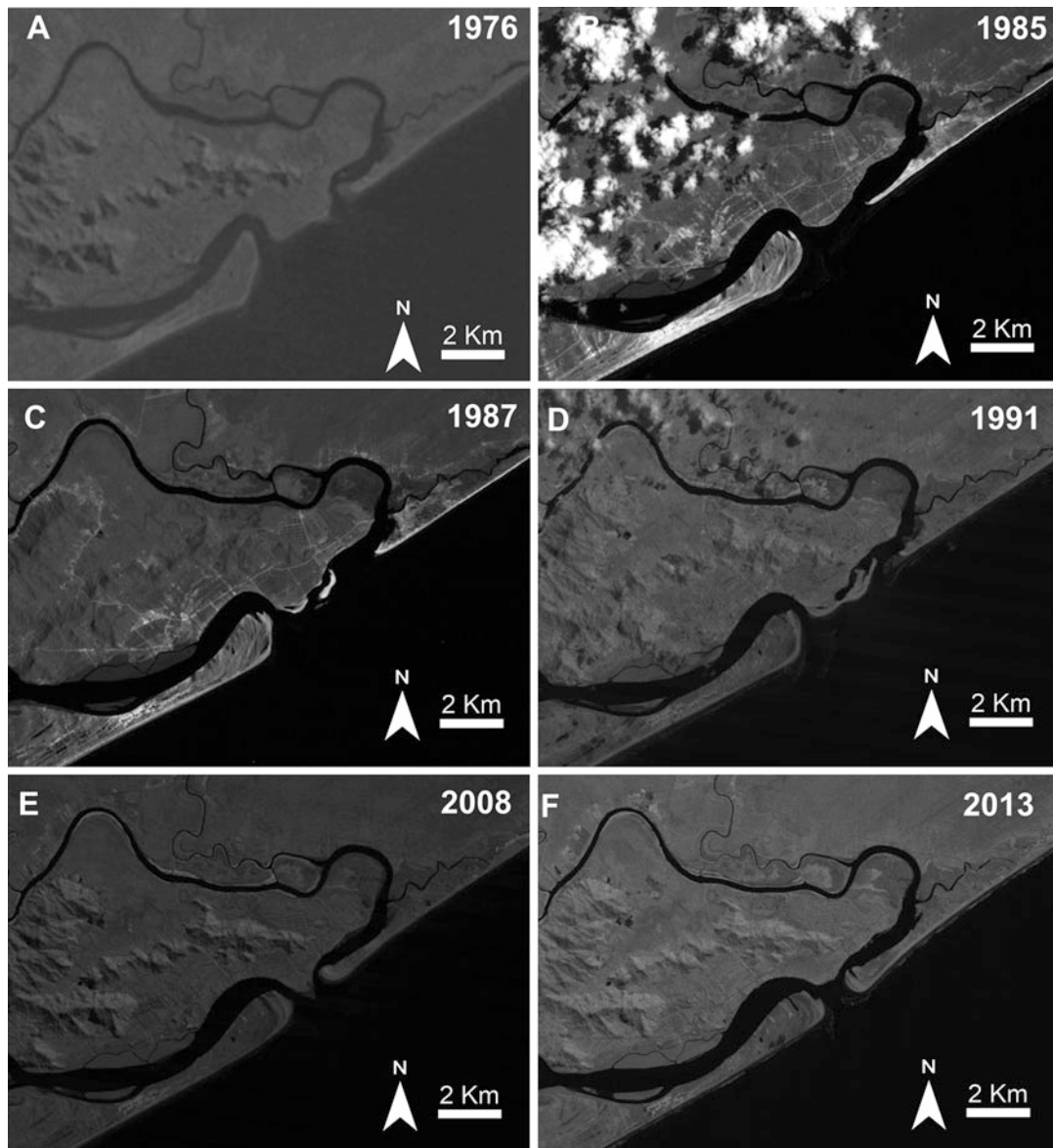
Model 5: Convergence of Two Outlets and their Associated Spits

Previous models for the formation of baymouth spits have been related to the presence of a single outlet, which can correspond to a tidal inlet or a river mouth. However, two independent spits, associated with different bays or river

mouths that initially were located far away, can also become baymouth spits. In this model, one of the spits is formed by the dominant longshore drift, i.e., it is a down-drift spit associated to a bay with either or tidal currents or river flow with low competence. On the contrary, the second one is a counter-drift spit detached fronting a high-competence river flow. One or both outlets, during decades or centuries, progressively erode the central headland that isolated them, until both outlets became merged, with the associated down-drift and counter-drift spits outside (Figs. 2e and 4e, f). In this model, both spits have been developed during decades, centuries, or even millennia, and this configuration can remain for a long period too, although the location of the combined outlet between both spits can migrate (Alcántara-Carrió et al. 2018).

Numerical Modeling of Paired Baymouth Spits

Formation of paired baymouth spits has been obtained by numerical simulations, firstly for symmetrical spits and later, using different seaward arrangements of the sandy headlands, for asymmetrical spits (Uda and Serizawa 2011; Watanabe et al. 2014). These numerical simulations only consider wave action, without fluvial flows or tidal currents, and they assume some other simplifications (Watanabe et al. 2014). Then, they only have simulated the second model of formation, defined by Zenkovich (1967). Nevertheless, they suppose a useful tool to understand the initial formation and evolution of paired



Paired Baymouth Spits, Fig. 4 Formation of baymouth spits at the Ribeira de Iguape River mouth (SE Brazil), after the fourth (a–d) and fifth (e and f) models (modified from Alcántara-Carrió et al., 2018) (Source of LANDSAT images: Instituto Nacional de Pesquisas Espaciais, Brazil)

baymouth spits, and it is expected that they could simulate the rest of the models of formation.

Summary and Conclusions

Paired baymouth spits are two convergent spits that partially or almost completely enclose a bay or river mouth. They have also been named paired spits, double spits, and baymouth barriers, but these other terms can result unclear.

They are distinguished from other coastal barriers with inlets and closing the bay or river mouth because baymouth spits show a convergent trend during their geomorphological evolution. Most of the paired baymouth spits are located in microtidal regions; some of them have been identified in mesotidal and macrotidal coasts, and any study case has been reported for megatidal coasts. Five models have been described for their formation: (i) break of a coastal barrier, (ii) convergent longshore drift in a narrow bay, (iii) one outlet with high longshore drift and hydraulic blockage,

(iv) one outlet with high hydraulic blockage of longshore drift, and (v) convergence of two outlets and their associated spits. All these models are based on a bidirectional longshore drift by waves, and most of them include a hydraulic blockage. Numerical modeling of hydrodynamics and sedimentary processes of paired baymouth spits is still incipient.

Cross-References

- [El Niño–Southern Oscillation \(ENSO\)](#)
- [Longshore Sediment Transport](#)
- [Spits](#)
- [Tidal Inlets](#)

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Paleocoastlines

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As R.W. Fairbridge (1992) noted, “Paleogeography is an exercise in imaginative insight. One needs first to observe and understand the dynamic physical processes and landforms and then transfer and reapply these concepts to past landscapes.” In essence, the study of paleocoastlines requires the application of fundamental concepts of geology. *Uniformitarianism* and its modern version *actualism* provide the conceptual basis for modern sedimentary environmental patterns to be used in analogous comparisons with older facies in the construction of paleogeographies. Equally Walther’s dictum on *the correlation of sedimentary facies* provides the conceptual basis for differentiating *possible* conformable lateral and vertical facies patterns from the *improbable* in the delineation of ancient coastal landscapes. Climate, including glacial and other cycles, tectonics and eustasy are long-term forcing factors. Nevertheless, knowledge of local relative sea-level fluctuations are requisite to precise paleogeographic reconstruction.

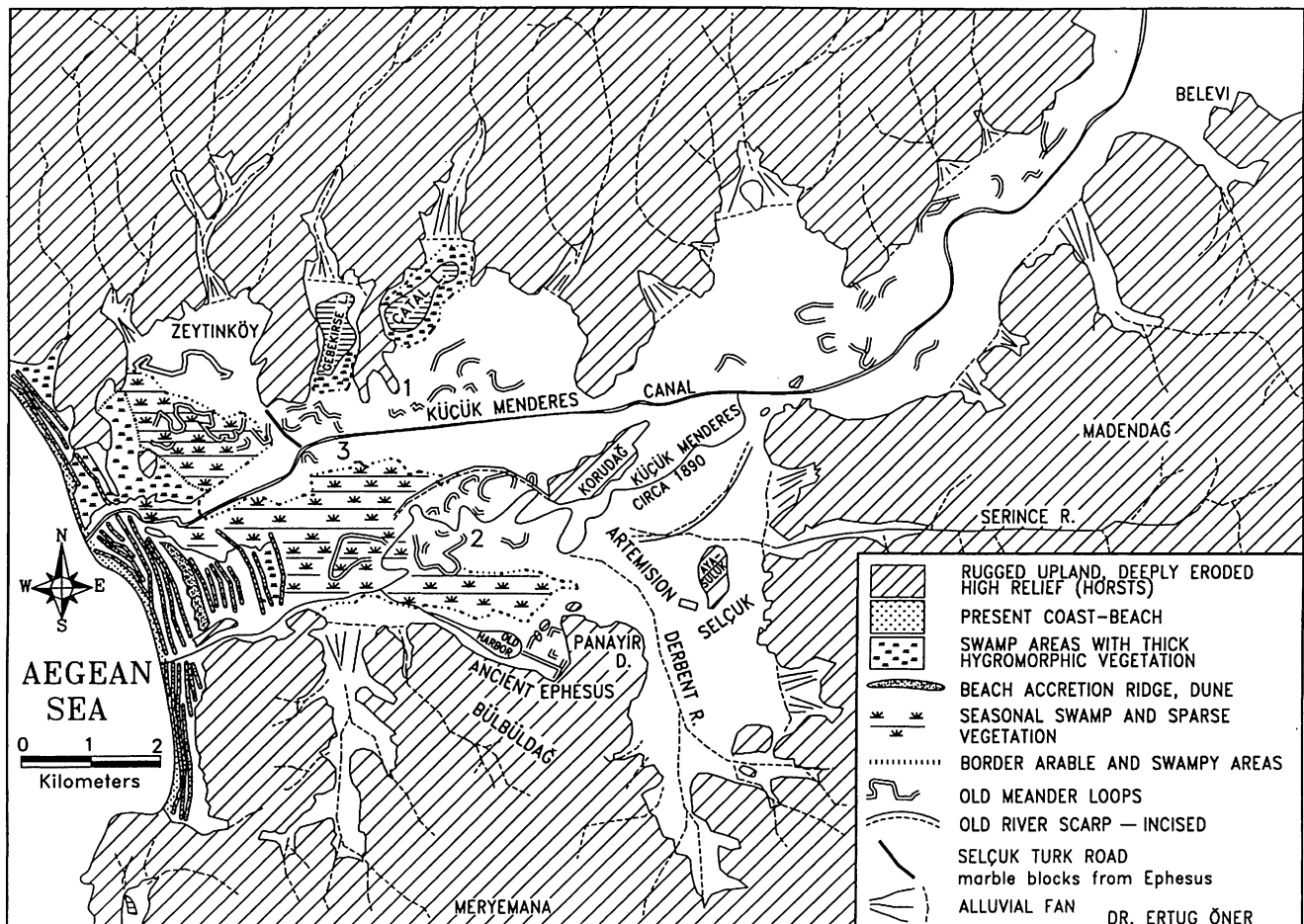
Major advances in paleoenvironmental reconstructions in the mid to late twentieth century were driven by the needs of the petroleum and coal industries for precision in sedimentary facies studies. For instance, Harlan Fisk, and later his students Hugh Bernard and Rufus LeBlanc, pioneered in intensive three-dimensional analysis of fluvial, deltaic, and coastal/nearshore marine facies. Emphasis was placed on sedimentary processes and depositional geometries of the various sedimentary environmental lithosomes, including lateral and vertical conformable and disconformable relationships. Studies of sedimentary environmental lineaments, geometries, internal sedimentary structures, and grain sizes and composition coupled with floral and faunal (fossil) analyses and ¹⁴C dates therefrom allow precise definition of ancient landforms, physical processes, water salinities, and their lateral and vertical distributions. These factors are a requisite to the accurate reconstruction of ancient coastal landscapes. Kraft and Chrzastowski (1985), Belknap et al.

(1994), and Kraft et al. (1987) as well as many other papers in the three referenced volumes of the Society for Sedimentary Geology (SEPM) provide many modern examples of sedimentary facies studies relevant to paleocoastline delineation.

Interdisciplinary Study of Paleocoastlines

Perhaps nowhere are paleocoastline morphologies of more interest than in the environs of historical and archaeological sites. Here, archaeology, history, and the Classics both complement and supplement each other in a synergistic manner clearly more effective than the sums of the disciplines applied separately. In the Aegean Sea, on the western coastline of Anatolia, sea level rose to its present position about 6000 BP. Since then, sea level dropped 1–2 m relative to land, rising

again to its present level. From the Würm glacial low stand to 6,000 BP fluvial and coastal environments were transgressed by nearshore marine environments, infilling valleys incised during Würm and earlier times. Since then, the Küçük Menderes (ancient Cayster River) has prograded 15 km seaward burying the marine sediments of the ancestral Gulf of Ephesus with a thin veneer of fluvial deltaic and barrier accretion plain sediments. The impact of this major change in coastal configuration had a profound effect on the peoples occupying the region from Neolithic time to present. Since the arrival of Ionian Greek colonists 3000 years BP, the residents of the ancient city of Ephesus, its many harbors over 3 millennia of time, the Artemision (Temple of Diana of the Ephesians, one of the seven wonders of the ancient world), and the later Turkish peoples of the region have been continually adapting to major changes in coastal environmental morphologies (Kraft et al. 2000).

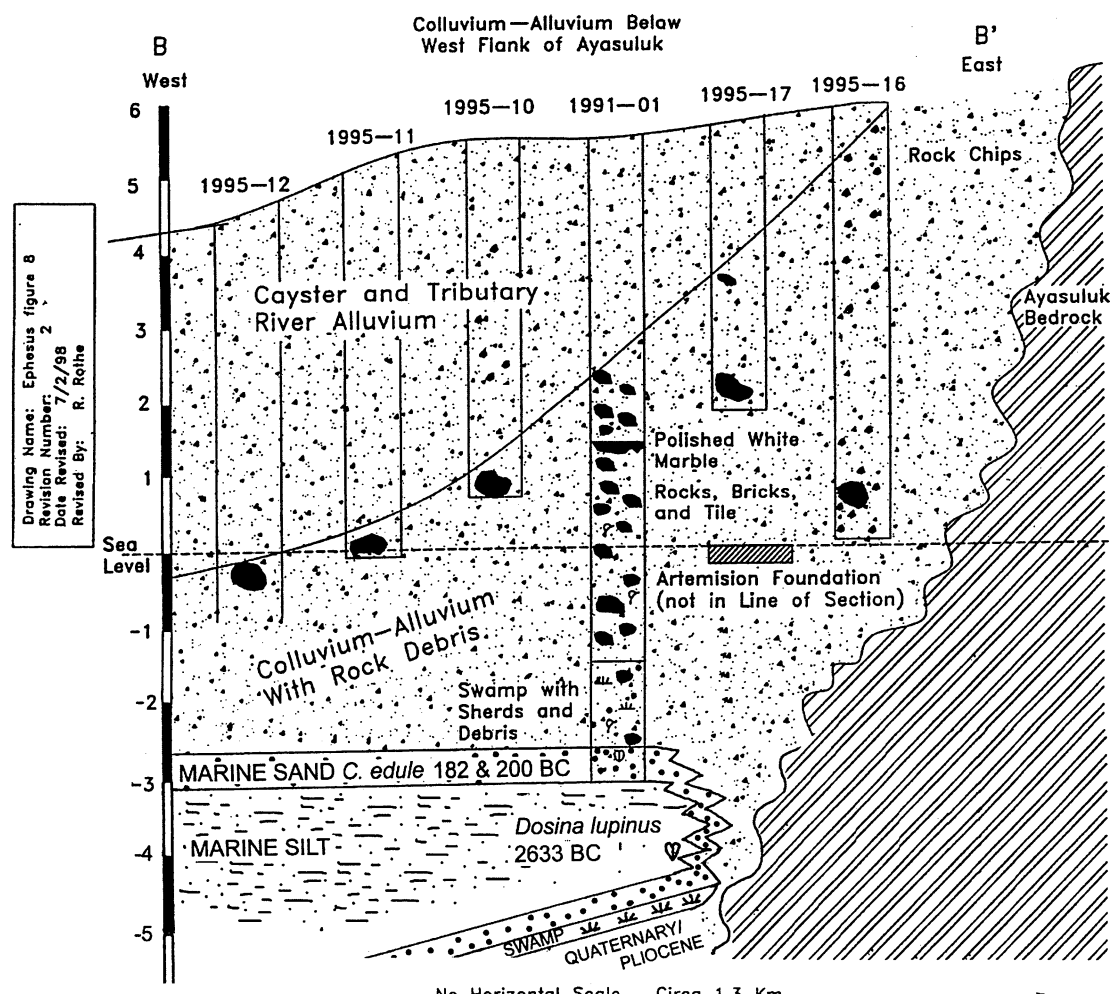


Paleocoastlines, Fig. 1 Holocene Epoch geomorphic elements of the Küçük Menderes (ancient Cayster River) in western Anatolia. (From Kraft et al. 2000. With permission, Österreichischen Archäologisches Institut)

The floodplain and delta of the Küçük Menderes have been much altered by twentieth century drainage and irrigation works. The coastal barriers form a damming effect that leads to widespread seasonal swamps and short-term floods in the wet season. Many palimpsests of morphic features may be observed from air photo, satellite imagery, and correlated with a carefully documented Austrian survey in the late nineteenth century (Fig. 1). Coastal barrier accretion ridges only occur in the lower 2.5 km of the delta-floodplain. Thus, in much of the earlier, relatively sheltered marine embayment from 6000 BP to Late Byzantine time (ca. 1,300 AD) fluvial-deltaic progradation was into a shallow marine prodelta shelf, probably with birdfoot like distributaries and flanking swamps, ponds, and marshes.

Details of sedimentary environmental lithosome geometries, sediment characteristics, micro- and macro-faunal and ^{14}C dates were attained by an intensive coring program (over 120 drill cores, Kraft et al. 2000, 2001). Figure 2 is a

schematic cross section based on 26 drill cores in the vicinity of the Artemision (Temple of Diana) constructed in the seventh century BC and finally destroyed by earthquake in the third century AD. The Artemision was built on an alluvial fan of a flanking river and eventually, after destruction, buried therein. However, we can also show that the site of the Artemision was at one time, pre-3000 BP, a coastline of the marine embayment. Therein lies an interplay with legend. *Callimachus: Hymn III to Artemis* tells us that "Amazons, lovers of battle set up a wooden image under an oak, in seaside Ephesos (II) and hippo offered a holy sacrifice to you; around the oak they danced you a war dance. . . . Afterward around that wooden image, wide foundations were built. . . ." Our paleogeographic evidence confirms the coastline under the Artemision site in Bronze Age times. Further, we can position this "wonder of the ancient world" in a coastal setting overlooking the ancient Gulf of Ephesus, as late as Hellenistic time, ca. 200 BC. Eventually, two sets of alluvium



Paleocoastlines, Fig. 2 A schematic cross-section through the environs of the Artemision (Temple of Diana of the Ephesians, one of the Seven Wonders of the Ancient World). Constructed mid-seventh century

BC, destroyed by earthquake third century AD, based on 26 drill cores. (From Kraft et al. 2000. With permission, Österreichischen Archäologisches Institut)

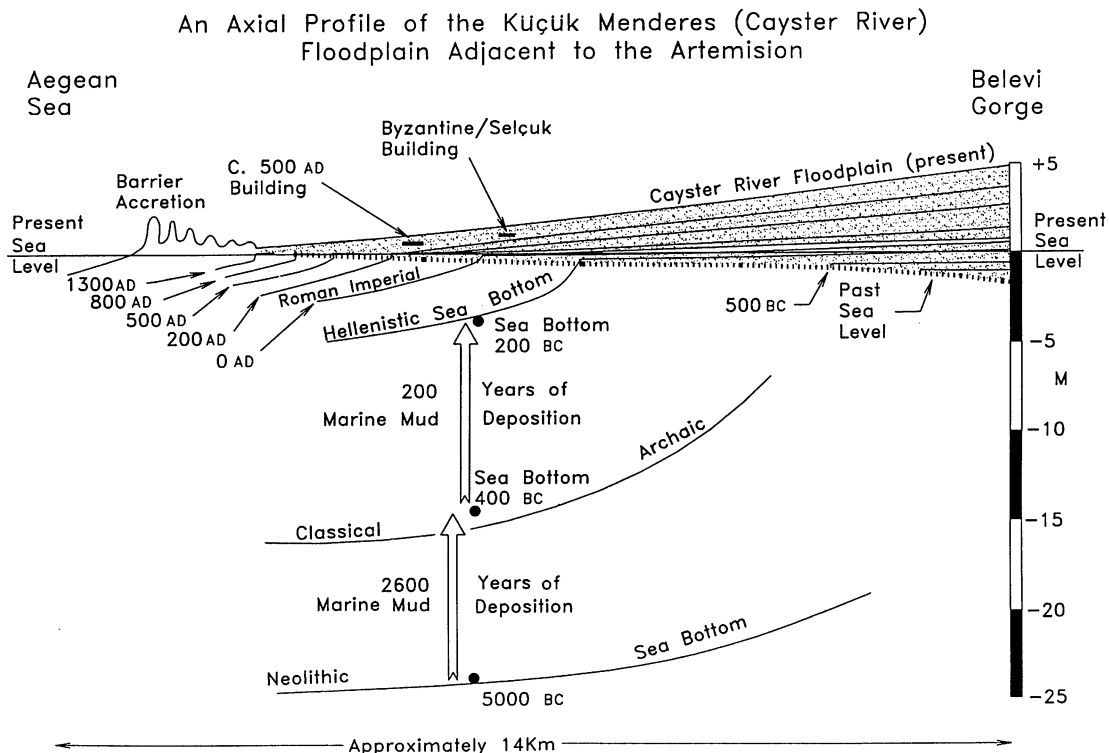
inundated and buried the ruins of the Temple of Artemis (as shown in Fig. 2).

Using drill core evidence of sedimentary facies in their lateral and vertical distribution, we can show that the major portion of the Holocene strata that infilled the ancestral valley of the Küçük Menderes River was deposited in an open marine embayment, initially clear water, followed by distal and nearshore prodelta silts (Fig. 3). Deposition was more rapid in the past 2400 years, as the delta prograded and alluvium aggraded. A few archaeological sites in the floodplain provide minimal dates for the delta advances.

Figure 4 is a summary delineation of delta coastline progradation for the southern portion of the former marine embayment. The Neolithic coastline shows the maximum extent of the marine embayment of mid-Holocene time, ca. 6000 BP, while, the Archaic coastline of 600 BC continues in an irregular configuration of distributaries, marsh, swamps, etc. until significant wave generated littoral transport of sands in barrier accretion started ca. AD 1300. The city of Ephesus varied in location over 4 km along the southern flank of the embayment, eventually being built on portions of the delta including massive fill. Many documents attest to the “conflict or fight” with the river progradation over more than a millennium of time.

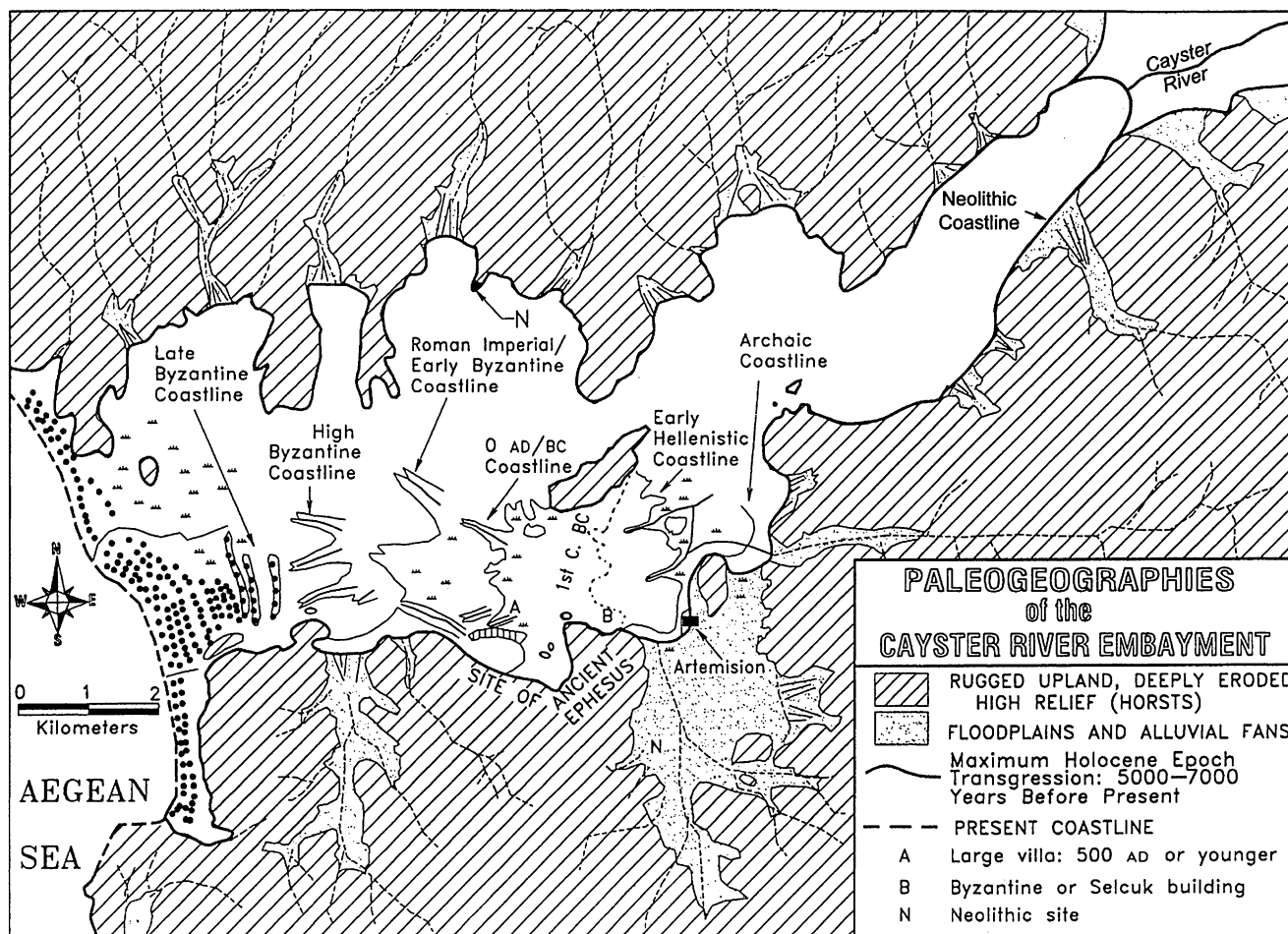
One of the clearest examples of negative environmental impact is a quote from Strabo (XIV 1, 24), “The city has both an arsenal and a harbor. The mouth of the harbor was made narrower by the engineers, but they, along with the King who ordered it, were deceived as to the result. I mean Attalus II Philadelphus (159–138 BC); for he thought that the entrance would be deep enough for large merchant vessels—as also the harbor itself, which formerly had shallow places because of silt deposited by the Cayster River—if a mole were thrown up at the mouth, which was very wide, and therefore ordered that the mole should be built. But the result was the opposite, for as the silt, thus hemmed in, made the whole of the harbor, as far as the mouth, more shallow. Before this time the ebb and flow of the tide would carry away the silt and draw it to the sea outside.”

The literature abounds with writings of the problems that the ancient Ephesian peoples had with maintenance of their harbors over the millennia of existence of Ephesus, capital of the ancient Kingdom of Asia (Kraft et al. 2000, 2001). Clearly interdisciplinary approaches to the study of paleocoastlines increases our abilities to precisely delineate ancient coastal configurations.



Paleocoastlines, Fig. 3 Cross-section of the lower Küçük Menderes (Cayster River) delta-floodplain. Seven thousand years of depositional valley infill including 25 m of marine/estuarine sediments covered by a

thin veneer of prograding floodplain alluvium and barrier accretion plain sands. (From Kraft et al. 2000. With permission, Österreichisches Archäologisches Institut.)



Paleocoastlines, Fig. 4 Paleocoastline progradation in the Küçük Menderes (Cayster River) embayment of the Aegean Sea in the vicinity of the ancient city of Ephesus and the Artemision (After Kraft et al. 2001)

Cross-References

- [Archaeology and Sea-Level Change](#)
- [Changing Sea Levels](#)
- [Coastal Changes, Gradual](#)
- [Coastal Sedimentary Facies](#)
- [Coastal Warfare](#)
- [Coastline Changes](#)
- [Holocene Coastal Geomorphology](#)
- [Shoreline and Coastal Terrain Mapping](#)

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Paraglacial Coasts

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Definition

Paraglacial coasts are defined as those occurring on or close to glaciated terrain, where glacial erosion features or deposits have a recognizable effect on the nature and evolution of coastal morphology and sediments (Forbes and Syvitski 1994). They thus encompass a range of coastal environments and landforms from proglacial fjord and outwash coast-lines to sand- and gravel-dominated coasts deriving their sediment supply from glacial deposits, but where the glacial imprint is older and subdued. Such coastlines also occur in settings where other phenomena such as a *periglacial* and related processes (including permafrost, ground ice, thaw subsidence, sea ice) or *human impact* (*q.v.*) may appear to dominate the coastal landscape. Essential features of paraglacial coastlines include a sediment source or physiographic setting in which the glacial origin exerts a continuing influence on the form and evolution of the coast.

Geographic Extent

The geographic range of paraglacial coasts is broadly coincident with the maximum extent of the last major continental glaciation in the *Pleistocene Epoch* (*q.v.*), approximately 20,000 years ago (Fig. 1). The most extensive paraglacial coasts (broadly defined by ellipses in Fig. 1) are those of the Arctic circumpolar region, including parts of the North American Pacific and Atlantic coasts, Greenland, Iceland, north-west Europe including the Baltic, and many parts of the Arctic

coasts of Canada, Norway, and Russia. Most of the Antarctic coast, where not presently glaciated, forms an equivalent southern circumpolar paraglacial region. Two other areas in the Southern Hemisphere showing extensive glacial influence on coastal development are the South Island of New Zealand and the west coast of South America in southern Chile, extending to Tierra del Fuego, where the southernmost coast of Argentina is also included.

Recognition and Terminology

The term *paraglacial* was originally applied to river systems (Church and Ryder 1972). In this context, it is used to describe the effects of disequilibrium excess sediment supply following deglaciation and the sub-sequent relaxation or decrease of sediment supply through surface reworking and downstream transport with time (cf. broken line in Fig. 2b). While he did not use the term *paraglacial*, Johnson (1925) recognized the distinctive nature of coastal morphology directly arising from glacial erosion or indirectly sourced from glacial deposits such as moraines, drumlins, and outwash sediments. The term was adapted in the early 1980s for application to coasts where glacial sediment supply or morphology remain dominant controls (Forbes and Syvitski 1994). Since that time, it has been most frequently employed in relation to coarse-clastic beaches and barriers sourced from glacial deposits (e.g., Forbes et al. 1995; Orford et al. 1996) but has also been applied to sediment sinks in coastal embayments (e.g., Carter et al. 1992; Shaw and Forbes 1992) as well as to glaciated shelves.

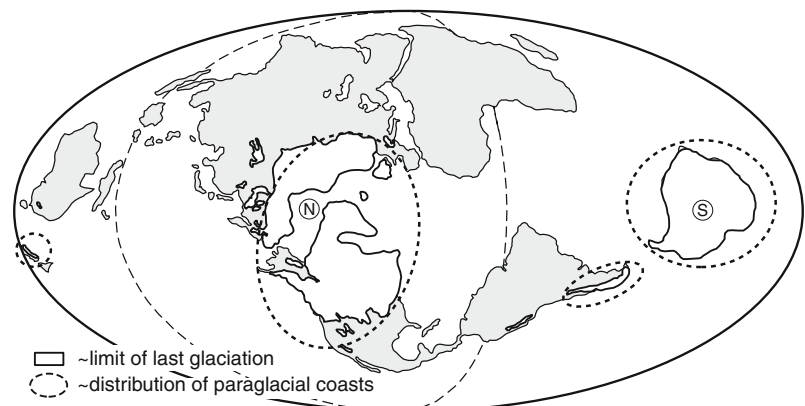
Classification

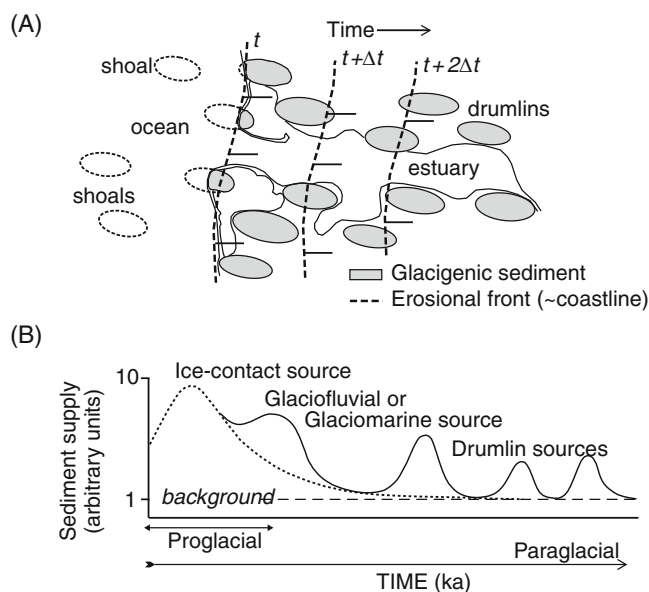
Paraglacial coasts can be classified into nine broad categories:

- sediment-starved coasts dominated by glacial erosion – fjords, overdeepened basins, and coastal barrens;

Paraglacial Coasts, Fig. 1

Approximate extent of last major glaciation and the geographic distribution of paraglacial coasts





Paraglacial Coasts, Fig. 2 (a) Schematic illustration of marine transgression and passage of a coastal erosional front through a drumlin field. (b) Schematic time variation of sediment supply in paraglacial coastal systems. (Modified after Forbes and Taylor 1987)

- depositional systems in confined fjord-head and proximal basin settings;
- open outwash coasts (not embayed or confined);
- basin-margin or open coasts with abundant glacigenic sediment in coastal cliffs;
- drumlin coasts characterized by sediment source switching and time-varying supply;
- embayed coasts with sediment supply controlled by sea-level change;
- coasts with patchy glacial cover or terraced outwash sediments – supply limitation or restricted access;
- forced-regressive coasts with limited sediment supply; and
- boulder-rich coasts in otherwise sediment-starved settings.

It is obvious from this list that the geomorphic expression may be highly variable and that sediment supply can range from negligible to abundant. However, apart from cases of significant fluvial or glaciofluvial sediment input (fjord-head deltas and outwash coasts), previous work on paraglacial systems has emphasized the rationing of sediment supply from glacigenic sources as a major control (and often the dominant influence) on coastal evolution in paraglacial settings. In this entry, we consider two examples of time-varying sediment supply and its implications for morphosedimentary outcomes at the coast.

Sediment Supply Control

Coastal morphology and deposits are strongly dependent on the geological setting, climate, oceanographic environment,

coastline orientation and exposure, and the rate and sign of relative sea-level change. Sediment supply is another critical variable, often in paraglacial settings the dominant factor defining the geomorphological response (Orford et al. 1996). If sufficiently large, sediment supply may counteract the transgressive effects of relative sea-level rise and produce a regressive sequence on a prograding coast (Forbes et al. 1995).

Persistent changes in relative sea level, related to post-glacial *eustasy* (*q.v.*), *hydro-isostasy*, and *glacio-isostasy*, are near-ubiquitous on paraglacial coasts, where they exert a profound influence on coastal morphology and development, including control of delayed shore-zone access to glacigenic sediment sources. However sea-level change is neither fundamental nor unique to paraglacial environments and may occur on all types of coast throughout the world. The implications of these relative sea-level changes for coastal evolution in paraglacial systems include highly varying rates of landward retreat or seaward advance of the coastline, and significant variation in the rate of access to glacigenic sediment supplies.

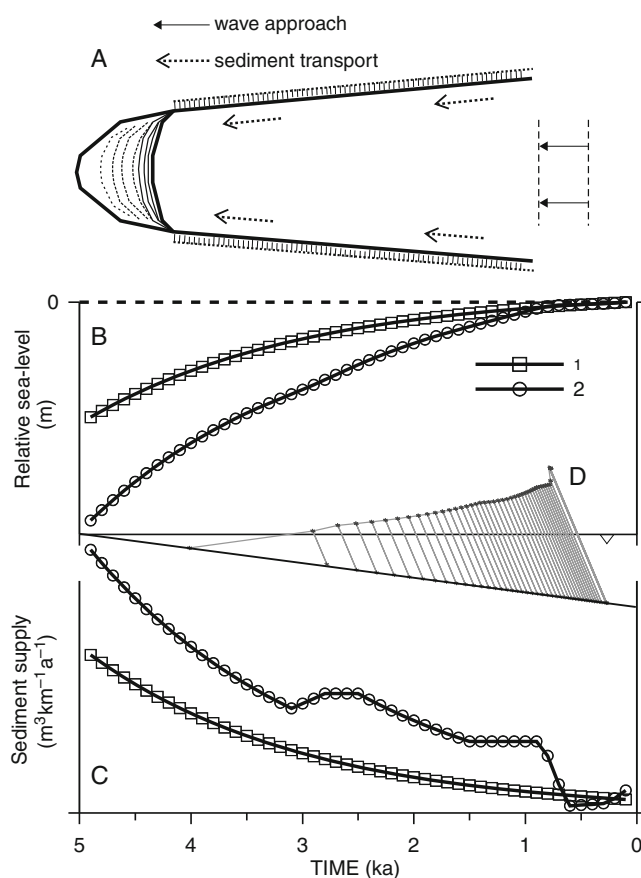
On paraglacial coasts, sediment is predominantly derived from erosion of glacial deposits. In many cases these represent the only significant source. This has profound implications for the nature of coastal sedimentary facies and shore-zone morphology. Grain-size distributions in glacial deposits run the full gamut from clay to large blocks the size of houses. Glaciolacustrine and glaciomarine facies may contain a large proportion of clay and silt (glacial flour) with small proportions of sand and minor fractions of ice-rafted gravel. In other cases, the sediment source facies may include glaciofluvial outwash, till, or other ice-contact diamicts (Fig. 2). The grain size in ice-contact sediments can range over more than six orders of magnitude (Forbes and Syvitski 1994) and coarse clastic components (pebble and coarser) are often abundant. This leads to a preferential tendency for the development of gravel (including sand-pebble, cobble, and boulder) beaches on paraglacial coasts. Where the underlying bedrock is a soft sandstone or similarly friable lithology with a large proportion of sand, the ice-contact facies may be relatively sandy and large volumes of coastal sand derived from both bedrock and the overlying glacial units can feed extensive sandy beaches, barriers, nearshore bars, coastal dunes, and tidal inlet deposits, as in the southern Gulf of St. Lawrence.

The spatial disposition of source sediments on paraglacial coasts also exerts a major influence on patterns of coastal evolution. Landward movement of a coastal *erosional front* across a glaciated landscape (Fig. 2a), as occurs with marine *transgression* under rising sea level, can lead to rapid mobilization of glacial deposits. The volume and spatial distribution of sediment may be discontinuous and vary widely,

from a thin till veneer to larger volumes concentrated in distinctive glacial landforms such as eskers, kames, deltas, moraines, and drumlins. In such cases, the start of erosional incision (such as when barrier retreat exposes a back-barrier hillslope to wave action for the first time) can mark the beginning of significant sediment release. In the case of marine transgression through a drumlin field or multiple moraine complex, this process may involve repetitive injections of finite sediment volume into a coastal region (Fig. 2b), spawning cyclic patterns of littoral-zone deposition with barrier growth, supply depletion, barrier breakdown, and landward transfer to begin the cycle over. This drumlin coast model has formed one of the dominant paradigms for understanding paraglacial coasts over the past 15 years (e.g., Boyd et al. 1987; Carter et al. 1990; Forbes et al. 1995; Orford et al., 1996).

Prograded bayhead barriers provide another example of time-varying sediment supply as a dominant control on coastal geomorphology. Experiments with numerical simulation of barrier growth in confined embayments with a thin till veneer provide useful insights on the role of relative sea-level change and sediment supply. Published examples (e.g., Forbes et al. 1995) are based on coastal embayments in southeast Newfoundland. The plan view (Fig. 3a) shows a narrowing embayment with ocean wave approach up the axis of the bay to a barrier complex at the head. Breaking of refracted waves along the shore is assumed to mobilize all un lithified glacial cover within reach of the waves and transport it alongshore to the head of the bay. Assuming simple geometry, the supply of sediment to the bayhead barrier is a function of access to erodible material, controlled by the rate of rise in relative sea level through time. The more rapid the rate of relative sea-level rise (Fig. 3b), the greater the sediment supply; subtle variation in the rate of sea-level change can produce significant variation in the supply (Fig. 3c). Progradation of the bayhead barrier (Fig. 3d) is a function of sediment supply and accommodation space (increasing depth and bay width). Therefore, with diminishing relative sea-level rise, typical of the last few millennia in Newfoundland, the rate of simulated barrier progradation decelerates in response to reduced sediment input. As the progradation rate decreases, the potential for severe storm impact leading to overtopping and growth of the active beach ridge may increase. The morphological outcome is a seaward-rising sequence of barrier ridges, flooded at the back and reworked into a high transgressive storm ridge at the front.

These examples demonstrate the critical importance of sediment supply and the ways in which glacial inheritance may direct and ration the delivery of sediment to the coast. They also show the propensity for high spatial and temporal variation of sediment supply in some paraglacial settings. Some systems suffer from slow supply reduction, with



Paraglacial Coasts, Fig. 3 Numerical simulation of beach-ridge barrier growth in a bayhead setting. (a) Schematic plan view of embayment. (b) Arbitrarily selected relative sea-level histories. (c) Simulated sediment supply to bayhead barrier corresponding to the two sea-level histories shown above. (d) Model output showing schematic section through simulated beach ridges with high transgressive storm ridge at seaward margin

distinctive morphological outcomes (Fig. 3). Others experience highly fluctuating sediment budgets (Fig. 2), resulting in abrupt coastal changes when supply from a given source is exhausted or other stability thresholds are exceeded (Forbes et al. 1995). Delayed paraglacial sediment input to coastal systems, ultimately controlled by access through changing sea level, can prolong the paraglacial signal, extending the relaxation time through many thousands of years. Coastal morphology in glaciated regions of the mid-to high latitudes over a large part of the globe is thus affected to this day by the physiographic and sedimentary legacy of the last major glaciation.

Cross-References

- [Eustasy](#)
- [Glaciated Coasts](#)
- [Pleistocene Epoch](#)

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Peat

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Peat forms from the accumulation of plant remains in waterlogged, anaerobic conditions that inhibit decomposition. In coastal situations peat forming environments occur within the intertidal zone, typically as *salt marsh* or *man-grove* peat, to the landward transition to freshwater environments, *wetlands* and *bog*. Accumulation requires a relatively low-energy environment, such as an *estuary* or behind *barriers* or *dunes*. Geographically the range is from the high latitudes to tropics, limited by water balance and the local conditions that constrain plant growth. If it is too dry, the plant material will decompose rapidly and peat will not accumulate.

Peat comprises authochthonous material, decomposed plant material deposited *in situ*, and allochthonous material,

that which is transported to the site of deposition. The latter includes organic material, plants and fauna, and minerogenic material, usually clay, silt, and sand size fractions but occasionally coarser. The organic component includes the visible, or macrofossil, remains and a fine organic matrix. Separate constituents of this matrix that are indistinguishable with the naked eye include totally decomposed organic material and microfossils, such as pollen, spores, diatoms, thecamoebians, and foraminifera (see Troels-Smith 1955, a classic work on the description of peat and other unconsolidated sediments, and recently summarized in chapter 11 of Jones et al. 1999). The balance between organic and minerogenic sediment covers the full transition between the end members, with the intermediates described by terms such as sandy peat (predominance of organics) and organic sand (predominance of minerogenics). Peat-forming plant communities exhibit a spatial zonation in response to a series of environmental gradients; the most important of which in the coastal zone is usually ground surface elevation with respect to the tidal regime (Waller 1994, chapters 3, 4, and 5). The spatial zonation changes through time, described as plant succession, as a result of autogenic and allogenic processes. Autogenic processes are those directional changes that operate within an ecosystem, trending toward equilibrium with the environment, the classical concept of climax vegetation. Allogenic processes are those operating externally to the vegetation. Examples include changes in sea level or sediment budget.

Where the balance of processes result in net accumulation, peat provides an excellent archive of environmental change. Successional changes that occur through time are recorded in the stratigraphic column. The physical, chemical, and biological properties of the sediment help to reconstruct the continuum from tidal flat, through marsh to wetland or bog. Although the classification into discrete units is difficult and somewhat arbitrary, it is relatively easy to reconstruct the trends or direction of change. This involves both the horizontal component of change (e.g., vegetation succession) and the vertical component (e.g., *sea-level change*). Early scientific studies, such as the geological surveys of coastal wetland areas carried out in the nineteenth century (or earlier) in northwest Europe, described the visible plant remains (macrofossils) within peat layers and from these inferred environmental change. With scientific advances during the twentieth century, an increasing range of techniques has been used to analyze peat (van de Plassche 1986). While macrofossils, especially the roots and stems of the plants that were growing in the sediment, provide extremely good evidence of site conditions at the time of formation, their preservation is very variable and because only a small volume of sediment may be available (e.g., from a borehole) they may not give a reconstruction of the whole community or wider environment (Waller 1994). From the microscopic sediment



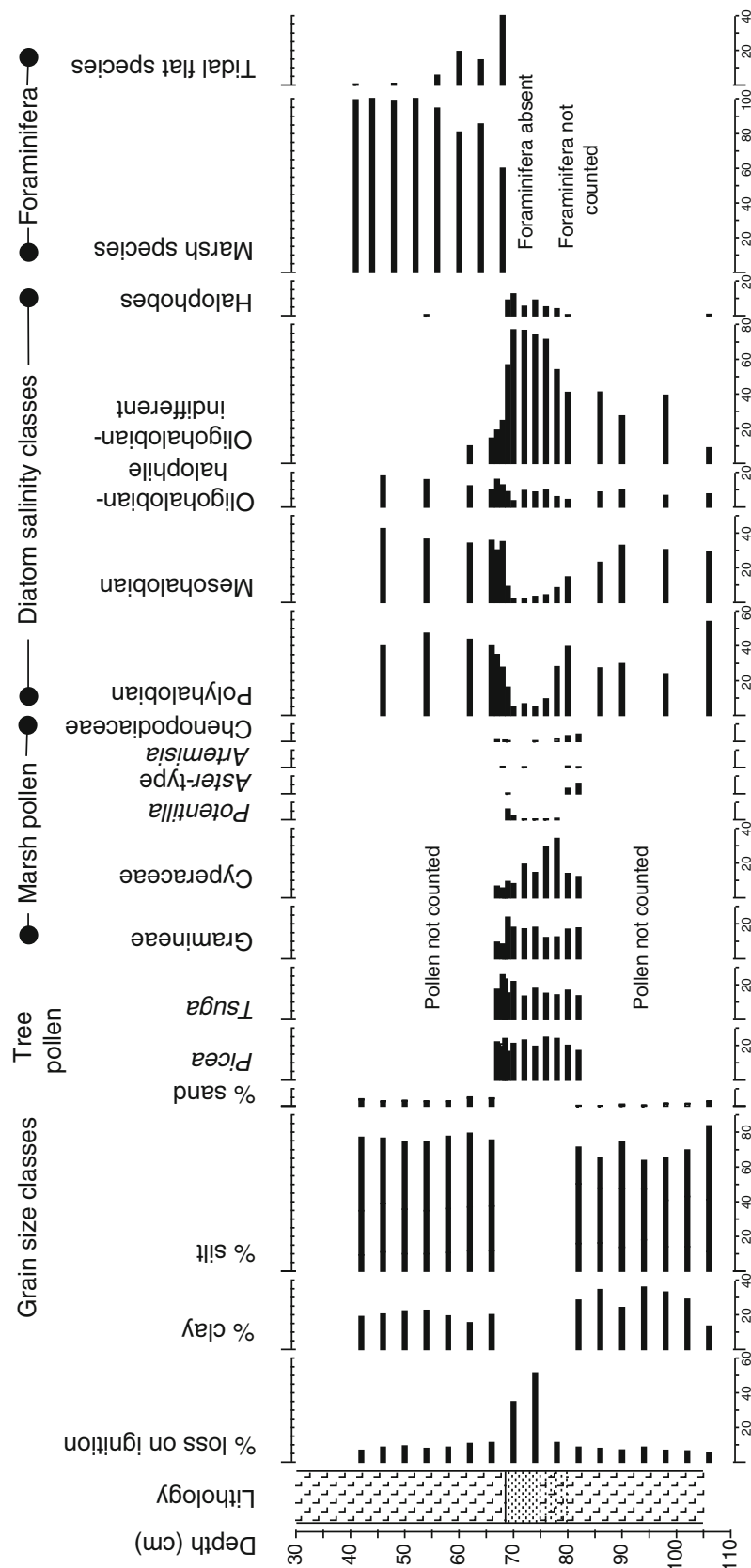
Peat, Fig. 1 Salt marsh and tidal flats at Girdwood, in the upper Turnagain Arm of the Cook Inlet, Alaska. The dead spruce trees are rooted in a peat layer that had developed over the tidal flat sediments visible in the exposure. Coseismic subsidence during the earthquake of 1964 dropped the forest to a level below high tides. Tidal inundation killed the trees and gray tidal silt accumulated over the peat and around

the tree trunks. At this location the silt is approximately 50 cm thick. The present salt marsh, dominated by sedges, represents the first stage in the renewed growth of peat as gradual land uplift since 1964 allows saltmarsh, and eventually, freshwater environments to develop once more

matrix the most widely studied element has been the microfossil content (especially pollen, spores, diatoms, and foraminifera).

Analysis of peat layers within a sediment sequence and the microfossils contained in the different layers can give reconstructions of a wide range of different environments, ranging from the coastal environments of previous interglacials, through the whole of the current interglacial to the present day. Peat can form quickly on the geological timescale. For example, the marshes around the Cook Inlet, Alaska subsided by up to 1.7 m during the great earthquake (M_w 9.2) of 1964 and were rapidly buried by tidal flat silt deposits (Fig. 1). The 1964 marsh peat is now exposed along the banks of tidal creeks beneath minerogenic sediments, more than 1 m thick in some places, that grade upwards with an increasing organic content to the present marsh surface where there is already a centimeter or more of peat accumulating. While this example of recent marsh burial was the result of rapid coseismic change in land level, the gradual submergence of marsh peat beneath tidal flat minerogenic sediment records relative sea-level change due to nonseismic processes, especially the interaction of *eustasy* and *isostasy* (e.g., Allen 2000).

Buried peats contain critical information on the long-term dynamics of coasts prior to the period of scientific observation and measurement. The threat of large earthquakes, comparable to the 1964 Alaska event, occurring in Washington, Oregon and British Columbia is clearly indicated by the multiple peat layers buried beneath many of the coastal marshes in these states (e.g., Shennan et al. 1996). Figure 2 shows a peat layer from the Johns River, Washington. The sequence from 105 to 69 cm shows the colonization of a marsh onto tidal flat. As the marsh peat accumulates, low-marsh plants, shown by the *Chenopodiaceae* and *Aster-type* pollen, are replaced by high-marsh plants (e.g., indicated by *Potentilla* pollen). At the same time, marine (polyhalobian) and brackish (mesohalobian) diatoms give way to increasingly freshwater (oligohalobian) and salt-intolerant (halophobes) diatom species. The top of the peat marks the rapid return to tidal flat sedimentation, interpreted as the result of an earthquake that caused about 1 m of ground subsidence. The radiocarbon age of the peat indicates this occurred about 300 years ago. Other peat layers from the estuary indicate a total of eight such events during the last 5,000 years. The earthquakes result from the collision of the Juan de Fuca and American plates.



Peat, Fig. 2 A buried peat layer from Johns River, Washington. (After Shennan et al. 1996). The development, then rapid burial caused by an earthquake ~300 years ago. The horizontal axes show percentage abundance for selected species for three microfossil types: pollen, diatoms, and foraminifera. The full data set comprises microfossils from over 100 different species

Cross-References

- [Barrier](#)
- [Bogs](#)
- [Coastal Soils](#)
- [Dune Ridges](#)
- [Eustasy](#)
- [Isostasy](#)
- [Mangroves, Geomorphology](#)
- [Salt Marsh](#)
- [Vegetated Coasts](#)
- [Wetlands](#)

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Photogrammetry

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Photogrammetry can be defined simply as the science of making reliable measurements from photographs. Unlike a map, however, a photograph contains a number of distortions that require correction before accurate measurements can be made. A number of photogrammetric techniques can be employed to remove these distortions and obtain useful measurements. In coastal studies, photogrammetric techniques are commonly employed to establish the positions of historical and modern features-of-interest (e.g., shorelines (defined as the high-water line or wet-dry boundary), cliff edges, dune positions, etc.). Historically, the focus of study has been overwhelmingly on the use of vertical aerial photography to

derive accurate shoreline positions, although photogrammetric applications using ground-based photography, videography and integration with other types of remotely sensed data (e.g., lidar) are becoming widespread. Most often, a time series of feature positions is compiled for the purpose of studying coastal dynamics, such as the evolution of geomorphic features, or determining rates of coastal change (e.g., shore erosion and accretion). These time series have also been used to guide the delineation of coastal erosion or flood hazard areas and building setback lines.

Historical Development

Early studies of coastal erosion (e.g., Stafford and Langfelder 1971) used point measurements made on individual air photos to determine shoreline rates of change. The procedure is straightforward: stable reference points such as buildings and road intersections are selected on air photos taken in different years, and the distance between these points and the shoreline reference feature (typically the boundary between wet and dry beach sand, which approximates high water) is measured. These measurements are then multiplied by the nominal scale of the aerial photograph to obtain ground distances. The change in distance over time yields a rate of shoreline change at that location. Subsequent shoreline delineation methods expanded upon this basic approach, using techniques such as photo-enlargement or a Zoom Transfer Scope to match a base map (e.g., Dolan et al. 1978, 1980; Smith and Zarillo 1990). These methods permitted a continuous representation of the shoreline to be mapped from the photos, rather than discrete points, and were a significant improvement over point measurements in terms of the amount of information that could be extracted from aerial photography.

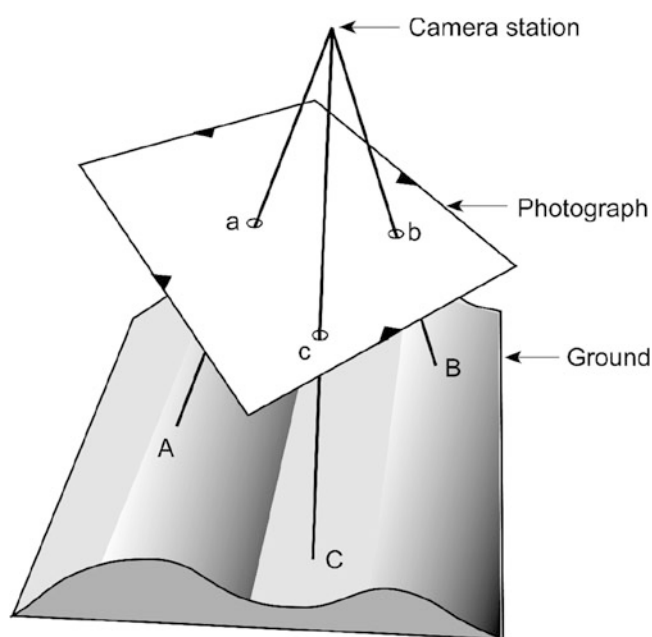
One of the primary limitations in early historical shoreline mapping studies was the inability to quickly and accurately correct aerial photographs for their inherent distortions. The first integrated coastal photogrammetry software package to address this need was the Metric Mapping System (MMS) (Leatherman 1983; Clow and Leatherman 1984). For aerial photographs, this package furnished the ability to correct individual photographs for image displacements and produce shoreline positions on a plotted map. The Digital Shoreline Mapping System (DSMS) (Thieler and Danforth 1994a, b) expanded upon this approach to include a full photogrammetric adjustment of large groups of photographs and produce data suitable for plotting shoreline change maps in a Geographic Information System (GIS). Both the MMS and DSMS software packages utilized *x-y* tablet digitizers to obtain vector data from aerial photographs. While this approach kept the amount of data acquired from the photographs to a manageable size, much useful visual information

was not used. More recently, however, advances in computer storage and processing capability allow photogrammetric techniques to be applied to digitally scanned photographs. This process, known as softcopy photogrammetry, has recently seen increased use in coastal studies. A comprehensive review of available shoreline mapping techniques is furnished by Moore (2000).

Characteristics of Aerial Photography

The basic principles involved in the extraction of geographic data from aerial photographs are derived from the geometric relationships between image space and object space. Image space refers to the world inside the camera (i.e., the photographic image and measurements obtained from it). Object space refers to real-world geographic coordinates outside the camera.

In distortion-free space, points in image space can be projected to points in object space. This relationship is based on the principle of collinearity: the perspective center of the camera lens (which is considered a point), an image point and its corresponding ground point all lie on the same straight line (Fig. 1). Aerial photographs, however, are subject to a number of distortions introduced at various stages in the photographic process that perturb the collinearity condition. These perturbations affect both image space and object space.



Photogrammetry, Fig. 1 In distortion-free space, a projective relationship exists between image space (points on an aerial photograph) and object space (points on the ground). The camera station, an image point (a, b, c), and its corresponding ground point (A, B, C) all lie on the same straight line. (After Thieler and Danforth 1994b)

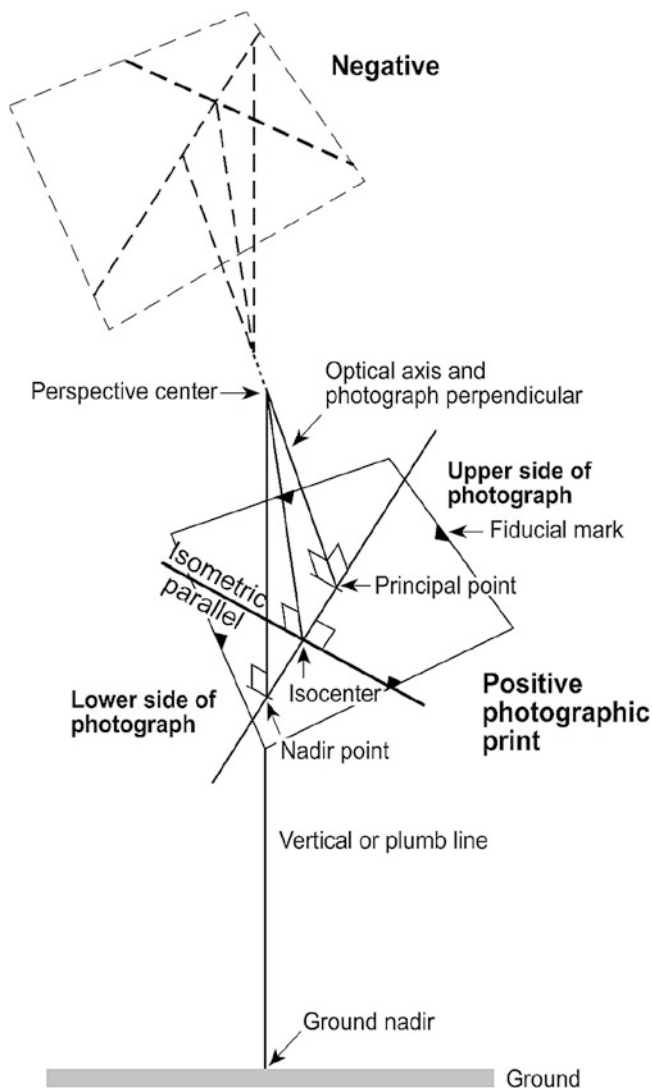
The image space coordinate system is defined by the locations of the fiducial reference marks on a photograph, the calibrated focal length, and the geometric distortion characteristics of the lens system in the camera. Image space is described by a three-dimensional, rectangular Cartesian coordinate system with the origin located at the principal point (the center of an aerial photograph as defined by fiducial reference marks around the edge of the photograph). For aerial photographs, the x -axis is typically positive in the direction of flight. The z -axis corresponds to the optical axis of the camera.

The distortions affecting image space result from lens distortion and film deformation. All camera lenses have measurable distortions and optical defects that affect the representation of image points on film. Lens distortions can be radial or tangential. Radial distortion is symmetric around the principal point and is caused by optical defects in the lens. Tangential distortion is symmetric along a line through the principal point and results from the lens being slightly off-center in the camera. In a well-adjusted camera, however, only radial distortions are present.

The magnitude of lens distortion is highly variable. Some lenses used today have up to 0.110 mm radial distortion (American Society of Photogrammetry 1980). Similar and sometimes greater amounts of lens distortion commonly are present in photographs taken prior to World War II, which brought an increased demand for accurate photography and improved lens manufacturing techniques. This is a particularly important consideration in using historical coastal photography. In most modern camera systems, however, lens distortion is negligible.

Two types of film deformation exist. Deformation can be introduced in the camera during the aerial survey or in subsequent processing. Film buckling, for example, may occur during the photographic survey due to irregularities in temperature, humidity or film spool tension in the camera. Further deformation is introduced not only in the development of the original negatives, but also in each generation of prints and transparencies (typically used by coastal researchers) made from the original negatives. The end result of these deformations is a photograph that no longer represents accurately the true geometric relationships between the fiducial reference marks and image points in the photo.

In addition to deformation occurring in the camera, the amount of film deformation present in a given photograph depends upon the age and type of material (glass, film, or paper), processing techniques used, and the temperature and humidity at the time measurements are made. Standard diapositive (transparency) film is generally stable within 0.005 mm. Photographic paper, however, is far less stable and may change in size up to 1% during processing alone (American Society of Photogrammetry 1980).



Photogrammetry, Fig. 2 Definition sketch of terms used to describe the various elements of a tilted aerial photograph. Some degree of tilt is always present in an aerial photograph, which causes image points to be displaced from their true position. (Modified after American Society of Photogrammetry 1980)

The characteristics of object space cause image points on film to be displaced (as opposed to distorted) from their true position as a result of three factors: relief displacement, tilt displacement, and atmospheric refraction. Relief displacement is caused by changes in ground elevation within a photo that cause objects closer to the camera to be larger (i.e., at a larger scale) than those farther away. Relief displacement takes place radially from the nadir (Fig. 2). Objects higher than the ground elevation at the point where the nadir intersects the ground (the ground nadir; Fig. 2) are displaced outward; objects lower than the ground nadir point are displaced inward.

The determination and magnitude of relief and tilt displacements have been widely discussed in the context of

coastal photogrammetry (e.g., Anders and Byrnes 1991; Crowell et al. 1991). Tilt displacement occurs due to the inability to keep the aerial camera perfectly leveled during photography. Some degree of tilt is always present in an aerial photograph. On a tilted photograph, the sense of displacement depends on whether the image point is on the low or high side of the isometric parallel (Fig. 2). Points on the low side of the isometric parallel are displaced outward from the isocenter; on the high side they are displaced inward. Points on the isometric parallel are not displaced.

Atmospheric refraction (the bending of light rays through the atmosphere) also causes photograph image points to be displaced. The displacement occurs radially outward from the nadir. The magnitude of the displacement depends on the aircraft flight height, direction of the optical axis relative to the ground, and the focal length of the camera. For most coastal applications, atmospheric refraction is negligible because low-altitude photography is used.

Extracting Geographic Data from Aerial Photographs

The traditional approach to analytical photogrammetry (American Society of Photogrammetry 1980) is composed of three steps that remove the perturbations described above, and exploit geometric relationships between overlapping aerial photographs to extract geographic data: (1) preprocessing; (2) aerotriangulation; and (3) postprocessing. Preprocessing reduces measured image coordinates to the image space coordinate system described above, as well as removes systematic errors such as lens distortion and film deformation effects. The goal of preprocessing is to transform measured image data from a photograph to an idealized image space in which distortions do not exist. In other words, the measured coordinates are refined to remove image space distortions before the data are used in subsequent processing.

Aerotriangulation is used to solve simultaneously for the camera position of each photograph in a large group of overlapping photographs (a strip or a block), as well as the coordinates of unknown ground points. The absolute orientation performed in an aerotriangulation adjustment is basically an extension of the technique of space resection, which determines the six elements of exterior orientation for a photograph, including the position (latitude, longitude, and elevation), and attitude (roll, pitch, and yaw; designated ω , ϕ , and κ , respectively) of the aerial camera. Atmospheric refraction effects can also be removed during space resection or aerotriangulation by applying a correction function each time the orientation of the camera is updated in the solution process.

Postprocessing typically involves transforming the camera position information into instrument settings used in

stereoplotters or other photogrammetric equipment in order to compile basemaps or generate rectified photographs (also known as orthophotographs). In digital photogrammetry, camera information is input into computer software that performs orthorectification on scanned photographs. The result is a geographically corrected photograph that may be used directly in a GIS.

Ground Control

The common requirement of all forms of photogrammetric manipulation is adequate ground control information so that quantitative, geographically accurate information can be extracted from the photographic data. Thus, establishing an adequate ground control network is critical. This is particularly true in studies using historical aerial photography, since points on the ground are the only source of position information.

A ground control network is a set of points that appears in one or more photographs, and provides the basic means of establishing a correspondence between the photographs and the ground. There are two types of points used in photogrammetry: ground control points and tie points. A point appearing in one or more photos for which information about its location is known (e.g., one or more of latitude, longitude, and elevation) is called a control point or ground control point. The image space coordinates of ground control points and their corresponding geographic coordinates are used to establish the collinear relationship between image space and object space (Fig. 1).

There are a number of ground control data sources, including maps, field surveys and geodetic control tables. The locations of well-defined points shown on maps can be digitized, converted to geographic coordinates and used as ground control points. These points are termed "supplemental control points" because they are obtained from a map rather than by direct field survey. Supplemental control points, consisting primarily of buildings and road intersections, are the primary source of ground control for many historical shoreline mapping projects because other data are unavailable.

A tie point is defined as a point appearing in two or more photos, for which a corresponding ground position is not known. Tie points are used to "pass" or extend control between overlapping photos. These points are used in addition to ground control points to establish the relative orientation of photos to each other, such as is done when viewing a pair of overlapping photos through a stereoscope. Tie points commonly include features such as trees, buildings, and road intersections.

For most applications, at least four and preferably six to nine tie points are needed to provide adequate control for a

given stereo-pair, and should be distributed throughout the photograph. Only a few control points are needed to establish the geographic orientation of a group of photographs in object space; most points used to control a photograph may simply be tie points.

Ideally, the exact planimetry (latitude, longitude, elevation) would be known for a large number of spatially and temporally well-distributed ground control points throughout all photographs. This is never the case, however, and ground control points of varying quantity and quality must be used when constructing the control network for a given mapping project.

In historical shoreline change studies, when several sets of photographs spanning several years of the same geographic area are used, the common ground control and tie points form a model that establishes correlations between images, in space and through time, that are readily exploited in aerotriangulation and error analysis.

It is often problematic, however, to furnish an adequate quantity and distribution of ground control and tie points due to the nature of photography along the shoreline and changes in coastal environments over time. Most photographs that include the shoreline, for example, are devoid of usable control points seaward of the shoreline. Coastal areas may also change rapidly over time, due to natural processes or human development, which reduces the number of stable points that can be used as ground control or tie points. In these situations, it is often necessary to use additional overlapping photos taken of more landward areas in order to augment the control network for the shoreline photographs.

Modern Techniques and Current Investigations

Several different methods of digital rectification (Table 1) are available through the various softcopy photogrammetry software packages currently available. Not all rectification techniques, however, are equally accurate. The two most robust rectification methods are full orthorectification based on viewing and manipulating digital stereo-pairs, and pseudo-rectification, in which stereo-pairs are used but there is no ability to manipulate stereo-imagery. These two techniques both apply photogrammetric principles to produce orthophotographs, and involve the creation of a Digital Terrain Model (DTM) to remove relief displacement from the imagery. Generation of DTMs involves taking the userinput measurements (ground control and tie points) and deriving the DTM from a stereo model through interpolation (Ackerman 1996). In many softcopy photogrammetry systems, DTM collection is automated but the user can edit the model. Editing a DTM (which is only possible with stereo-viewing photogrammetry) increases the accuracy of the ground surface model, allowing more complete removal of relief

Photogrammetry, Table 1 Comparison of softcopy photogrammetry processing techniques

Technique	Process (result)	Products
Full orthorectification (stereo-viewing)	Input camera system parameters (remove camera distortions) Input tie points (remove aircraft attitude distortions) Add ground control points (georeference imagery) Create DTM (remove relief distortion) Edit DTM (correct for detailed topography) Aerotriangulate (tie all photos together, orthorectify)	Orthophotographs Stereo imagery High-accuracy DTMs
Pseudorectification	Input camera system parameters (remove camera distortions) Input tie points (remove aircraft attitude distortions) Add ground control points (georeference imagery) Create DTM (remove relief distortion) Aerotriangulate (tie all photos together, orthorectify)	Orthophotographs Uneditable DTMs
Single frame resection	Input camera system parameters (remove camera distortions) Add ground control points (georeference imagery) Least squares resample (statistical best-fit)	Pseudo-orthophotograph (relief distortion not removed)
Warping (rubber sheeting)	Add ground control points (georeference imagery) Stretch pixels to best fit approximation	Georeferenced image

displacement in the orthophotograph. Using an unedited DTM may affect the spatial accuracy of the orthophotograph, and may not approximate rapid changes in topography. Another rectification technique, single frame resection, can be used to remove distortions associated with the camera system, but does not remove aircraft attitude or relief displacements. The final method available is warping (also known as “rubber sheeting”). In this case, no parameters are input to remove displacements; instead, ground control points are used to “stretch” the pixels of an individual image to a best georeferenced fit.

In comparing the various methods described above in coastal change applications, recent studies (Hapke and Richmond 2000) have found that a shoreline position digitized on a stereo-based orthophotograph is the most reliable in terms of reproducing shoreline position. The technique of warping is the least reliable method and can be very inconsistent. In areas of high relief, or where high accuracy (submeter) is required, use of a stereo-edited DTM is recommended.

The recent proliferation of digital photogrammetry and GIS software has greatly increased the accessibility of aerial photograph analysis. Many software packages can be run on a standard desktop computer and do not require formal training in photogrammetry or GIS. Using such software requires scanning the aerial photographs to convert to digital format. There are several options as to the media of the aerial photography and the choice of scanners, all of which may have noticeable impact on the spatial accuracy of the final orthophotograph.

The first step in digital aerial photograph analysis is to convert the photograph into a digital format using a scanner. The most accurate choice, a geometrically correct

photogrammetric scanner, will produce minimal distortion. This ensures that the spatial distribution of objects in the original photography is reproduced accurately in the digital version. Photogrammetric scanners provide the most accurate digital imagery, but these scanners and/or the contracting of photogrammetric scanning may be too costly for smaller-scale coastal mapping projects. As a less-expensive (and less-accurate) substitute, high-quality graphic design scanners or desktop scanners are available. Graphic and desktop scanners are not designed to assure geometrically accurate data conversions, and thus may introduce distortion to the digital imagery, which results in nonsystematic positional errors in the resulting orthophotograph (Hapke et al. 2000). A further limitation of desktop scanners is that the scanning resolutions rarely exceed 800 dpi (dots per inch), whereas most photogrammetric workflows recommend 1200 dpi or higher.

Standard aerial photography is typically available from the original negatives in two different media: contact (paper) prints and diapositives (positive film transparencies). Although contact prints cost less and are readily available, as noted above photographic paper may undergo stretching, shrinkage, and warping, resulting in positional errors. Diapositives are a much more stable media, since film is much less likely to undergo distortion.

Most digital softcopy photogrammetry software packages provide an overall RMS (root mean square) value as an indication of error associated with rectification. US National Map Accuracy Standards (NMAS) (Falkner 1995) state that in order to conform to the accuracy standards, 90% of identifiable stationary objects should be accurate to within a specified RMS error. In a study comparing media types and scanners, Hapke et al. (2000) found that at a scale of 1:

12,000, the only combination of scanner and media type that conforms to NMAS is a diapositive scanned on a photogrammetric scanner. The RMS error increases with the lower precision scanner and with the use of contact prints versus diapositives. The NMAS standards also require that the maximum error at any given point, either horizontal or vertical, does not exceed three times the magnitude of the RMS. For images scanned on a nonphotogrammetric scanner (either graphic design or desktop), this value is almost always exceeded in the X and Y, and in every case in the Z direction. If a coastal mapping application involves delineation of absolute position (e.g., establishing setback lines, determining hazard zones at the scale of individual property boundaries, etc.), or if the resulting data could be used for such applications, then it is especially important to use a technique that adheres to the NMAS.

For other types of coastal applications, larger errors may be quite acceptable (e.g., rate calculations, especially over long periods of time), such that using contact prints and/or nonphotogrammetric scanners is appropriate to the study design. These methods, however, may introduce large nonsystematic errors into the final orthophotograph. These nonsystematic errors are most likely due to stretching or shrinking of the contact prints. Table 2 provides an overview of the recommended scanner and media choices for a variety of coastal research activities utilizing aerial photography. The image quality of scanned contact prints is significantly degraded as compared to scanned diapositives, regardless of the quality of the scanner. This may result in the inability to measure ground control and tie accurately.

The full stereo-orthorectification process requires the generation of a DTM that defines topography within the stereo-overlap region of images using a series of points and is required if the distortion from terrain relief is to be removed from the orthorectified images. Two formats of DTMs may be generated from stereo-images: grid or TIN (triangulated irregular network). The grid is a regularly spaced network of points where the elevation of individual points can be edited but the network spacing must remain constant, and thus the horizontal position of each point cannot change. In a TIN model, however, not only the elevation of points may be

edited, but they may also be added in areas where a greater density is desired, or deleted in problematic areas. Another advantage of TIN models is the ability to add breaklines, which allows for more accurate definition of subtle topographic changes. Breaklines are crucial to accurately define the topographic signal of narrow or sharp features such as beach scarps, cliff edges, or dune ridges in the surface model. A breakline is a manually entered line composed of a series of points that are incorporated into the DTM. Breaklines can only be added to the model while viewing in stereo, as the operator must be able to identify the elevation change in order to correctly place the line.

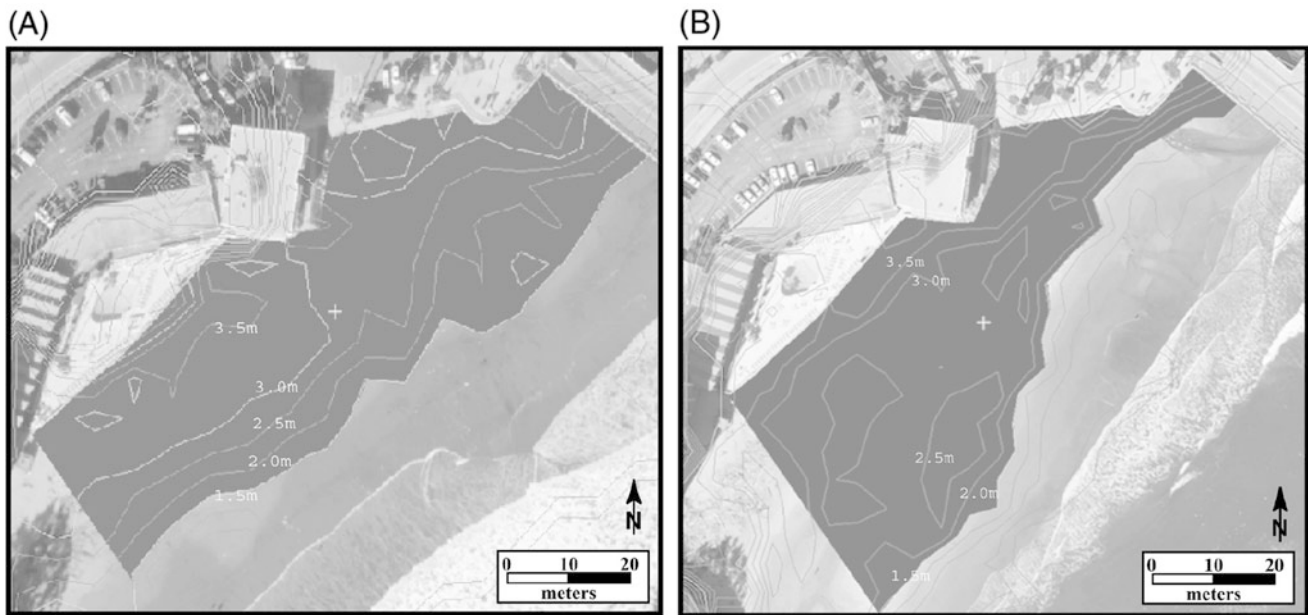
DTMs generated from stereo-models can be used to derive volume, and hence volumetric changes. However, the automated process of DTM generation has distinct, nonsystematic errors that must be corrected if DTMs are to be used for accurate surface modeling and volumetric change analyses. Automated collection with no editing is problematic in images of coastal regions due to typical low-contrast beach conditions and high visual interference areas like the swash zone, where the water has moved and thus varies in position from one image to the next. In order to utilize the DTMs for time-series volumetric studies, careful editing is crucial.

Beach volume (Hapke and Richmond 2000) and/or dune volume (Brown and Arbogast 1999) can be calculated using an edited TIN format DTM derived from each date of photography (Fig. 3). For the volume calculations, a base elevation must be defined, and 0 m (MSL) is typically used. However, surface elevations below the 0.5 to 1.5-m zone contour on a subaerial beach are commonly in, or very near, the swash zone that creates severe interference when viewing in stereo and thus greatly reduces the accuracy of the contours. Merging other topographic data in high interference areas or areas prone to shadow (e.g., base of seacliff, scarp, or dune) is highly desirable if such data is available. In addition, once ground control points are measured within a block or strip of air photos, other sources of DTM data can be used for final orthorectification. These sources include lidar as well as existing contour maps.

Technological advances in surveying techniques and digital data collection are rapidly influencing the field of photogrammetry, and will eventually result in reducing the currently required processing time by several orders of magnitude. As described above, orthorectification and DTM generation both require a network of accurate ground control points that are visually identifiable on imagery. In order to remove displacements associated with aircraft attitude and topographic relief, a network of tie points are used to connect the images and then the ground control points must be digitally located in order to successfully aerotriangulate and rectify imagery. Airborne kinematic GPS used in conjunction with the collection of the photographic data eliminates the

Photogrammetry, Table 2 Suggested scanner and media types for coastal research

Project scope and examples	Recommended scanner and media
Absolute position: Setbacks, hazard zones, elevation changes, short-term changes	Photogrammetric scanner, diapositives
Long-term erosion rates: Bluff or shoreline change	Graphic arts scanner, diapositives
Thematic mapping: Vegetation, land use, watershed mapping	Desktop scanner, diapositives



Photogrammetry, Fig. 3 Topographic contours generated from an edited grid DTM of Cowell Beach, Santa Cruz, California, for two time periods: (A) January 27, 1998 and (B) March 6, 1998. The seaward extension of the area under which the volume is calculated is the 1.5-m contour. The volume of sand above a zero datum on January 27 is

49,740 m³, as calculated from the topographic model. Almost one half the volume of sand on this beach was lost during the 1998 El Niño winter. By March 6, the volume of sand above a zero datum is 24,380 m³. Note how the position of the 1.5-m contour has changed between January 27 and March 6

need to collect ground control points as well as measure tie points on the imagery. Changes in aircraft position from frame to frame can be recorded using an inertial measurement system as the photography is being collected; the attitude values (ω , ϕ , κ) can be entered into processing software. Airborne kinematic GPS can be used to record simultaneously the exact location of the center of the camera lens as the instant exposure occurs, thus eliminating the requirement of locating photo-identifiable ground control points in a strip or block of images (Lucas and Lapine 1996).

For use in traditional softcopy photogrammetry, photographs are presently converted from film to digital format using a photogrammetric scanner. With continued advances in the resolution of sensor arrays in digital cameras, it is inevitable that photographic collection will eventually be dominated by digital systems.

Summary

The use of photogrammetric techniques in coastal studies has progressed in concert with technological advances over the past several decades. Early studies of coastal change used simple point measurements at control point locations on individual photographs. This was a time-consuming process and did not yield a great amount of information relative to the effort involved. In most cases, a simple rate of shoreline change at discrete locations was produced. Modern

applications of photogrammetry, although computerized and somewhat automated, are similarly time-consuming. The data produced from this process, however, enables the delineation and measurement of a variety of parameters, such as morphology, position and volume changes in coastal environments that are of use to a wide audience including coastal managers and scientists.

Cross-References

- [Beach Features](#)
- [Beach Profile](#)
- [Coastal Boundaries](#)
- [Geographic Information Systems \(GIS\)](#)
- [Global Positioning Systems \(GPS\)](#)
- [Nearshore Geomorphological Mapping](#)
- [Remote Sensing of Coastal Environments](#)
- [Shoreline and Coastal Terrain Mapping](#)

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Physical Models

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In the context of coastal science and engineering, the term physical model can be defined as the physical reproduction of the environment of interest in the laboratory, normally at reduced scale. To many readers, the use of physical models in an age of increasingly sophisticated numerical, or computer

models must appear antiquated. Indeed, Leonardo da Vinci employed the technique to examine a variety of hydraulic problems in the 1500s (American Museum of Natural History 1996)! Physical models retain their usefulness in the study of coastal processes and design for a variety of reasons, including:

1. When scaled correctly, physical models can simulate complex turbulent phenomena to a degree of accuracy that is not possible with analytical and numerical models.
2. The inherent control that laboratory experiments afford allows for the careful investigation of dominant physics, and the evaluation of cause-and-effect relationships among the various forcings of interest. These advantages are not attainable via field experiments because of the prohibitive expense of long-term coastal measurements having the requisite spatial coverage and resolution.
3. Physical models facilitate the development of visual-based, “hands on” knowledge of the environment and problem being investigated. This advantage can be particularly important when dealing with complex dynamics, and/or when examining the impact of coastal engineering design alternatives.

Physical models are however, not without disadvantages. These include:

1. Scale effects, which are present in virtually any reduced-scale physical model. These are associated with the improper scaling of one or more relevant forces or flow features. The degree of influence of scale effects depends on the relative importance of the poorly scaled variable(s) to the problem of interest. The optimization of model accuracy in the presence of scale effects remains part of the “art” of physical modeling.
2. The possibility that the natural environment cannot be fully reproduced because of (e.g.,) incomplete representation of all natural forcings, or inadequate model dimensions leading to boundary effects.
3. The high expense associated with the construction and implementation of physical models. When comparing the costs of physical and numerical models, however, one should consider the full range of costs associated with each, including, for example, the developmental costs associated with a numerical model, and the field data acquisition necessary to calibrate and validate both types of models.

Similitude

Similitude between a reduced-scale physical model and the full-scale prototype is simply defined as the condition when

the reduced-scale model is an exact reproduction of the full-scale system, or prototype, being examined. There are three basic forms of similitude, including:

- Geometric – wherein the ratios of corresponding lengths in the model and prototype are identical, for example, $(L_{\text{model}}/L_{\text{prototype}})$ $(d_{\text{model}}/d_{\text{prototype}})$. In other words, the model and prototype have the same shape.
- Kinematic – wherein the ratios of corresponding flow velocities and accelerations in the model and prototype are identical, for example, $(V1_{\text{model}}/V1_{\text{prototype}})$ $(V2_{\text{model}}/V2_{\text{prototype}})$, where $V1$ and $V2$ represent velocities at corresponding locations in the model and prototype.
- Dynamic – wherein the ratios of corresponding forces in the model and prototype are identical, for example, $(F1_{\text{model}}/F1_{\text{prototype}})$ $(F2_{\text{model}}/F2_{\text{prototype}})$, where $F1$ and $F2$ represent forces acting on corresponding locations in the model and prototype.

Assurance of geometric similarity is relatively straightforward, and requires knowledge only of the linear dimensions of the physical system to be modeled. These dimensions must then be reduced in equal proportion in all directions – vertical, lateral, and longitudinal. This assures that the reduced-scale model accurately reproduces both the scale and the shape of the prototype. In the case of reduced-scale models of coastal systems, the large spatial extent of the domain of interest and/or the restricted size of the laboratory facility, will at times require the use of a “distorted” model, in which the vertical and horizontal scales are reduced by different ratios. The use of distorted models is normally avoided because of the problems assuring kinematic and dynamic similarity in such models.

Assurance of kinematic and dynamic similarity requires knowledge, if not of the detailed physics of the full-scale system being modeled, then at a minimum knowledge of the important variables that describe the system. These variables include forces such as gravity and pressure, fluid characteristics such as density and viscosity, and flow characteristics such as velocity. One must first determine which among the various variables are important to the problem at hand. Proper representation of these variables can then assure that both kinematic and dynamic similarity are achieved. If, as is often the case in coastal systems, the number of variables is large, the proper scaling of each individual variable can become problematic. As an alternative, one can group the various variables into dimensionless parameters, which, if properly reproduced in the physical model, can assure similarity. Of course, the challenge lies in the selection of the dimensionless parameters. This process is known as *dimensional analysis*, and is accomplished either by inspection, using experience and/or knowledge of the physical system, or through a more formal process such as the Buckingham- π

theorem (e.g., see Fischer et al. 1979). For an excellent treatment of scaling considerations for a variety of practical coastal problems, the reader is referred to Hughes (1993).

The construction of physical models for coastal systems is usually accomplished through the use of one or both of the following dimensionless parameters: the Froude number: $Fr = V/(gL)^{1/2}$, where V is the velocity of interest, L is the length of interest, and g is the acceleration due to gravity; and the Reynolds number: $Re = VL/\nu$, where ν is the kinematic viscosity of the fluid. Both of these dimensionless parameters are constructed via the ratio of the force of interest – the force due to gravity in the case of the Froude number and the force due to viscosity in the case of the Reynolds number – to the inertial force. We discuss each parameter below.

Froude Number Similarity

The Froude number is employed in virtually every physical model study dealing with coastal processes, because of the importance of free surface (gravity) effects. Using our definition, $Fr = V/(gL)^{1/2}$, Froude number similarity is achieved by assuring that:

$$\frac{V_m}{(gL_m)^{1/2}} = \frac{V_p}{(gL_p)^{1/2}}$$

where the subscripts “m” and “p” denote model and prototype values, respectively. Since gravity remains constant at model scale, this relation can be rewritten as:

$$\frac{V_m}{V_p} = \frac{L_m^{1/2}}{L_p^{1/2}}$$

which indicates that the scale reduction for the velocity is equal to the square root of the length scale reduction. Alternatively, since velocity is (length/time), we can restate this by saying that the scale reduction for time is equal to the square root of the length scale reduction. In other words, using geometric similarity and Froude number similarity, and employing a model that has (undistorted) length scale reduction of 9, the corresponding timescale reduction is equal to 3. If one was to use this model to simulate a full-scale 12-h flow event, for example, one would employ a 4-h event in the model.

Reynolds Number Similarity

The Reynolds number was originally employed to predict the onset of turbulence in flow through pipes. It is commonly used in physical model studies that address problems where viscous forces are expected to play an important role. Examples of such problems in the fields of coastal science and engineering include fluid flow within porous media such as sand or a stone structure, and boundary layer flows. Using our

definition, $Re = VL/\nu$, Reynolds number similarity is achieved by assuring that:

$$\frac{V_m L_m}{\nu_m} = \frac{V_p L_p}{\nu_p}$$

This relation can be rewritten as:

$$\frac{V_m}{V_p} = \frac{L_p \nu_m}{L_m \nu_p}$$

which indicates that the scale reduction for the velocity is in this case equal to the *inverse* of the length scale reduction, multiplied by the viscosity scale reduction. If we assume that the same fluid is employed in the model as exists in the prototype, we can restate this by saying that the scale reduction for time is equal to the *square* of the length scale reduction. In other words, using geometric similarity and Reynolds number similarity, and employing a model that has (undistorted) length scale reduction of 9, the corresponding timescale reduction is equal to 81.

Clearly from this discussion, both Froude number similarity and Reynolds number similarity cannot be achieved for a given flow regime without much difficulty (e.g., using a model fluid having a different viscosity than that of fresh or salt water). It is therefore left to the coastal physical modeler to ascertain whether viscous or gravity forces dominate the system being examined. Once this decision is made, the physical model is developed using geometric similarity and either Froude or Reynolds number similarity. In most cases involving coastal systems, Froude number similarity is chosen, because of the dominance of free surface (gravity) effects. However, care must be taken to assure that viscous forces – which in such cases have been assumed to be relatively small in the prototype – are not large (relative to gravity forces) in the model. This most often requires avoiding large length scale reductions, which give rise to small Reynolds number flows at model scale, leading to potentially large viscosity effects. One example of such a dilemma is the small interstitial spaces in a model stone breakwater structure that has been reduced in scale to the point of creating very low Reynolds number flows within the spaces, and therefore unrealistic behavior of both the water motion and the stone response. An example of overcoming this problem can be found in the use of small models for slow-moving surface vessels such as sailing yachts. When such models are scaled according to geometric and Froude number similarity, the Reynolds number is too low to assure that a natural, turbulent boundary layer exists along the hull of the vessel. “Turbulence stimulators,” in the form of a roughened bow of the vessel or small roughness elements placed along the leading edge of the vessel, serve to artificially stimulate the generation of turbulence at model scale.

Challenges in Coastal Physical Modeling

Scale Effects

The inherent difficulties associated with assuring both Froude number and Reynolds number similarity represent a formidable challenge to modelers seeking to simulate the behavior of the coastal environment over large prototype scales. Typically, such models are employed in the simulation of coastal wave processes, such as wave diffraction around a proposed breakwater, or wave refraction and shoaling in the nearshore region. In these cases, as in virtually every coastal physical model, the proper representation of surface wave dynamics is critical to success, and so Froude number scaling is employed along with geometric scaling.

The largest three-dimensional physical model facilities, or wave basins, have horizontal dimensions of the order 100 m. Even so, if one desires to model a coastal system over horizontal scales of 10 km or greater (e.g., to examine the high-tide shoreline impact of a coastal structure, geometric similitude requires a length scale reduction of the order 100). This presents immediate difficulties related to scale effects, the most obvious being the requirement that the water depth must like-wise be reduced at model scale by the scale reduction factor of order 100. If, for example, the offshore depths at full scale vary over a range of 1–10 m, the model depths must range from 0.01 to 0.1 m, or 1–10 cm. These shallow water depths give rise to viscous dissipation in the model which is not present at full scale.

Although the modeler may be tempted to employ larger water depths in the physical model so as to avoid the unwanted wave attenuation at model scale, this is strongly discouraged. The use of vertical scale distortion alters the wave kinematics by altering the vertical distribution of velocity and pressure. This in turn alters the wave transformation processes, including refraction, diffraction, and shoaling. Rather than employ scale distortion, the modeler should first investigate whether the viscous effects are appreciable – particularly in terms of the problem being examined. This can be accomplished, for example, by comparing the reduced-scale wave behavior with that expected from either field data or analytical or model results. Such comparisons are in fact advisable regardless of whether scale effects are suspected. If the modeler concludes that scale effects may impact the study results, a decision must be made between proceeding – perhaps accepting that the model results are only qualitative representations of the full-scale system – or reconstructing the model at closer to full scale so as to reduce the relative importance of the model-scale viscous effects.

An additional concern regarding scale effects arises in cases when wave–structure interactions are being examined (e.g., in the model analysis of the influence of a breakwater on wave propagation). The examination of these interactions requires accurate representation of wave transmission and

reflection at the structure face. This requires accurate representation of the wave–structure interaction and wave-induced flows. Since this localized behavior is not likely to be well reproduced at model scale because of the lack of Reynolds number similarity and the importance of viscous effects in such behavior (particularly for porous structures such as stone breakwaters), it is probable that the large-scale, three-dimensional model discussed here will not be adequate to address the problem. This situation is one example of a scenario in which a two-dimensional (cross-shore and vertical) physical model can be effective. Two-dimensional physical model facilities, that is “traditional” wave tanks, allow for modeling cross-shore processes and wave–structure interaction at scales larger than those normally utilized in three-dimensional facilities. The use of closer-to-full-scale models assures that even if Froude number scaling is used in the construction of the model, the model-scale Reynolds number is sufficiently high to assure that realistic turbulent flow exists (e.g., in the gaps within a model stone structure). It is in fact not uncommon for a coastal engineering physical model study to be composed of two separate components: (1) a two-dimensional study to examine wave–structure interaction (wave run-up, transmission, reflection, etc.), and (2) a three-dimensional study of wave transformation and other processes, which uses the results of the two-dimensional study in model design and interpretation.

Moveable Bed Models

The proper simulation of sediment transport dynamics remains one of the most challenging problems facing physical modelers. And yet, the simulation of sediment transport is essential if one is to examine the coastal environment in a complete manner, since we know that bottom topography is dynamic, changing over timescales of many wave periods in response to bottom shear stresses and currents associated with surface waves. In fact, knowledge of the “equilibrium” beach geometry and bottom topography (or nearshore profile) is often a desired output of coastal physical model studies. Since the scope and cost of field observations required to achieve the same results are prohibitive, and since numerical model algorithms to address the problem must still be considered an area of active research, physical model studies remain an important tool in addressing the problem.

For an excellent discussion regarding moveable bed modeling, the reader is referred to Kamphuis (1985) and Kamphuis (2000). A fundamental problem with moveable bed models is that the model sediment grain size obtained via the same geometric scale ratio used for the hydraulic model development will in all likelihood be so small as to introduce scale effects (e.g., cohesive forces, that are not present in the prototype).

For many problems of interest, the selection of the sediment grain size scale reduction factor begins with the

assessment of whether bed-load or suspended load sediment transport dominates the coastal system being examined. For example, we might expect that suspended sediment transport dominates the environment in the breaking wave region, or surf zone while bedload transport dominates moveable bed dynamics in deeper, less energetic regions of the coast.

In the case of bedload-dominated systems, a reasonable approach appears to be to select the length scale reduction factor based on the ratio of the prototype and model sand grain diameter, with the model sand grain diameter being chosen as to avoid cohesive forces, or roughly speaking, greater than 0.06mm. Additionally, the model sand must have the same density as that of the prototype sand. In practice, this requires either a large model facility in order to accommodate scale reduction factors of order 10, or the good fortune of dealing with prototype sand having grain diameters measuring several millimeters.

In the case of suspended load-dominated systems, one must correctly scale the sandfall velocity, as this becomes a critically important parameter. Dean (1973) proposed the use of a dimensionless parameter that includes a measure of the erosion capability of the wave and the bias toward deposition represented by the fall velocity of the sand. This parameter, now commonly referred to as the Dean number, can be written:

$$\frac{H}{wT}$$

where H is the wave height, T is the wave period, and w is the sediment fall velocity.

The fall velocity of the model sediment can be selected by forcing similarity of the Dean number between model and prototype, using geometric similarity for the wave height and Froude number similarity for the wave period. Since sediment fall velocity is a strong function of grain diameter, this provides the scale reduction factor for the model sediment grain size, which is expected to differ from the length scale reduction factor.

Concluding Remarks

Numerical models continue to become more sophisticated in their handling of the complex, turbulent coastal environment. Improved instrumentation allows for the collection of high-resolution data that are being used to both improve our understanding of coastal dynamics, and to provide essential data for numerical model calibration and verification. However, our knowledge of the complex interactions of waves, currents, and (moveable) bottom bathymetry across scales ranging from wave period to several years, has not progressed to the point where accurate simulations of coastal processes can be

made solely with the use of numerical models. It would appear therefore, that physical models will remain an important tool for coastal scientists and engineers for the foreseeable future. It is likely in fact that for many problems (e.g., the influence of structures on the coastal environment), the future holds promise of stronger-than-ever collaboration between the physical and numerical modeling community, with physical models being used to provide valuable insights and verification data to the developers of numerical model algorithms.

Cross-References

- [Beach Processes](#)
- [Erosion: Historical Analysis and Forecasting](#)
- [Geohydraulic Research Centers](#)
- [Shore Protection Structures](#)
- [Wave Climate](#)

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Placer Deposits

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A placer is any waterborne deposit of sand or gravel that contains concentrated grains of valuable minerals such as gold or magnetite, grains that had originally been eroded from bedrock but were then transported and concentrated by the flowing water. Placers can be found in rivers (alluvial placers) and on the coast, particularly in beaches (beach placers). In some locations, the ore minerals that initially were concentrated into a beach placer have been blown inland

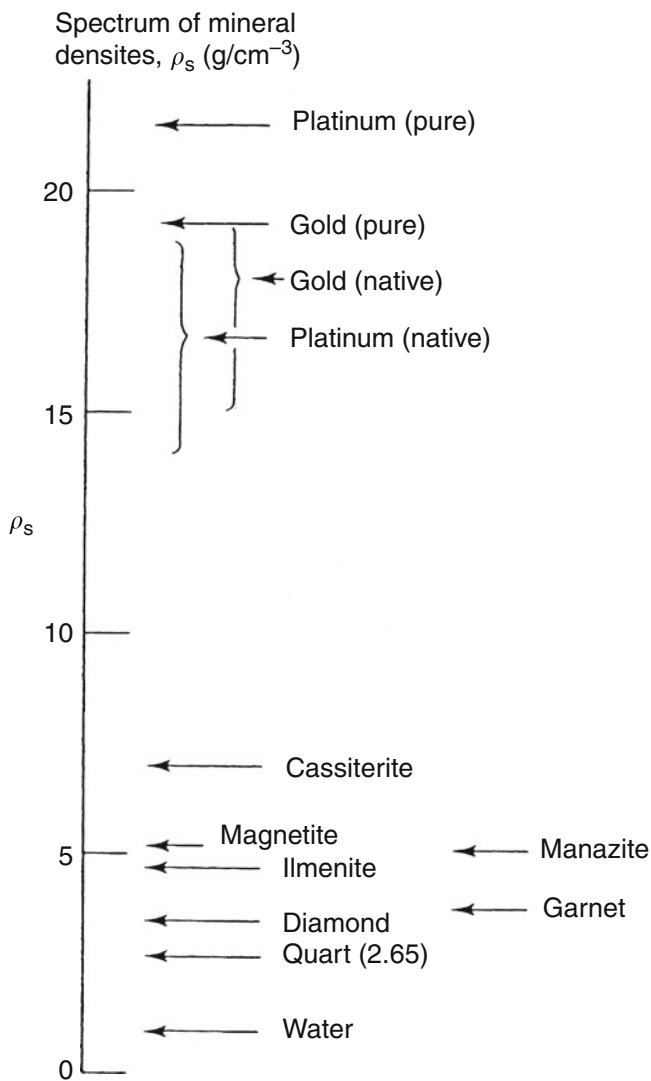
by coastal winds to form mineral-rich dune sands. Beach placers also have been submerged by the rising level of the sea, and are now found as relict placers within sediments on the continental shelf. However, some placer deposits found in continental shelf sands may have originated in that environment, formed by waves and currents that transported the shelf sands and concentrated the ore minerals.

Coastal placers, including those in beaches, dunes, and on the continental shelf, have been mined for gold, platinum, diamonds, the titanium-bearing minerals rutile, monazite and ilmenite, cassiterite (containing tin), magnetite (iron), zircon (zirconium), and garnet (used for making sand paper). Coastal placers are found throughout the world, with the economically most significant deposits being in Alaska (gold), India (magnetite), New Zealand (magnetite), Australia (rutile, monazite, and zircon), Indonesia (cassiterite), and South Africa (diamonds).

Placers are formed under the action of flowing water, by a current or the to-and-fro movement of water beneath waves, which transport the sand and gravel, leading to the concentration of the placer minerals due to their high densities. The range of mineral densities is shown in Fig. 1, illustrating that platinum and gold have extremely high densities, while cassiterite, magnetite, etc. (generally referred to as heavy minerals) range in densities from about 3.5 to 7.0 g/cm³. Of importance, these densities are greater than those of quartz and feldspar (the light minerals), which have a density on the order of 2.65 g/cm³. Included in Fig. 1 is the density of water, 1.00 g/cm³ for fresh water and slightly greater for sea water.

Most deposits of sand on the continents are composed mainly of quartz and feldspar, with only small percentages of heavy minerals. When transported by currents and waves, the processes of selective grain sorting can occur where the quartz and feldspar are transported more easily and at higher rates, leaving behind a concentrated deposit of heavy minerals (Komar 1989). In that the heavy minerals are generally dark in color, the result is the formation of a “black sand” deposit; if it contains a valuable ore mineral, the deposit then also qualifies as a placer.

The process of formation of black sand and placer deposits can be readily observed on beaches, especially during a storm that erodes back the beachface and berm, transporting the eroded sand to offshore bars. The tan-colored grains of quartz and feldspar are easily picked up by the swash of the waves at the shore, and are quickly transported offshore, leaving behind the heavy minerals as a concentrated black-sand layer on the eroded beach. In that the different heavy minerals have a range of densities and colors (e.g., pink garnet versus black magnetite), some selective grain sorting between the heavy minerals may also be seen (Komar 1989). With the return of smaller waves following the storm, the quartz and feldspar sand moves back onshore, burying the black-sand layer. The search for beach placers, therefore, often



Placer Deposits, Fig. 1 The range of densities of quartz grains and representative minerals found in placers. (From Komar 1989. With permission of CRC Press)

necessitates digging trenches through the beach sand in order to reach the black-sand deposit at depth.

Having initially formed within the active swash zone of the beach, the placer minerals may be washed inland by still stronger storms, or blown there by the wind to form dunes. If the area of coast is accreting due to an abundant supply of sand, with the beach environment slowly shifting seaward, beach placer deposits dating hundreds to thousands of years old may be found well inland. This is the case on the southeast coast of Australia, where research has focused on the formation of such deposits (Roy 1999), which are also the deposits most actively mined at present, the mineral resource on the modern beach having been exhausted by mining in the past. The search for placer deposits also has extended out onto the continental shelf off the southeast coast of Australia, but generally with disappointing results. Research suggests that

ocean waves may preferentially transport the heavy minerals and placer minerals onshore from the shelf sands, contributing to their concentration in the beaches, but depleting the shelf sands (Cronan 2000). The principal minerals that are being mined from continental shelf placers are gold (Alaska), cassiterite (Indonesia), and diamonds (South Africa).

Cross-References

- [Beach Features](#)
- [Beach Sediment Characteristics](#)
- [Beach Stratigraphy](#)
- [Cross-Shore Sediment Transport](#)
- [Eolian Processes](#)
- [Mining of Coastal Materials](#)
- [Shoreface](#)
- [Surf Zone Processes](#)

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Pleistocene Epoch

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Definition

The Pleistocene Epoch is the first long period of the Quaternary, also named the “Ice Ages.”

It started at the end of the Tertiary less than three Mio years ago with a remarkable cooling of the Earth’s climate, possibly triggered by the closing of the Panama land bridge (O’Dea et al. 2016). The beginning of the Pleistocene is set at 2,588,000 years ago from a stratotype section in Sicily (“Gelasian Stage,” Gibbard et al. 2010; Head and Gibbard 2015). This is close to the Gauss-Matuyama geomagnetic reversal and lasted till 1,800,000 years ago, when the first “cold guests” like the mollusk *Cyprina islandica* or the foraminifer *Hyaline balthica* appeared in the Mediterranean and extended glaciations occurred on the continents (see also

Table 1). The Gelasian was followed by the Calabrian (1,800,000–780,000 years ago), its end again close to a paleomagnetic boundary (Brunhes-Matuyama, Shackleton and Opdyke 1976). The Calabrian was followed by the Ionian Stage (Middle Pleistocene), comprising the main glaciations (Table 1 and Ehlers 1996). The middle Pleistocene ended 126,000 years ago at the warm peak of the last interglacial

phase (Eemian, Sangamon), when the oceans reached a worldwide level at +6 m.

In regions of long uplift, interglacial shorelines are preserved, regionally many 100 m high in steps as along the Huon Peninsula of Papua New-Guinea or Barbados. If coral reefs are uplifted, radiometric ages can be taken from fossil corals (Schellmann and Radtke 2004).

Pleistocene Epoch, Table 1 The Pleistocene Epoch

Years BP	Chrono-stratigraphy		Magnetic polarity	Glacial epochs	Sea level
0	QUATERNARY	HOLOCENE	NORMAL	Postglacial	high
10,000		LATE PLEISTOCENE		WISCONSINAN GLACIATION	low
100,000				Sangamon Interglacial	high
		MIDDLE PLEISTOCENE	BURUNDI	ILLINOIAN GLACIATION	low
				Yarmouth Interglacial	high
				KANSAN GLACIATION	low
500,000				Altonian Interglacial	high
0.78 Ma		EARLY PLEISTOCENE	MATUYA	NEBRASKAN GLACIATION	low
1.0 Ma				Pre-Nebraskan Epochs	various sea level oscillations
1.5 Ma					
2.0 Ma					
2.5 Ma					
	TERTIARY		GAUSS		

Multi-temporal changes between cold and warm periods are analyzed by oxygen isotope measurements (ratio of $^{16}\text{O}/^{18}\text{O}$ in carbonate shells of microfossils (foraminifera) in bottom layers of the tropical Pacific (Murray-Wallace and Woodroffe 2014; Shackleton and Opdyke 1976). Drop of temperature in high to middle latitudes during glacial peaks was around 12–15 °C, whereas in tropical latitudes, it was much lower (about 4 °C), and belts of coral reefs and tropical rain forests survived. Fauna was characterized by large mammals like mammoth and mastodon, long horned bison, cave bear, or sable-toothed tiger as well as camels and horses, most of them extinct at the end of the Pleistocene Epoch, probably accelerated by hunting man. The many and intensive climatic changes during the Pleistocene Epoch certainly have had impacts on the evolution of plants and animals including man, and the Pleistocene is the period of human development from early hominids to *Homo sapiens sapiens* and its distribution over all continents.

The last phase of the Pleistocene (Late Pleistocene) lasted more than 100,000 years from the warmest peak of the last interglacial at 126,000 years ago (Marine Isotope Stadium = MIS 5e), covering all phases of the last glaciation (Würm, Wisconsin) including late-glacial oscillations to its end 11,700 years ago, when the second epoch of the Quaternary, which is the Holocene, began. Dating of all the variations during this phase was supported by counting annual ice layers from Greenland ice cores. The Last Glacial Maximum (LGM) was around 23,000–18,000 years ago, when sea level was as low as –120 m. By this, land bridges occurred between continents and islands, and wide areas of shelves have been exposed, enabling mankind to spread and bridge distances which now are covered by the ocean again.

At the end of the Pleistocene, strong and abrupt oscillations in climate, ice extend, and sea level occurred, also identified by Greenland ice core data. The LGM ended at about 14,700 years ago (Oldest Dryas period), followed by the first warm interstadial (Bölling, 14,700–14,200 years ago, with a short Older Dryas interval to its end between about 14,200–13,900 years ago). The warm interstadial Alleröd (13,900–13,000) reached temperatures equivalent to modern conditions in its peak but changed into a longer lasting cold period of the Younger Dryas stadial (13,000–11,700 years ago). The end of the Younger Dryas at 11,700 years ago also was the abrupt termination of the Pleistocene with a rapid temperature rise, glacial recession, and sea level transgression. Temperature variations during the Holocene had maximum ranges of ± 2 °C into modern times.

Conclusion

Although by far most of the coastal forms we see today belong to a Younger Holocene period, the Pleistocene Epoch with its remnants of older shorelines and geo-archives pointing to

former environmental conditions along the coastlines of the world has delivered a large amount of data and hints as analogues. Those from the warm last interglacial with a higher sea level as today also is used as a scenario for future conditions under climate warming with accelerated sea-level rise.

Cross-References

- [Geochronology](#)
- [Holocene Epoch](#)
- [Last Interglacial](#)
- [Late Quaternary Marine Transgression](#)
- [Marine Terraces](#)

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Pluvial Lake Shore Deposits

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Pluvial lakes are lakes that existed in enclosed basins in the interior of continents during times of enhanced rainfall, reduced evaporation, or a combination of both. These lakes in many cases left a valuable record of paleoclimates in the form of lake sediments, of erosional shore features,

and of gravelly to sandy shore deposits. Pluvial lake shore features have been recognized in many parts of the world, but are particularly well known from North America. Here, they were first described, identified, and interpreted in classical studies by geological pioneers I.C. Russel (1885), and G.K. Gilbert (1890) who worked on the shore deposits of Lake Bonneville (“the proto Great Salt Lake”), a pluvial lake that existed during the Wisconsinan/Weichselian Ice Age.

In the last few years, Lake Bonneville shore deposits have received renewed attention. Many of the internal structures of these deposits remained unknown because of slumping or other types of outcrop deterioration. Only where gravel pits, road building, or fluvial dissection afforded access, could internal sedimentary structures be defined and interpreted (Oviatt and Miller 1997). Now, however, the advent of Ground Penetrating Radar (GPR) permits identification of previously unknown sedimentary structures. For instance, Smith et al. (1997) described gravelly shore deposits from Lake Bonneville, and identified a sequence of sedimentary processes that led to the deposition of gravelly shore features.

Other gravelly shores exist further west, in California and Nevada. One of the pluvial lakes that existed here was Lake Lahontan (Fig. 1) which left a variety of shore features. This is illustrated for the Jessup embayment, Nevada, by Adams and Wesnousky (1998, and references therein). The reader is referred to this excellent description of shore features and shore processes. The authors were successful in separating shore features formed during both transgressive and regressive as well as highstand phases of lake-level changes. Examples of other pluvial lakes that existed are Lake Russel (the “proto-Mono Lake”) and Lake Manly (Fig. 2) that filled the central part of Death Valley.

Although these pluvial lakes left a legacy of both erosional and depositional features, this entry is primarily concerned with depositional features, consisting mainly of beach gravel. After a general description of beach gravel, we will illustrate these deposits, based on a study of a deposit from Lake Manly by Ibbeken and Warnke (2000).

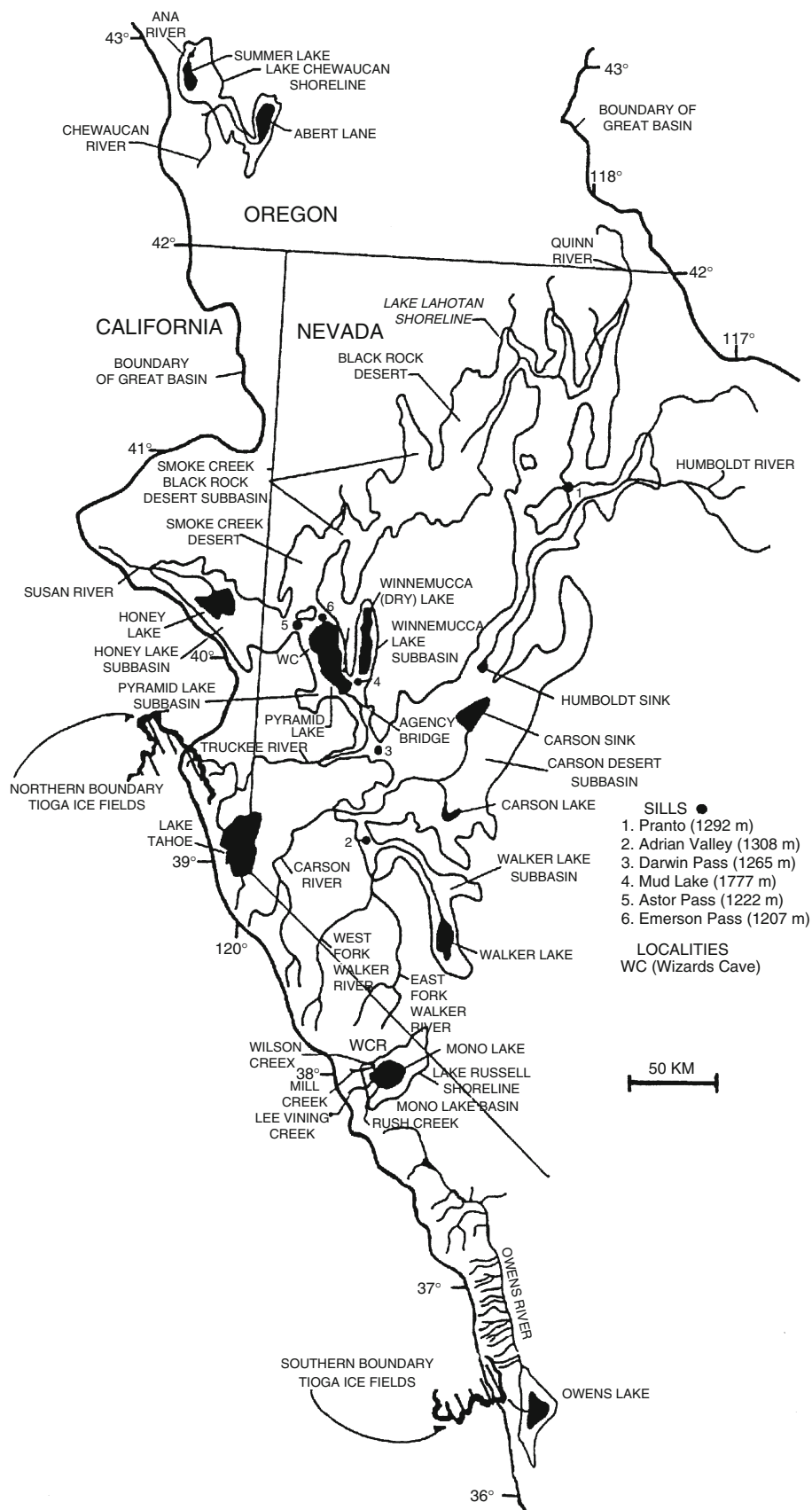
Beach gravel consists of clasts which “tend to be well rounded, well stratified, and well sorted within strata” (Adams and Wesnousky 1998, p. 1319). Clast sizes range from pebbles to cobbles or larger. In most cases beach-gravel deposits are clast supported, but may have a matrix of coarse sand to granules (Adams and Wesnousky 1998). Depositional features are various types of barriers, spits, etc., similar to equivalent features on marine coastlines, and produced by washover processes during storm events,

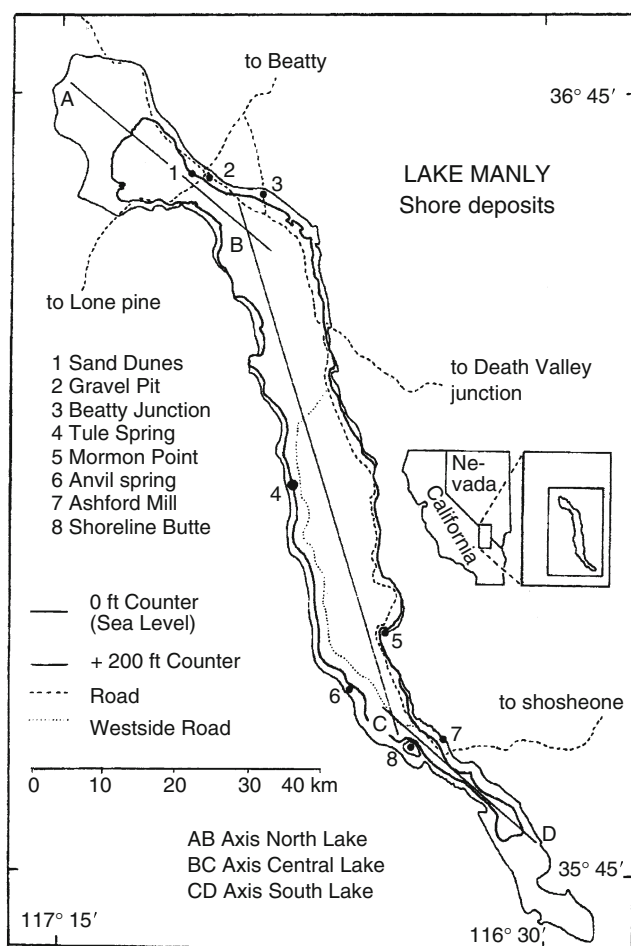
longshore transport, and simple aggradation on the beach-face. However, the rapid lake-level changes can produce lakeshore deposits with a history and internal architecture “notably more complex than their marine counterparts” (Blair 1999, p. 217). Blair (1999) studied the Churchill Butte shore deposit in western Nevada, about 52 km east-southeast of Reno. The age of the deposit is about 14.5 ka. The shore remnant is easily accessible because it is bisected by US Hwy Alt. 95, leading from Fallon, Nevada to Carson City, Nevada. The locality affords an excellent introduction to gravelly lakeshore deposits, based upon the analysis of the deposit provided by Blair (1999). The following description is extracted from Blair (1999) to which the reader is deferred for more detailed information. “The beach deposit consists of a lakeshore barrier spit and a lake lower-shoreface spit platform. The lake-ward side of the barrier consist of beachface deposits (1–10-cm-thick beds of granules and pebbles) sloping 10–15°. These interfinger downs-lope with thicker and less steep lakeward-dipping beds of pebble gravel of the lake upper shoreface. Interstratified with these beds are high-angle cross-beds that dip southward, along-shore. Landward-dipping (15–20°) sets form the proximal backshore of the barrier, deposited by overwash processes during storms. Fossiliferous sand and mud exist landward of the barrier in what was a lagoon, separated from the lake by the barrier. The lake lower shoreface consists of a southward prograding spit platform constructed by longshore drift, and is characterized by ‘Gilbert’ foresets of pebble gravel dipping southward 16°.” The various facies identified by Blair (1999) are described in Table 1.

The Hanaupah-Fan Shore Deposit at Tule Spring, Death Valley National Park, California

During the Pleistocene, a series of lakes existed in the enclosed basin of Death Valley, collectively referred to as Lake Manly. Higher lake levels existed during pre-Wisconsinan/Weichselian ice ages. For instance, the higher Beatty Junction bar complex (Fig. 3) was deposited during the Illinoian (?) according to Orme and Orme (1991) although Klinger (2001) favors a younger age, based on soil development on the main spit. The reason(s) for higher lake levels in pre-Wisconsinan times are still in debate. A comprehensive description of shoreline elevations is provided by Meek (1997). During the Wisconsinan, the lake level oscillated about the 0 m contour line in the valley, and left a series of shore features that have been and continue to be, the subject of investigations. (Note added in proof: Machette et al. 2003, how consider the deposit described

Pluvial Lake Shore Deposits,
Fig. 1 Map of Lake Lahontan,
 from Benson (1999). (Reproduced
 with permission of American
 Geophysical Union)





Pluvial Lake Shore Deposits, Fig. 2 Sketch map of several, major known localities of shore features in Death Valley National Park. Zero feet (0 m) and 200 ft. (61 m) contour lines indicated. (From Ibbeken and Warnke 2000. Reprinted with permission from Kluwer Academic Publishers)

below, to be 128–145 ka, correlative with oxygen-isotope stage 6).

Here, we describe a remarkable, gravelly shore deposit that exists west of Tule Spring (Fig. 2). It can be seen from the Westside Road in Death Valley National Park, and can be reached by a short hike across the lower fan surface. The deposit consists of a gently sloping, WSW-ENE elongated ridge, about 600 m long, 165 m wide, and 8 m high (Fig. 4). Its surface extends from –12 to +28 m in elevation. The sedimentary inventory consists of cross-stratified gravel beds (Fig. 5) of various size ranges dipping in all directions (beachface and overwash sets), of horizontal berm gravel beds, and horizontal silt layers, probably lagoonal overwash deposits landward of a barrier. The deposit is the erosional remnant of a once much larger deposit that extended from the lakeshore east into the lake. Waves from both the north

Pluvial Lake Shore Deposits, Table 1 A list of sedimentary facies identified by Blair (1999) from a beach deposit of Lake Lahontan. Note the attitude of the beds with respect to the paleoshore. (Modified after Blair 1999)

Facies	Sedimentary features	Depositional environments
A	Unsorted, unstratified, angular, muddy bouldery pebble cobble gravel	Subaerial bedrock-fringing colluvium
B	Thickly bedded, matrix-supported, muddy cobble pebble gravel	Subaerial alluvial-fan debris flows
C	Lakeward-dipping low-angle (5–15°) beds of granular fine to medium pebble gravel	Lake beachface and upper shoreface
D	Landward-dipping (10–20°) foresets of sandy granule fine to medium pebble gravel	Proximal washover in a back-barrier pond
E	Irregular pebbly granule gravel, fossiliferous sand, mud, and diatom beds	Distal back-barrier pond
F	Poorly sorted, sandy granule pebble gravel in south-dipping (10–16°) foresets	Spit-front foresets, lower lake shoreface
G	Burrowed and oscillation ripples, granular sand in south-dipping (1–6°) toesets	Spit-front toesets, lower lake shoreface
H	Horizontally laminated silt and thickly bedded mud and clay	Lake bottom, below storm wave base

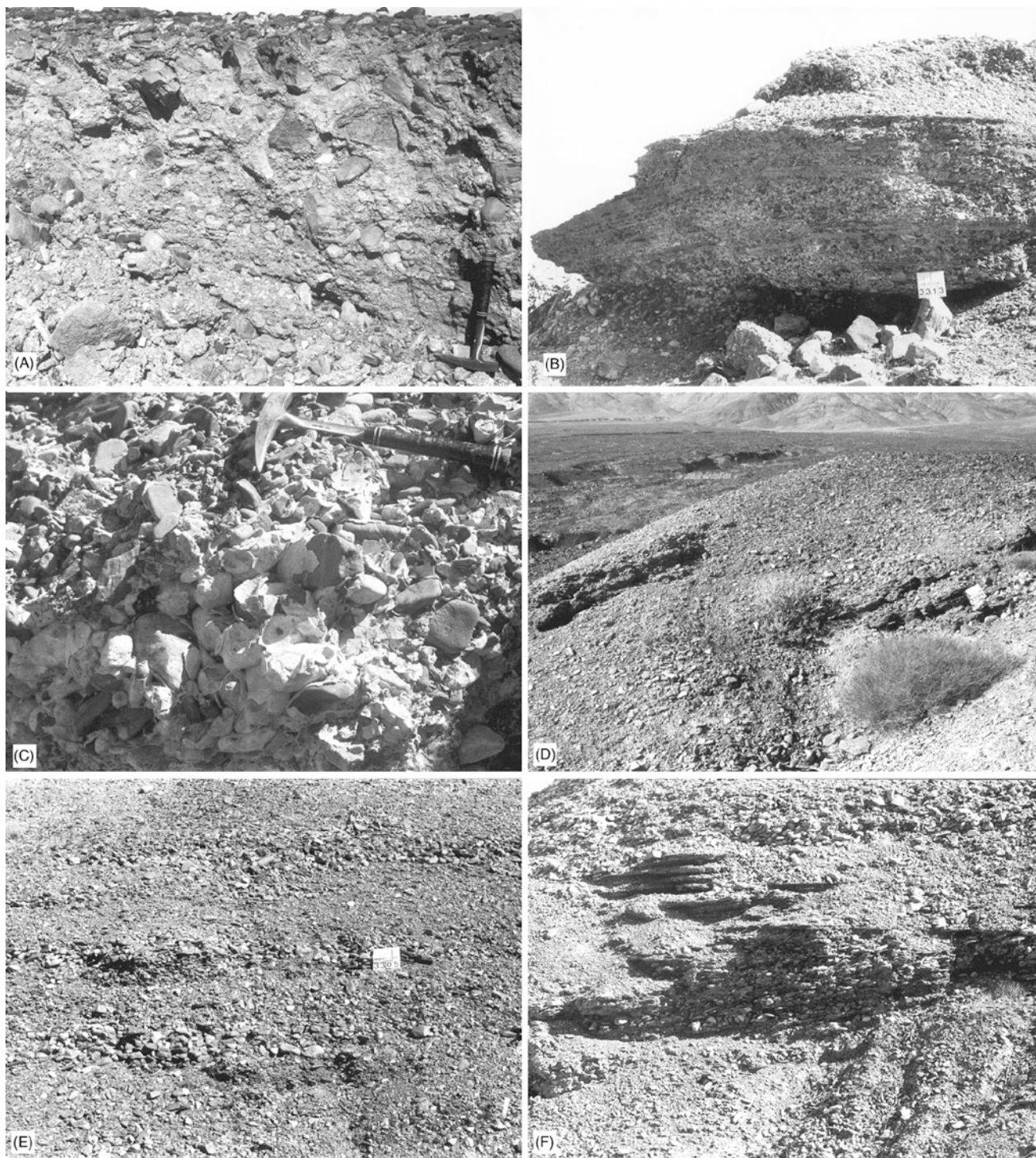
and the south eroded fan materials, and produced a sediment body with a complex architecture. Although at first we favored the idea that the deposit was formed during lake-level rise, albeit with major oscillations (Ibbeken and Warnke 2000), subsequent ground-penetrating radar GPR profiles (unpublished data) show that the gravel beds have a predominant offlap architecture. As a working hypothesis we suggest that the deposit started to form during a highstand of Lake Manly, but at an unknown elevation since the uppermost part has been lost by erosion. Falling lake level caused the deposit to be extended eastward, and produced the sloping surface that exists today, albeit modified by faulting. In a final phase, a discordant gravel layer was deposited over the entire surface of the deposit. This uniform gravel layer is quite distinct from the surrounding fan surfaces. It is relatively fine grained, better sorted, and densely packed. Rock varnish is very well developed (Fig. 6), resulting in a dark surface color that makes the deposit recognizable on aerial photographs. The exact depositional history and facies architecture still remains to be worked out, but the deposit has the potential of yielding valuable information on wave approaches and wave strength during the existence of Late Wisconsinan Lake Manly.

Pluvial Lake Shore Deposits,
Fig. 3 Beatty Junction bar
complex. This is the main spit
(Spit B) of Klinger (2001)



Pluvial Lake Shore Deposits,
Fig. 4 The Hanaupah-Fan shore
deposit at Tule Spring. The view is
from the Hanaupah-Fan toward
the east. Geologists for scale





Pluvial Lake Shore Deposits, Fig. 5 Sedimentary features of the Hanaupah-Fan shore deposit. The number plate is $25 \times 8 \times 2.5$ cm for scale. Modified from Ibbeken and Warnke (2000). (a) Debris flow near the base of the deposit. Southeastern limit of deposit. (b) NE dipping beachface sets, capped by berm crest sets. Northeastern limit of deposit.

(c) Shore gravel encrusted by tufa. Northeastern limit of deposit. (d) SE dipping beachface deposits. Southern slope of deposit. (e) Beachface (?) deposits. Southwestern limit of deposit. (f) North-dipping beachface deposits, capped by berm crest sets. Northwestern limit of deposit

Pluvial Lake Shore Deposits,
Fig. 6 Smooth, varnished surface
 of deposit



Cross-References

- Bars
- Beach Features
- Gravel Barriers
- Paleocoastlines

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Polders

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Poldering is practiced for reclaiming arable land in many countries of the world in lacustrine, riverine, and coastal lowlands in areas with impeded drainage.

The art of poldering will be exemplified for the Flemish-Dutch-Northern German lowlands around the eastern shores of the North Sea. Poldering essentially consists of isolating a

certain area by diking and improving the drainage of this area by expelling the surplus water. So poldering requires three major abilities: the art of diking, of drainage and of the discharge of surplus water.

The Art of Poldering in Medieval Times

The earliest diking and reclamation of peat areas was reported from around 800 AD. Monasteries and cloisters, being centers of knowledge, played an important role in the development of techniques and abilities for diking, poldering, and river training. The dikes were preferentially located on the natural levees of the creeks and rivers. Land reclamation required detailed local knowledge of the topography and hydrography of the area.

In early times dewatering was accomplished by digging of ditches according to the (small) slopes of the terrain. A major technical invention was the so-called “ebb-valve” which opened when the outside water level was lower and closed by the water pressure while the outer level was higher. It was only since 1200 AD that larger channels were dug to connect the major drainage creeks and small channels in order to facilitate the discharge of surplus water. But it was still a drainage under

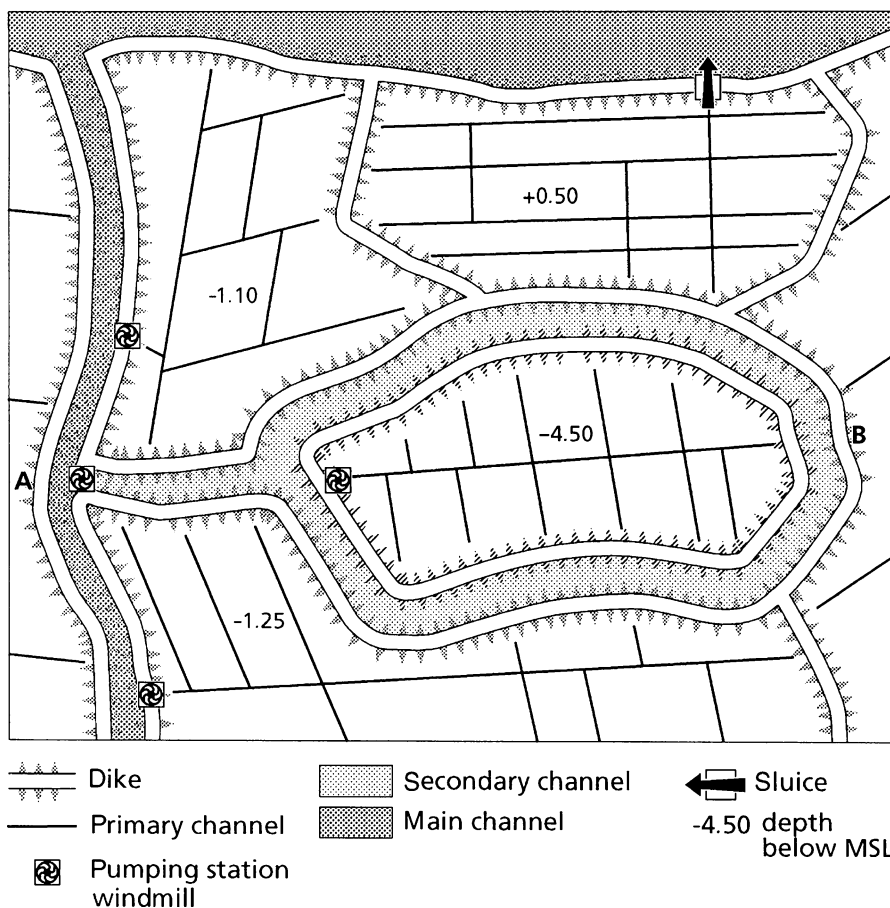
natural gradients, thus on the supratidal areas. Similar practices were employed in the riverine areas by digging of ditches, almost parallel to the main river courses, thus using the natural downslope gradient of the terrain.

It was in the fourteenth century that the sluice was developed, which made it possible to have different levels in and outside the polder. The sluice allows ships to pass the obstruction.

Poldering and land reclamation not only had beneficial effects by increasing the extension of arable lands. On the negative side, poldering decreases the area of the supratidal and higher intertidal lands. This reduced the tidal storage capacity. The drainage was more concentrated in the channels, which eroded. As a result the tidal wave could penetrate further inland and because of the funnel shape of estuaries and tidal basins the tidal range increased too. Not only the HW levels were heightened but also the storm surge levels. This is especially problematic in coastal lowlands experiencing a sea-level rise.

Dewatering resulted in bio-oxidation and compaction of the peat and clay layers, which obstructed the drainage. In addition, the lowering of the land surface by compaction was increased by the excavation of peat for domestic use and for salt extraction by an ever-increasing population. Many

Polders, Fig. 1 Drainage system in larger polders. (From: A compact Geography of the Netherlands. With Permission of the Royal Dutch Geographical Society)



disastrous floods have been reported, and considerable acreage of land was lost.

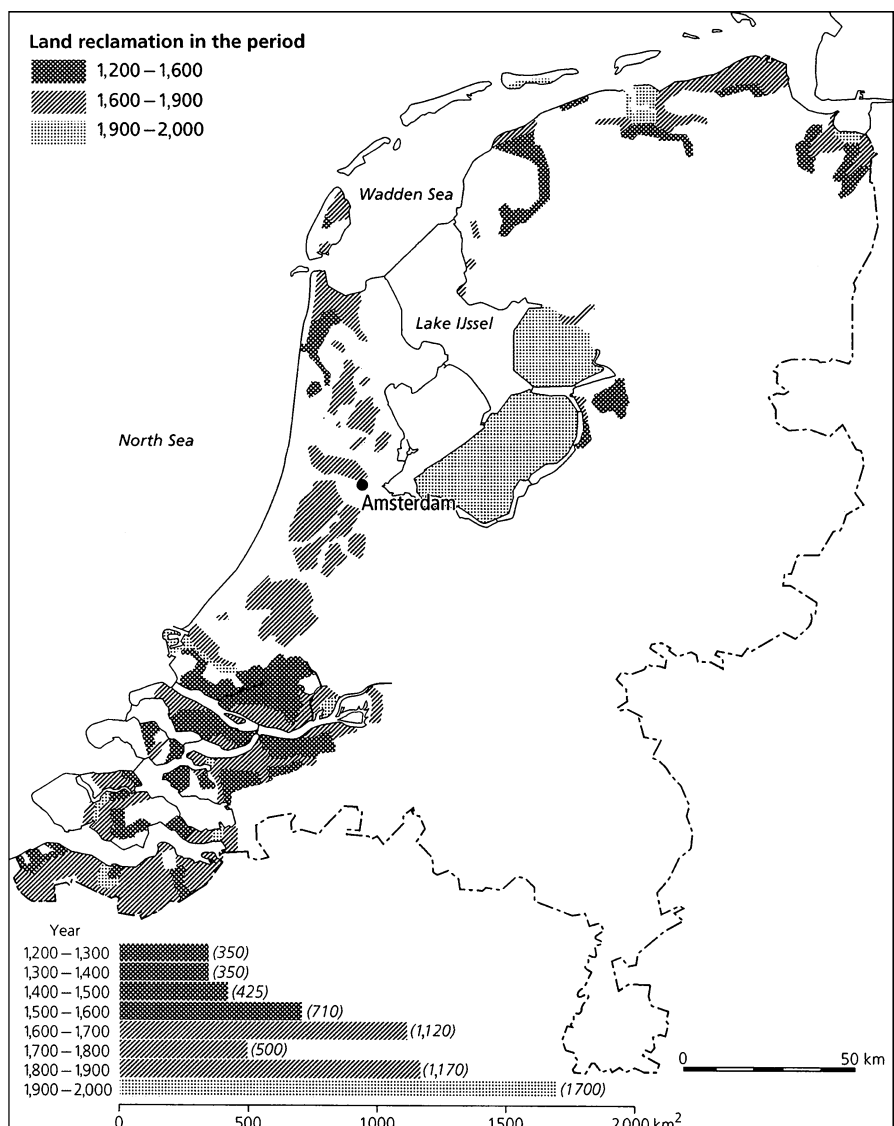
The Art of Poldering in the Fifteenth to Eighteenth Century

The problem of the drainage of the ever-sinking polders was solved effectively in the beginning of the fifteenth century by the introduction of the windmill as a pumping device. Much more water from lower levels could be pumped up and discharged and greater areas (combined polders) could be handled. From then on, the way of poldering showed a new setup (Fig. 1) consisting of diked areas with ditches draining into internal drainage channels, which in turn drain into natural waters. Also in the riverine areas, the development

of the windmills offered tremendous opportunities for improvement of the drainage. A large number of windmills were built here too. These developments resulted in new, larger reclamations, a well-balanced and organized hydrological drainage system and better water management.

However, windmills were not applied everywhere in the coastal and riverine areas. Windmills were costly (equivalent of US \$500,000 at present) and have a low capacity. In addition, they could only function effectively at a wind speed above some 4 Beaufort, effectively only some 25% of the time (Van der Ven 1993). Other areas were too low to be handled by windmills or required too high a discharge capacity. This was especially the case in Flanders and the northern part of the Netherlands and Germany. These areas could only be poldered after better techniques became available.

Polders, Fig. 2 Land reclamation in the Netherlands. (From: A compact Geography of the Netherlands. With Permission of the Royal Dutch Geographical Society)



Poldering in the Nineteenth and Twentieth Century

Steam-powered pumping stations took over the function of the numerous windmills, because they were more powerful and cheaper. Again because of their greater capacity they made the reclamation of larger lakes possible (Fig. 2). In addition, this resulted in an increase in scale of the polder units and a rationalization of the system of drainage channels. In the twentieth century motor and electric power increased the capacity of the drainage devices and the size of the polders further. After completion in 1932 of the Closure Diike, which separated the former Zuiderzee from the Wadden Sea (Fig. 2) three new modern polders were constructed in Lake IJssel between 1937 and 1968, totally covering some 145,000 ha.

Recent Developments

In the last decade a discussion was started about depoldering some polders by opening dikes or substantially reduce the drainage. Two arguments are at stake. The first is the expected consequences of an increased sea-level rise viz. a heightening of the HW levels along the coast and the increase in precipitation in the drainage areas of the major rivers, which will negatively affect the safety of the dikes and drainage of the polders areas.

The second argument is that because of the developments of agriculture in the European Community (EC) the need for arable land has reduced considerably and more and more land becomes available for other purposes, including housing, recreation, and landscape restoration and nature conservation. So there is a shift in priorities in rural planning. Several plans are under consideration viz. the use, as in former times, of overflow dikes, to temporarily store water in the polders during peak river floods to reduce the top water levels. Another plan is to reduce the drainage of some higher polders, creating lakes in which housing, floating, or on heightened lands, recreation and nature conservation are foreseen. A third type of plan is to depolder areas in the upper parts of estuaries in order to increase the tidal volume and the tidal currents to reduce the necessary dredging in the shipping routes.

Conclusion

Poldering requires special techniques and abilities: precision leveling and positioning, adequate information of water levels, current velocities, drainage systems and facilities, diking, discharge techniques, and water management policies and organization. In past time these challenges forced the ever-increasing population to solve these problems. In recent

times, facing problems of climatic change, there is a need for more space for water management and depoldering of some areas is a serious option.

Cross-References

- [Alluvial-Plain Coasts](#)
- [Changing Sea Levels](#)
- [Coastal Subsidence](#)
- [Coastal Management Practices](#)
- [Dikes](#)
- [Human Impact on Coasts](#)
- [Reclamation](#)
- [Salt Marsh](#)
- [Tidal Environments](#)
- [Wetlands](#)
- [Wetland Restoration](#)

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Pressure Gradient Force

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Definition

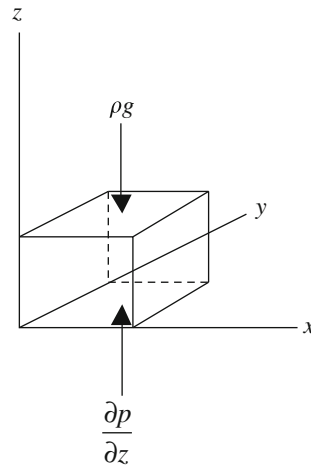
Pressure differences, not pressure per se, move fluids. A pressure gradient force (often abbreviated pgf) is a force per unit mass (i.e., an acceleration), expressed as the ratio of a pressure difference divided by the distance over which the difference occurs.

Introduction

Horizontal and vertical forces called pressure (p) gradient forces drive fluids. In vector equations, this is symbolically represented by $\frac{1}{\rho} \vec{\nabla} p$. The Greek letter “rho” (ρ) is density, and $\vec{\nabla} p$ is vector notation for “grad p” and is shorthand for $\frac{\partial p}{\partial x} \vec{i} + \frac{\partial p}{\partial y} \vec{j} + \frac{\partial p}{\partial z} \vec{k}$, where $\vec{i}$, $\vec{j}$, and $\vec{k}$ are unit vectors in the x, y, and z directions, respectively. A pressure gradient

Pressure Gradient Force,

Fig. 1 A cubic meter of seawater in hydrostatic balance where the upward pressure gradient force per unit volume $\frac{\partial p}{\partial z}$ is equal to the downward gravity force per unit volume ρg



force is a spatial expression of Newton's second law, $F = ma$, as applied to the ocean or the atmosphere.

Figure 1 shows a Cartesian coordinate system with the x-axis pointing eastward, the y-axis pointing northward, and the z-axis pointing upward in the opposite direction of the gravity vector $\vec{g}$. The x-y plane is parallel to a level surface and may be defined as coincident with the equipotential surface, known as the marine geoid, situated approximately a meter or so below mean sea level. An equipotential surface is a surface of equal potential energy per unit mass, everywhere orthogonal to the local gravity vector where the geopotential $\Phi \equiv gz$ is a constant.

The largest oceanic pressure difference, $\partial p = p_{z_2} - p_{z_1}$, occurs in the vertical over a distance $\partial z = z_2 - z_1$. The ratio $\frac{\partial p}{\partial z}$ is defined as the vertical pressure gradient and is given by the hydrostatic equation $\partial p = -\rho g \partial z$, where ρ is the density of seawater ($\rho_{\text{seawater}} \approx 1025 \text{ kg} \cdot \text{m}^{-3}$). If written in the form $\frac{1}{\rho} \frac{\partial p}{\partial z} = -g$, the hydrostatic equation defines a balance between the vertical pressure gradient force per unit mass and the acceleration due to gravity. In m.k.s. units $\frac{1}{\rho} \frac{\partial p}{\partial z}$ is dimensionally $\left[\frac{\text{m}^3}{\text{kg}} \right] \left[\frac{\text{N} \cdot \text{m}^{-2}}{\text{m}} \right] = \frac{\text{N}}{\text{kg}}$, which is a force (newtons) per unit mass (kilograms) when pressure difference is expressed in pascals (Pa) or newtons per square meter ($\text{N} \cdot \text{m}^{-2}$) and vertical distance is expressed in meters (m).

Discussion

In terms of ocean currents, the most important pressure gradient force per unit mass is in the horizontal $\frac{1}{\rho} \frac{\partial p}{\partial x}$ and $\frac{1}{\rho} \frac{\partial p}{\partial y}$ in the east-west direction and the north-south direction, respectively. Recalling that the partial derivative ∂p is taken with respect to a level surface ∂x or ∂y , the horizontal pressure gradient force per unit mass may be expressed as height departures of a constant pressure surface. From the hydrostatic equation p_2

$= - \int_{z_2}^0 \rho g \partial z = \rho g h_2$, and similarly for p_1 , the pressure difference on a level surface is just $\partial p = p_2 - p_1$. Thus the east-west pressure gradient force per unit mass $\frac{1}{\rho} \frac{\partial p}{\partial x} = g \frac{\partial h}{\partial x}$, and likewise for the north-south direction $\frac{1}{\rho} \frac{\partial p}{\partial y} = g \frac{\partial h}{\partial y}$, where h is the height of the sea surface above the equipotential (level) surface.

The pressure gradient force per unit mass may also be expressed as the horizontal derivative of the dynamic height.

Dynamic height is the vertical integral $D = \int_0^p \alpha dp$, where the

specific volume $\alpha_{STp} = \frac{1}{\rho_{STp}}$; STp in this notation is salinity (‰), temperature (°C), and pressure, respectively. The advantage of this approach (recall $\alpha dp = -g dz$) is that spatial variations of gravity are included in the pressure integral. Thus the horizontal pressure gradient force per unit mass can be written as $\alpha \frac{\partial p}{\partial x} = g \frac{\partial h}{\partial x} = \frac{\partial D}{\partial x}$ and $\alpha \frac{\partial p}{\partial y} = g \frac{\partial h}{\partial y} = \frac{\partial D}{\partial y}$. In the dynamic method, D is typically integrated over 0–1000 db (1 decibar = 10^4 Pa) or deeper, and the $p = 0$ db surface so created, subtracting the standard ocean $\alpha_{35, 0, p}$, is the oceanic sea surface topography of the dynamic height anomaly ΔD . The units of D and ΔD are $\left[\frac{\text{m}^3}{\text{kg}} \right] \left[\frac{\text{N}}{\text{m}^2} \right] = \text{m}^2 \text{s}^{-2} = \frac{\text{J}}{\text{kg}}$, which is geopotential; $\frac{\Delta D}{10}$ is called a dynamic meter. Oceanographers commonly calculate the dynamic height anomaly $\Delta D = \int_0^p (\alpha_{STp} - \alpha_{35, 0, p}) dp$ and compute horizontal currents with respect to a deep reference surface (which is assumed to be a level surface), using the geostrophic equation.

Geostrophy further illustrates the role of horizontal pressure gradients in the sea. Consider the geostrophic equation $f \vec{v}_g = \frac{1}{\rho} \frac{\partial p}{\partial x}$, where f is the Coriolis parameter ($1.4584 \times 10^{-4} \sin \phi$), ϕ is the latitude, $\frac{\partial p}{\partial x}$ is the east-west direction pressure gradient force per unit volume, and $\vec{v}_g$ is the north-south geostrophic current. Recall that pressure is an integral $p = - \int_h^z \rho g \partial z$ from the sea surface ($z = h$) to some deep layer where $z = -Z$. Derivatives of an integral require application of Leibniz's rule, which yields

$$\rho f \vec{v}_g = \rho g \frac{\partial h}{\partial x} + \int_h^Z g \frac{\partial \rho}{\partial x} dz.$$

The term on the left-hand side is the force per unit volume due to Earth's rotation creating a geostrophic current, $\vec{v}_g$. On the right-hand side, the first term is proportional to the slope of the sea surface (∂h) with respect to a level surface (∂x), and the second term is that due to the internal horizontal gradient of the density field. The term $\rho g \frac{\partial h}{\partial x}$ is the barotropic pressure gradient force per unit volume, from which the surface geostrophic current $\vec{v}_g(z = 0)$ is calculated, and $\int_h^Z g \frac{\partial \rho}{\partial x} dz$ is the baroclinic pressure gradient force per unit volume.

To illustrate these terms, consider an east-west crossing of the Gulf Stream from Charleston, South Carolina, to Bermuda. The sea surface slopes upward to the east a total of about 1 m, and thus the barotropic term is positive. The subsurface horizontal density gradient $\frac{\partial \rho}{\partial x}$ however is negative, and at some depth $z = Z_0$, the integral of the baroclinic term will be equal in magnitude but opposite in sign to the barotropic term. Thus at this depth (*i. e.*, $z = Z_0$), $v_g = 0$, a depth called the level of no motion. In the offing of Charleston, Z_0 typically is assumed to occur at about $z = -2000$ m.

Conclusion

In the equation that describes fluid motion, the term that quantifies the movement is the pressure gradient force per unit mass, an acceleration. The pgf is a diagnostic term, not a prognostic term, most often calculated from measurement of the distribution of temperature and salinity or by direct

measurement with pressure gauges. The hydrostatic equation is the primary vertical pressure gradient balance, and the geostrophic equation the primary horizontal pressure gradient balance.

Cross-References

- [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
- [Gulf Shorelines, Last Eustatic Cycle](#)

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RADARSAT-2

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RADARSAT-2, the second in a series of Canadian spaceborne Synthetic Aperture Radar (SAR) satellites, was built by MacDonald Dettwiler, Richmond, Canada. RADARSAT-2, jointly funded by the Canadian Space Agency and MacDonald Dettwiler, represents a good example of public–private partnerships. RADARSAT-2 builds on the heritage of the RADARSAT-1 SAR satellite, which was launched in 1995. RADARSAT-2 will be a single-sensor polarimetric C-band SAR (5.405 GHz).

RADARSAT-2 retains the same capability as RADARSAT-1 (Morena et al. 2004). For example, the RADARSAT-2 has the same imaging modes as RADARSAT-1, and as well, the orbit parameters will be the same thus allowing co-registration of RADARSAT-1 and RADARSAT-2 images. Furthermore, radiometric and geometric calibration is maintained thus permitting correlation of time series data for applications such as long-term change detection (Luscombe 2001).

The following features of the RADARSAT-2 system are thought to be the most significant in terms of their impact on existing and new applications.

Polarization modes. Three polarization modes: Selective, Polarimetry, Selective Single.

Resolution. 3 m ultra-fine mode and a 10-m Multi-Look Fine mode.

Programming lead time. Programming is defined as the minimum time between receiving a request to program the satellite and the actual image acquisition. Routine image acquisition planning is base-lined at 12–24 h, and emergency acquisition planning is baselined at 4–12 h.

Processing. Routine processing is base-lined at 4 h; emergency processing is base-lined at 3 h; and 20 min for processing a single scene.

Re-visit. Re-visit is defined as the capability of the satellite to image the same geographic region. Re-visit is improved through the use of left- and right-looking capability.

Georeference. Image location knowledge of <300 m at down-link and <100 m post-processing.

RADARSAT-2 Polarimetry Modes

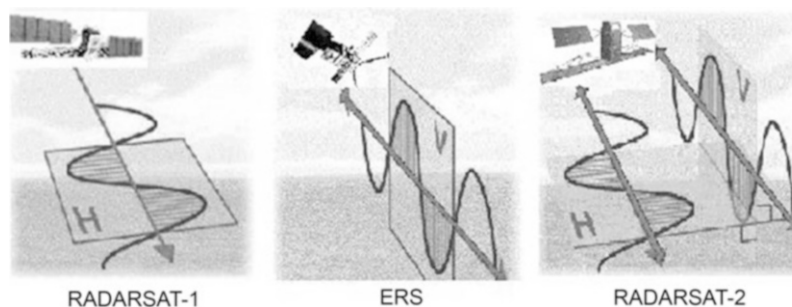
The RADARSAT-2 polarimetric capability is considered to be the most significant in terms of increasing the information content of the SAR imagery, and is subsequently discussed in more detail. To date, SAR data have been widely available from single channel (single frequency and polarization) spaceborne radars including ERS-1 and 2, JERS-1, and RADARSAT-1. RADARSAT-2 provides polarized data, and is the first spaceborne commercial SAR to offer polarimetry data (CASI 2004).

The intent here is not to outline polarimetry theory, but to present the concepts in an intuitive manner so that those not familiar with polarimetry can understand the benefits of polarimetry and the information available in polarimetry data. Many articles are available that discuss polarimetry theory, applications, and provide excellent background information (Ulaby and Elachi 1990). Notwithstanding the inherent complexity of polarimetry, polarimetry in its simplest terms refers to the orientation of the radar wave relative to the earth's surface and the phase information between polarization configurations.

RADARSAT-1 is horizontally polarized meaning the radar wave (the electric component of the radar wave) is horizontal to the earth's surface (Fig. 1). In contrast, the ERS SAR sensor was vertically polarized, implying the radar wave was vertical to the earth's surface. Spaceborne SAR sensors such as RADARSAT-2, ENVISAT, and the Shuttle Imaging Radar have the capability to send and receive data in both horizontal (HH) and vertical (VV) polarizations. Both the HH and VV polarization configurations are referred to as co-polarized

RADARSAT-2, Fig. 1

Orientation of horizontal (H) and vertical (V) polarization. Typical transmit and receive polarizations are HH, VV, and HV. (Adapted from the CCRS website)



RADARSAT-2, Table 1 RADARSAT-2 modes. Beam mode name, swath width, swath coverage, and nominal resolution

	Beam mode	Nominal swath width (km)	Swath coverage to left or right of ground track (km)	Approximate resolution Rng $\times$ Az (m ²)
Selective Polarization	Standard	100	250–750	25 $\times$ 28
	Wide	150	250–650	25 $\times$ 28
Transmit H or V	Low incidence	170	125–300	40 $\times$ 28
Receive H or V or (H and V)	High incidence	70	750–1000	20 $\times$ 28
	Fine	50	525–750	10 $\times$ 9
	ScanSAR wide	500	250–750	100 $\times$ 100
	ScanSAR narrow	300	300–720	50 $\times$ 50
Polarimetry				
Transmit H and V on alternate pulses	Standard QP	25	250–600	25 $\times$ 28
Receive H and V on every pulse	Fine QP	25	400–600	11 $\times$ 9
Selective Single Polarization				
Transmit H or V	Multiple fine	50	400–750	11 $\times$ 9
Receive H or V	Ultra-fine wide	20	400–550	3 $\times$ 3

modes. A second mode, the cross-polarized mode, combines horizontal send with vertical receive (HV) or vice versa (VH). As a rule, the law of reciprocity applies and HV–VH (Ulaby and Elachi, 1990).

A unique feature of RADARSAT-2 is the availability of polarimetry data, meaning that both the amplitude and the phase information are available. The amplitude information is familiar to SAR users, but the phase information is likely new and rather nonintuitive. In its simplest term, phase can be thought of as the travel time for the SAR signal: the travel time is the two-way time between the sensor and the earth, and includes any propagation delays as a result of surface or volume scattering. It is the propagation delays and the scattering properties of the HH and VV polarization configurations that make polarimetry data so powerful.

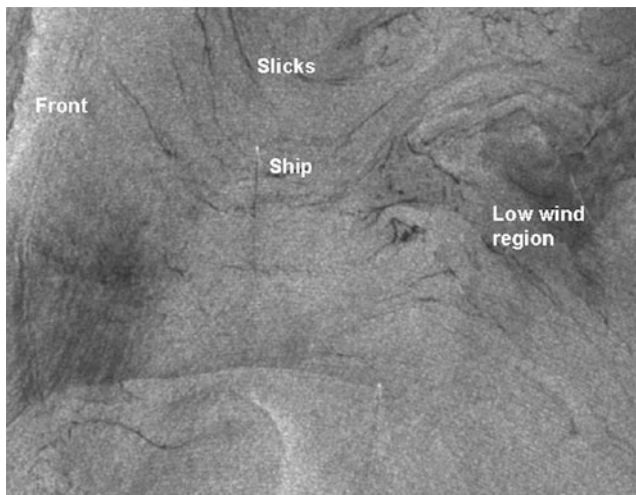
The RADARSAT-2 program has adopted the following terms to define the polarization modes (Table 1): Selective Polarization, Polarimetry, and Selective Single Polarization.

Selective Polarization and Selective Single Polarization modes imply the availability of amplitude data, but no interchannel phase data. For example, amplitude data may be HH, VV, or HV imagery. In contrast, the polarimetry mode (also called quad-polarized) implies the availability of both amplitude and interchannel phase information. The amplitude information is the same as the Selective Polarization and Selective Single Polarization modes, but adds phase information, such as the co-polarized phase difference.

Marine Applications

Marine applications of SAR data can be divided into three main categories: atmospheric phenomena, ship detection, and ocean features (Fig. 2).

Atmospheric phenomena include the effect of large-scale atmospheric features such as hurricanes on the ocean surface.



RADARSAT-2, Fig. 2 RADARSAT-1 SAR image acquired September, 1998 off the coast of Alaska. Typical ocean features and targets are shown. (Canadian Space Agency)

Although SAR images through the hurricane cloud-structure, the variability of the hurricane wind speed produces changes in the ocean surface-roughness that the radar detects. For example, the low-wind regime at the eye of the hurricane looks very different than the outer high-wind edges. The radar sensitivity to wind-induced roughness can also be used to map ocean-surface wind speed and direction. The use of VV polarization will be the preferred polarization configuration, largely due to a better radar response relative to HH or HV configurations.

Ship detection is optimal under low wind conditions, HH polarization, and large incidence angles. When co-polarized data and small incidence angles are used, there is increased radar return from the ocean surface, thus reducing the contrast between the ocean and the ship. The use of cross-polarized data (e.g., HV), however, enhances ship detection at small incidence angles due to the weaker return from the ocean surface, but similar return from the ship. Through the application of target decomposition algorithms, quad-polarized data can be used for ship detection and classification (Jeremy et al. 2001).

Ocean features include the detection of eddies, fronts, slicks, currents, surface waves, and internal waves. Radar return from the ocean surface is due to Bragg scattering. Bragg scattering is stronger for VV polarization, thus VV polarization is predominantly used versus HH or HV. The use of quad-polarized data will add significantly to the information content of the SAR imagery.

Cross-References

- [Remote Sensing of Coastal Environments](#)
- [Synthetic Aperture Radar Systems](#)

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Rating Beaches

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Definition

A beach rating survey was designed to provide an objective appraisal of the major public recreational beaches in the United States. Fifty criteria, which included physical and biological factors as well as human use and impacts, are used to evaluate beaches for their environmental qualities and utility by beach patrons, especially for swimming during the summer season. The factors were assessed on a scale from one to five in order to provide a quantitative comparison of beaches and to permit a ranking that is undertaken annually. The number one beach, which has the highest overall score, is retired the following year after selection as the national winner so that other great beaches can receive acclaim for their high-quality and good environmental management.

Introduction

Beaches are the number one recreational destination for Americans and Europeans, and a beach culture has developed worldwide. Nothing restores the body and soul like a stay at the beach. We are naturally drawn to the rhythmic pounding of the waves as if returning to our primordial beginnings. Recreational opportunities abound, and everyone, but perhaps children most of all, loves sand.

People are flocking to the shore in ever-increasing numbers for sun and fun. But most want much more from a beach experience – people are searching for real getaway places where they can escape from urban confinement and everyday pressures. The shore offers freedom from the “hemmed-in” feeling as we gaze out from the beach at the seemingly endless sea. The fresh, salty air invigorates the body as the sheer beauty and dynamic interplay between the waves and beach captures our imagination and refreshes our psyche.

So what do people look for in a beach? Water quality is probably the first concern. Polluted water can ruin any beach; the washing ashore of medical wastes along northern New Jersey beaches several decades ago nearly wrecked the local coastal economy as tourism plummeted. Some beaches, already stressed by pollution from upland sources, are also adversely affected by the onslaught of contaminants from coastal development and wastes dumped at sea. Sewage-associated wastes are an eyesore where inadequate treatment systems still exist. The Gulf coast states have the greatest accumulation of ship galley wastes on their beaches; these materials degrade near-shore coastal waters and foul beaches (Ocean Conservancy 2017). While concern about pollution is mounting as coastal development continues, some areas are addressing long-standing problems (e.g., construction of the massive Boston Harbor sewerage system). Actually, US beaches are generally quite clean compared to those in many countries, where pollution concerns receive far less attention and funding.

There is a worldwide coastward migration of the population, which itself is burgeoning at over seven billion at present. What some oceanographers refer to as the ring around the bathtub is some of the most expensive real estate in the world. These areas are rapidly urbanizing, and there is much public concern about the quality of coastal areas. For example, 39% of the US population, which amounted to 123 million people, lived in counties directly on the shoreline in 2010 (National Ocean Service 2017). As the population trend continues, many of the qualities that attracted people initially are diminishing.

At crowded beaches such as Jones Beach, New York, the sunbathing towels can run together like a gigantic patchwork quilt. While this beach can accommodate a huge number of people, few would rate this as the best beach although it does have good amenities. If coastal communities are too successful in encouraging development, overall quality can drop.

Beach weather and water temperature can greatly limit planned activities. The rugged beauty and serenity of Oregon's beaches appeal to the wilderness enthusiasts, but the water is much too cold for swimming, and frequent rainy days can spoil plans for sunbathing or more active activities. While the picturesque Northwest Pacific coast may be described as the most beautiful, others prefer the palm-treed beachscapes of Hawaii. Which is the most beautiful beach? Shakespeare correctly noted that "Beauty lies in the eyes of the beholder." But all of us can probably agree on a few basic characteristics, which include the physical condition, biological quality, and human use and development of beaches.

Beach Awards

Beach awards in Europe were once regarded as a valuable tool for promotion of beach tourism. The proliferation of awards

and awarding bodies in the United Kingdom, however, has led to low public awareness and distrust of their validity. Those currently in use in the United Kingdom include the Blue Flag Award (administered by the Foundation for Environmental Education; *FEED*), the Seaside Awards given by Keep Britain Tidy, and the *Good Beach Guide*, a book published by the Marine Conservation Society. These awards are based on a limited number of factors and do not approach coverage of all measurable aspects of the beach environment.

A total of 4,154 Blue Flags were awarded in 2015 in 60 countries. Only one-third of beach users have a reasonable understanding of the award criteria, which are largely based on water quality; 11% recognized the Blue Flag itself, and 7% thought the symbol represented danger (Williams and Morgan 1995).

The Seaside Awards were introduced in the United Kingdom in 1992 and is administered by Keep Britain Tidy. The award criteria are similar to the Blue Flag (e.g., water quality, beach cleanliness, and high standards of facilities and management). While the Blue Flag Award requires strict water quality standards, the Seaside Awards are less rigid so that more beaches in England can qualify for this recognition.

The Good Beach Guide helps one find the best UK beaches such as lifeguarded beaches and dog-friendly beaches. Beaches must exhibit a high standard of water quality and a low probability of sewage contamination (which is still a major problem for some European beaches).

The Blue Wave award was more recently introduced in the United States in order to mimic or perhaps confuse European tourists who might believe that it was the Blue Flag – which is certainly not the case. This program appears to be little more than a "pay-to-play" type of operation as the originator has no coastal scientific credentials.

Beach awards and flags are meant to signify good quality beaches and bring in money to the local economy, but what has developed in some cases is the flag jungle (Williams 2004). There is a high degree of bewilderment and lack of understanding with respect to beach flags, and their worth to the public is of dubious value. It appears that general media exposure and the Internet are a better bet to understand beach conditions than a flag flapping in the wind (Williams 2004).

Beach Ratings

The first beach rating scale was developed in the United States by Leatherman in response to an inquiry from the editor of a travel magazine in 1989. Fifty criteria were devised to measure the physical, biological, and human use/impacts while maintaining a balance between the natural and developed environment because most beachgoers are looking for some amenities (Table 1).

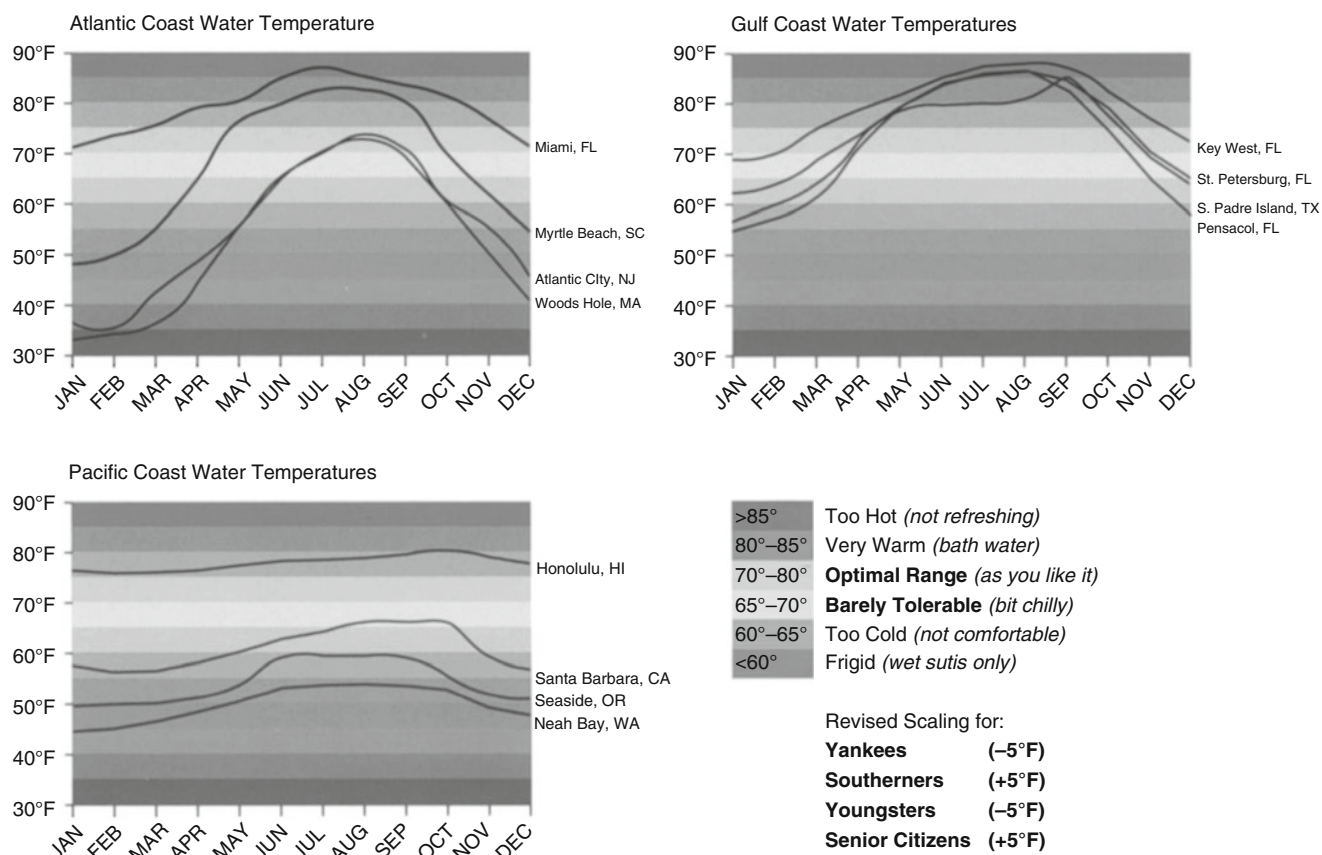
Rating Beaches, Table 1 Beach rating scale (Leatherman 1991, 1997; www.drbeach.org)

Criteria (relate to the vacation/holiday season)	Categories				
	1	2	3	4	5
1. Beach width at low tide	<10 m narrow	10–30 m	30–60 m	60–100 m	>100 m wide
2. Beach material	Cobbles	Sand/cobbles	Coarse sand	–	Fine sand
3. Beach condition or variation	Erosional	–	Stable	–	Depositional
4. Sand softness	Hard	–	–	–	Soft
5. Water temperature	Cold/hot	–	–	–	Warm (70–85 °F)
6. Air temperature (midday)	<60 °F	–	–	–	80–90 °F
	>100 °F				
7. Number of sunny days	Few	–	–	–	Many
8. Amount of rain	Large	–	–	–	Little
9. Wind speeds	High	–	–	–	Low
10. Size of breaking waves	High/dangerous	–	–	–	Low/safe
11. Number of waves/width of breaker zone	None	1	2	3	4+
12. Beach slope (underwater)	Steeply sloping bottom	–	–	–	Gently sloping bottom
13. Longshore current	Strong	–	–	–	Weak
14. Rip currents	Often	–	–	–	Never
15. Color of sand	Gray	Black	Brown	Light tan	White/pink
16. Tidal range	Large (>4 m)	3–4 m	2–3 m	1–2 m	Small (<1 m)
17. Beach shape	Straight	–	–	–	Pocket
18. Bathing area bottom conditions	Rocky, cobbles, mud	–	–	–	Fine sand
19. Turbidity	Turbid	–	–	–	Clear
20. Water color	Gray	–	–	–	Turquoise
21. Floating/suspended human material (sewage, scum)	Plentiful	–	–	–	None
22. Algae in water amount	Infested	–	–	–	Absent
23. Red tide	Common	–	–	–	None
24. Smell (e.g., seaweed, rotting fish)	Bad odors	–	–	–	Fresh salty air
25. Wildlife (e.g., shore birds)	None	–	–	–	Plentiful
26. Pests (biting flies, ticks, mosquitoes)	Common	–	–	–	No problem
27. Presence of sewage/runoff outfall lines on/across the beach	Several	–	–	–	None
28. Seaweed/jellyfish on the beach	Many	–	–	–	None
29. Trash and litter (paper, plastics, nets, ropes, planks)	Common	–	–	–	Rare
30. Oil and tar balls	Common	–	–	–	None
31. Glass and rubble	Common	–	–	–	Rare
32. Views and vistas – local scene	Obstructed	–	–	–	Unobstructed
33. Views and vistas – far vista	Confined	–	–	–	Unconfined
34. Buildings/urbanism	Overdeveloped	–	–	–	Pristine/wild
35. Access	Limited	–	–	–	Good
36. Misfits (nuclear power station, offshore dumping)	Present	–	–	–	None
37. Vegetation (nearby). Trees, dunes	None	–	–	–	Many
38. Well-kept grounds/promenades or natural environment	No	–	–	–	Yes
39. Amenities (showers, chairs, bars, etc.)	None	–	–	–	Some
40. Lifeguards	None	–	–	–	Present
41. Safety record (deaths)	Some	–	–	–	None
42. Domestic animals (e.g., dogs)	Many	–	–	–	None
43. Noise (cars, nearby highways, trains)	Much	–	–	–	Little
44. Noise (e.g., crowds, radios)	Much	–	–	–	Little
45. Presence of seawalls, riprap, concrete/rubble	Large amount	–	–	–	None
46. Intensity of beach use	Overcrowded	–	–	–	Ample open space
47. Off-road vehicles	Common	–	–	–	None

(continued)

Rating Beaches, Table 1 (continued)

Criteria (relate to the vacation/holiday season)	Categories				
	1	2	3	4	5
48. Floatables in water (garbage, toilet paper)	Common	—	—	—	None
49. Public safety (e.g., pickpockets, crime)	Common	—	—	—	Rare
50. Competition for free use of beach (e.g., fishermen, boaters, water skiers)	Many	—	—	—	Few

**Rating Beaches, Fig. 1** Water temperature scale for bathing and swimming

The beach rating survey was designed to provide an objective appraisal of the major public recreational beaches along the US Atlantic, Gulf, and Pacific Coasts. About 650 beaches were evaluated nationwide on the basis of 50 criteria with a sliding scale to quantify the elusive quality factor. In-state coastal experts provided information and assisted the author (Leatherman 1991, 1997) in this evaluation. Only the open ocean and Gulf Coast beaches were rated. Therefore, sea-shores along Long Island Sound in New York and Connecticut were not considered nor were the many small beaches in major bays such as the Chesapeake and San Francisco. Puget Sound is perhaps the most desirable coastal property in the State of Washington, but here again it is an inland marine water body. In addition, Alaska's long coastline with many sandy to cobble beaches was not evaluated, although the

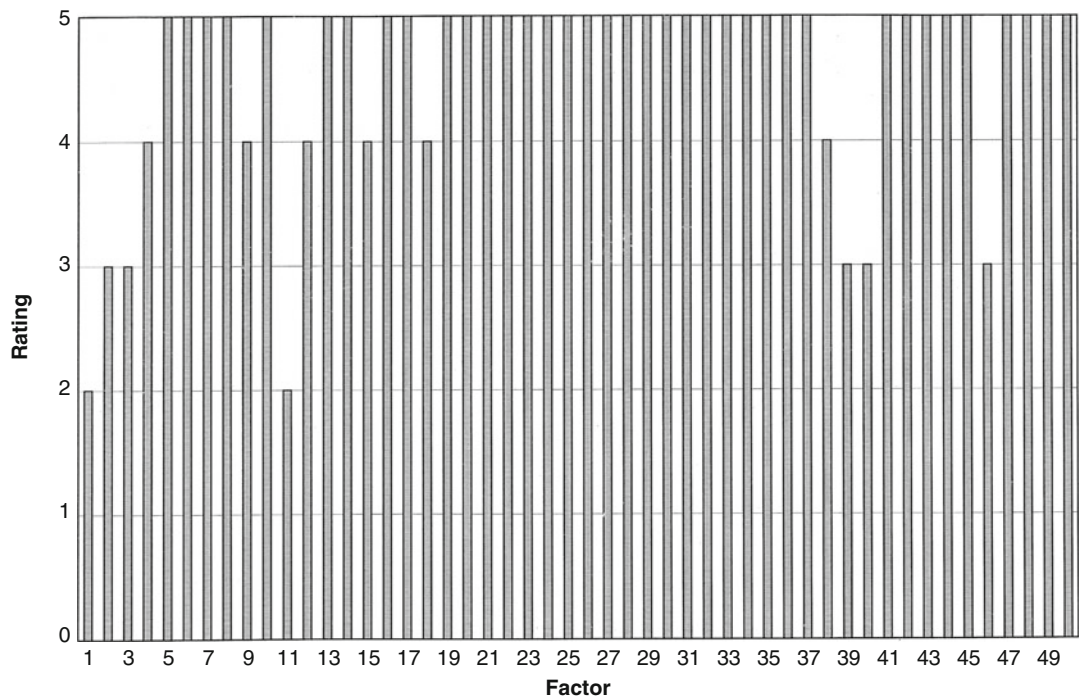
Valdez oil spill spotlighted this area's scenic beauty and the small pocket beaches where the oil tended to accumulate and pool.

A battery of factors was arrayed in order to allow for a quantitative comparison of the various beaches. The relevant criteria are those which influence beach quality as broadly defined. The factors considered in this analysis are of three types: physical, biological, and human use and impacts (Table 1). These ranged from 1 (poor) to 5 (excellent). This approach follows that developed by Leopold (1969) in her quantitative comparison of aesthetic factors for rivers.

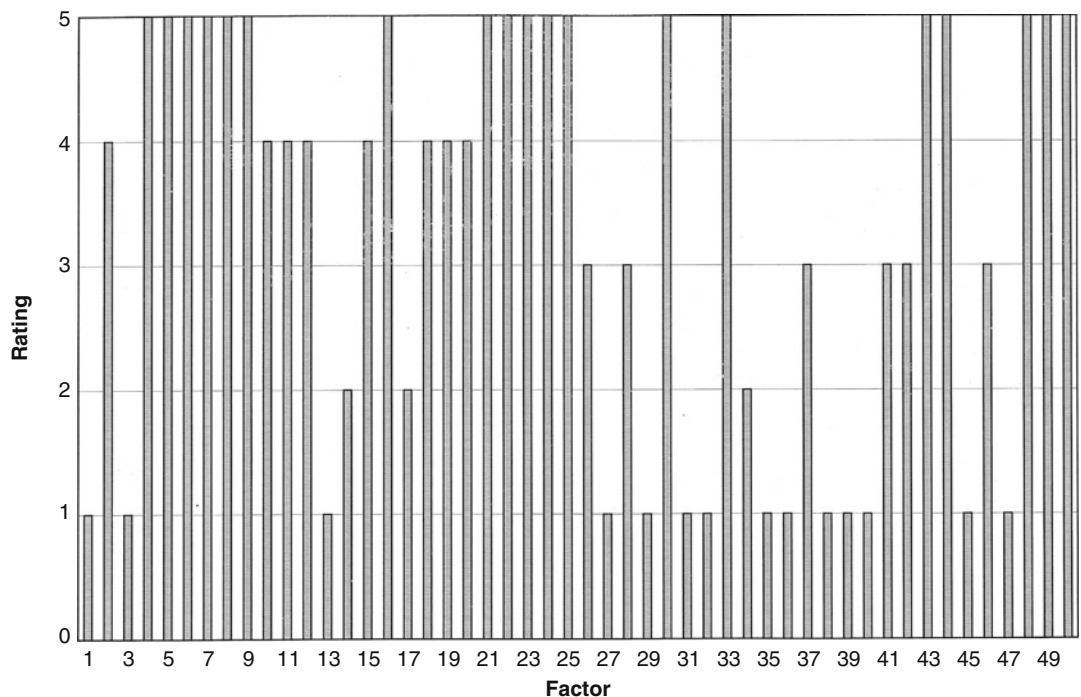
The survey was designed to reflect general beach usage with swimming water (see especially factors 5–14) being of primary importance. A water temperature scale for optimal

and tolerable conditions for swimming and bathing (Fig. 1) was also designed to aid in quantifying factor 5 of the questionnaire (see Table 1). In general, pristine beaches with limited development scored much better than the

overdeveloped and overcrowded urban resort areas. A profile can be developed for each beach based on the 50 factors evaluated; Fig. 2 illustrates this graphical representation for Kapalua on Maui, Hawaii – the top-rated beach nationally in



Rating Beaches, Fig. 2 Beach profile for top-rated Kapalua Bay Beach, Maui, Hawaii



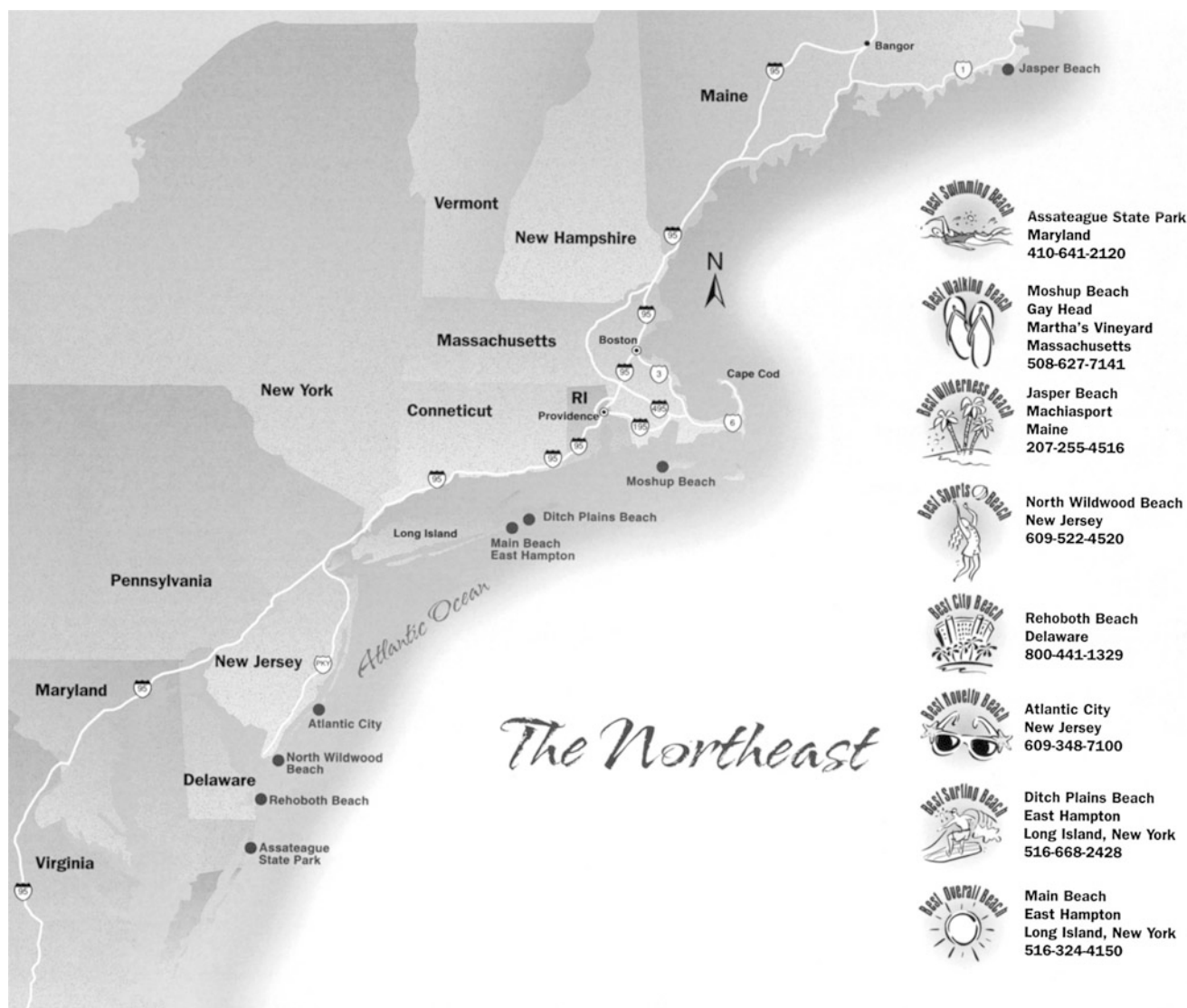
Rating Beaches, Fig. 3 Beach rating profile for Pike's Beach, New York. Note that the "sawtooth" graph indicates a less aesthetically pleasing and overall lower-quality beach

1991. For visual comparison, one of the nation's worst beaches (Pike's Beach in the New York City area) is presented (Fig. 3). This rating of America's best beaches has been conducted nearly three decades, and the results have appeared widely in the popular media. The Sarasota Convention and Visitors Bureau documented in 2011 that Dr. Beach's Top 10 Beaches List (www.DrBeach.org) was viewed by 425,959,156 people worldwide within several days after the Memorial Day announcement (Email from Erin Duggan, Sarasota CVB, June 3, 2011).

Specialty categories were determined by using a subset of the data. Figure 4 illustrates the ranking for the US northeast region as presented in America's Best Beaches (Leatherman 1998). Each region offers a different venue and recreational advantage. While swimming is the favorite activity of most beachgoers, many people just enjoy

walking along beaches for the exercise or to enjoy the scenery. Purists prefer wilderness beaches, but many vacationers are looking for creature comforts; the top city beaches offer all the amenities.

White and pink sand are the most highly rated, while gray sand is assigned the lowest rating. While pure white sand does cause much glare, the sugar-white beaches of the Florida Panhandle are considered the most beautiful by sunglass-wearing tourists. Also, the pink sand beaches of Bermuda are something to behold. Some factors are more important than others; this is dealt with by using a suite of factors that are all related to one variable such as pollution (Table 1, factors 21, 23, 27, 29–31, 43, 44, and 48) or beach safety (Table 1, factors 10–14, 21, 23, 28, 40, 41, and 49). Therefore, multiple factors are used to delineate important criteria in rating beaches.



Rating Beaches, Fig. 4 US beaches have been evaluated by region and category (Leatherman 1998)

America's most famous beaches were rarely rated the best ones for overall quality. For instance, the beaches at world famous Waikiki Beach in Honolulu, Hawaii, have experienced progressive erosion to the point that the beach simply does not exist in some sections. Coney Island in New York City was a very popular beach in the 1960s, but crime and other problems have taken their toll.

After a quarter of a century, the survey of the best US beaches was begun anew because some beaches had changed due to storm impacts, beach nourishment, pollution concerns, etc. The same 50 criteria are still being used, but extra points are now being awarded for no-smoking beaches because cigarette butts are the most common form of litter on many recreational beaches (Ariza and Leatherman 2012). Far too many people regard beaches as a large ash can instead of nature's greatest natural playground that needs to be cherished and protected.

Beach Scenery Surveys

Scenery is a very important component of beach tourism. Morgan (1999) developed a questionnaire to query recreational beach users in Wales, UK. Overall, scenic quality was rated as the most important factor in the beach environment, but it must be remembered that Welsh beaches are "cold water" and not that conducive to swimming. Sand and water quality were also highly rated. There were many observed differences in beach user preferences, depending upon the type of beach (e.g., urban to rural) that the user preferred to visit. Some vacationers wanted all the amenities and "creature comforts" of a beach resort (e.g., good hotels and restaurants, many water-based activities, and nightlife) compared to others who preferred the natural characteristics of a beach (e.g., fauna and flora, scenery, and camping). Morgan (1999) found that high environmental quality was a prerequisite for all beach users, emphasizing the high level of public concern for this aspect.

More recently, the Caribbean Coast of Colombia was assessed in terms of its scenery (Rangel-Buitrago et al. 2013). The study involved 135 sites which were analyzed based on 26 physical and human factors with a ranking scale from one to five. Natural areas ranked the highest with lower scores recorded for an increase in human parameters such as litter and sewage albeit beach nourishment was credited with higher scores. Litter can blight any beach.

Summary

Americans love beaches and lists, so the ranking of US beaches has been well received by the public as attendance

at the Top 10 Beaches spikes after the release of the list each Memorial Day weekend (e.g., typically the last weekend in May). A high beach rating is a huge positive for the tourism associations, but there are some citizens who would rather keep their beach out of the public eye. Beach ratings reward good beach management and encourage coastal managers and park officials to improve their beach in order to obtain a higher ranking. Beaches are always changing because of storm impacts, beach nourishment, pollution concerns, etc. so the rankings vary year to year (www.DrBeach.org).

Cross-References

- [Beach Awards and Certifications](#)
- [Beach Erosion](#)
- [Beach Management Tools](#)
- [Beach Safety Research](#)
- [Beach Use and Behaviors](#)
- [Carrying Capacity in Coastal Areas](#)
- [Coastal Scenery](#)
- [Marine Litter](#)

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Reclamation

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Although normally associated with the rehabilitation of land, reclamation in the coastal context refers to the exclusion of marine or estuarine water from formerly submerged land. The basic idea of reclamation is to win land from the sea, to displace water and to create new land (Plant et al. 1998, p. 563). The resulting land surface normally extends from the existing coastline and should be well above the level reached by the sea. Reclamation differs from the building up of shallow offshore grounds to form *artificial islands* (q.v.) (Kondo 1995). It also differs from *polders* (q.v.) (CUR 1993, p. 230, 244) in which the level of land subject to seasonal or permanent high water level is protected by dikes, and flood control and water management are important aspects.

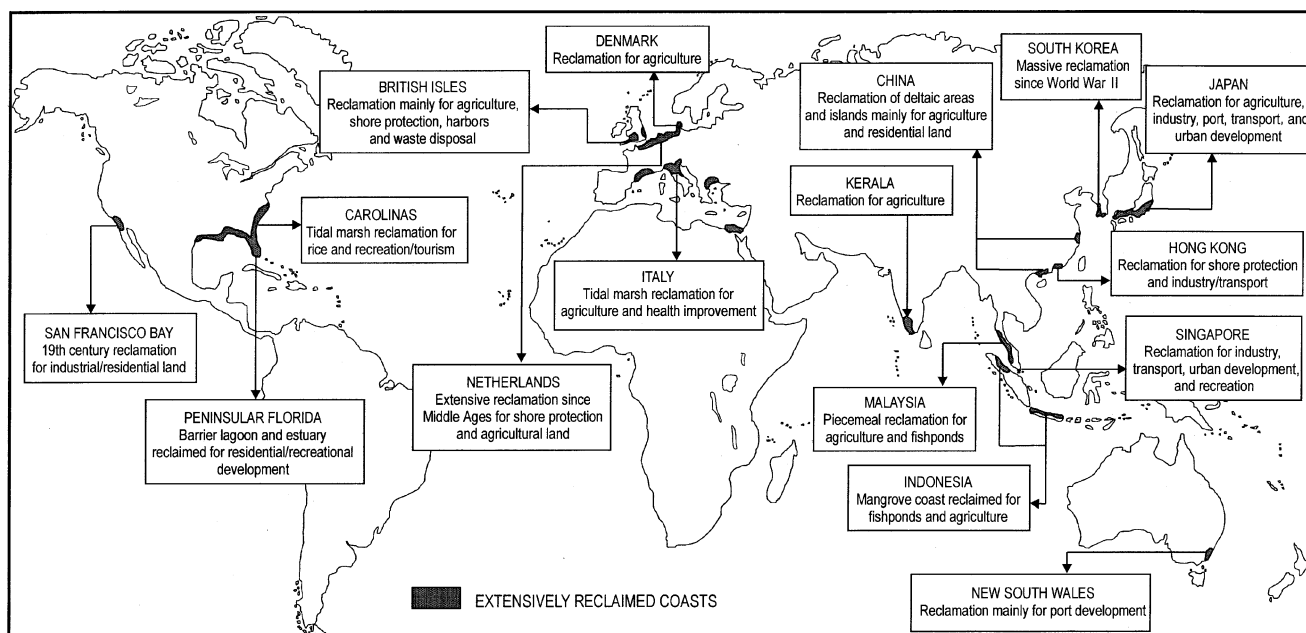
Historical and Geographical Brief

The origins of reclamation probably date back to human-kind's efforts in reclaiming land from estuaries and along coasts for their homes, livestock, and crops. For example, in the first few centuries, artificial clay mounds ("terps") were constructed by people in the lower parts of the Netherlands (Pilarczyk 2000). It is likely that reclamation initially

involved the draining of low-lying land and marshes. A survey of reclamation in temperate countries shows reclamation was primarily for agricultural purposes. Later, expanding population and increasing needs for industrial expansion and dock development led to a more focused objective of gaining land from the sea. It started with small-scale reclamation in almost all of the larger cities and ports situated on estuaries and coast throughout the world (Cole and Knights 1979) (Fig. 1). In fact, the reclamation of estuaries was better addressed, especially in the second-half of the nineteenth century to provide dock and harbor facilities to meet world trade expansion during the period of rapid industrialization in Europe and the United States (Kendrick 1994).

Any mention of gaining land from the sea cannot avoid the Netherlands because of the massive scale of reclamation in a country where the battle against the sea has become an integral part of the way of life. Reclamation was carried out in conjunction with poldering and the construction of sea defense works. Many lessons were learnt not only in engineering aspects but also in the physical and economic planning of reclaimed land. Another country in which reclamation of its estuaries and low-lying coastal areas has been carried extensively is Japan, which has at least 400 years of reclamation history. It has considerable experience in the consolidation and settlement of fill material, particularly in the treatment of soft foundations, such as the wide application of the sand drain method and the sand compaction method (Watari et al. 1994).

The need for reclaimed land for various developments varies worldwide and becomes particularly crucial around existing ports and cities and in countries where land scarcity



Reclamation, Fig. 1 Major reclaimed coasts in the world

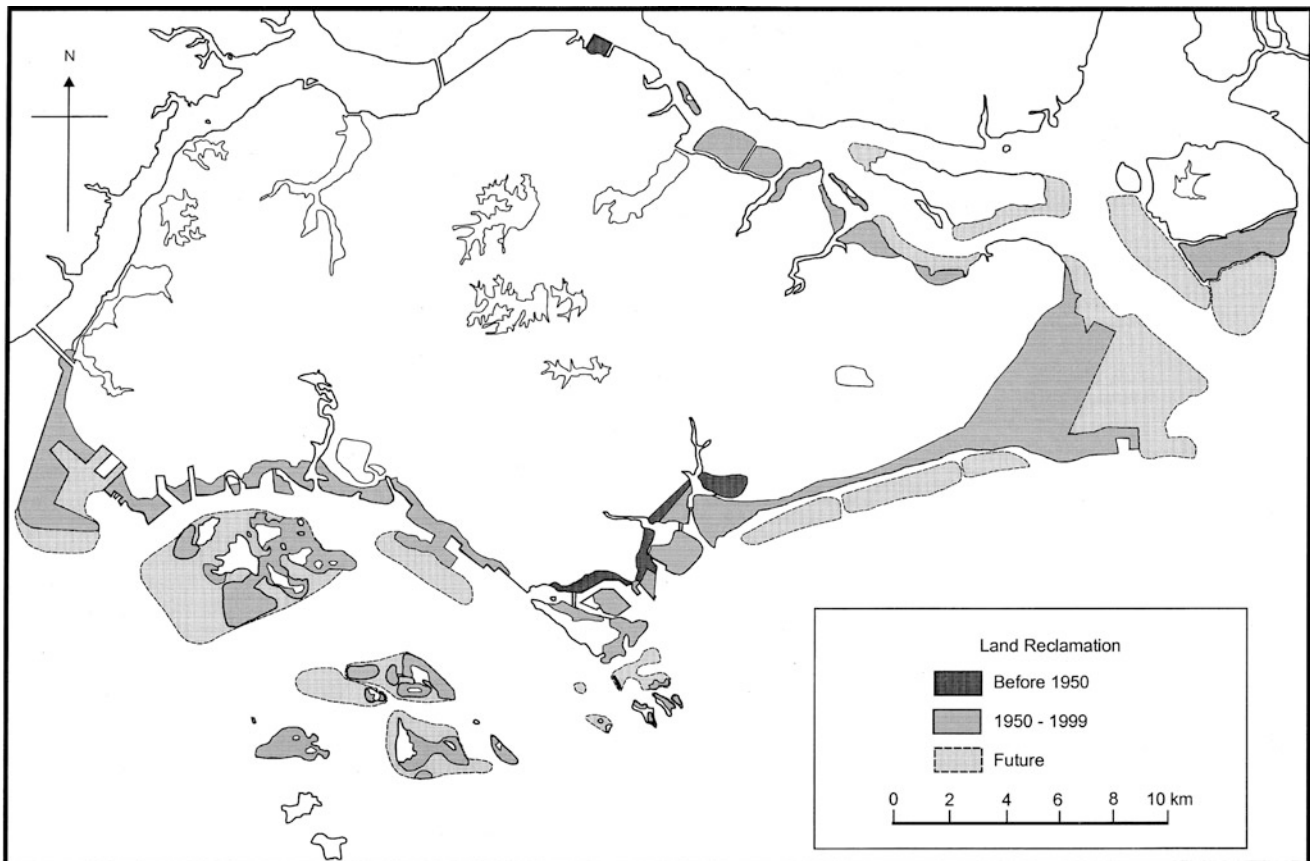
is a factor, for example, in Singapore and Hong Kong (Walker 1988) (Fig. 1). Faced with land scarcity, Singapore has resorted to large-scale reclamation from the 1960s although reclamation dated back to the nineteenth century. Through various projects, it has reclaimed about 10% of its area, which totaled 647 km² in 1998. Under the Concept Plan to accommodate a population of 4 million from the present 3.2 million, another 100 km² have to be reclaimed. This would amount to 25% in land area since the 1960s (Fig. 2). With current technology, it is not cost-effective to reclaim in waters deeper than 15 m. Another limitation is that reclamation encroaches into the sea space for shipping lanes and anchorage.

Reclamation Methods

Land won from the sea is normally carried out by raising the level of previously submerged land above sea level using materials dredged from the sea or excavated from the land. The reclaimed site is where reclamation is to take place and eventually forms the platform for required land use. The borrow area is where such material is obtained for reclamation and the disposal area is where material, unsuitable for fill, is to be disposed.

Depending on the type of landfill, there are two major types of reclamation and each has its advantages and disadvantages (Plant et al. 1998, p. 197, 561; CUR 1993, p. 244).

1. In the “drained” reclamation, fill is placed on the reclaimed site. The advantage is that existing soft marine material can be left in place, reducing the amount of fill required. The disadvantage is the longer time required for settlement and a greater uncertainty in the duration and rate of consolidation. This could increase the costs in follow-up structures to accommodate a larger settlement. Artificial drainage is required to reduce drainage paths and accelerate settlement (Plant et al. 1998, p. 561).
2. In the “dredged” reclamation, landfill by is by replacement, in which soft marine material is removed and replaced by imported fill. This means a larger volume of fill but the advantage of faster consolidation. The Hong Kong International Airport at Chek Lap Kok is an example of “dredged” reclamation in which three-quarters of the airport platform was reclaimed from the sea using a combination of dredging mud from the seabed, borrowing marine sand, excavation of islands, and construction of seawalls.



Reclamation, Fig. 2 Reclaimed land in Singapore

Site conditions have a strong influence on reclamation and factors such as climate, site access and site geology are important (Plant et al. 1998, p. 15). Inaccurate information on these factors has led to delays, disruptions, disputes, and operational failures. For example, heavy rainstorms can halt reclamation operation. Marine conditions as influenced by climate can be equally significant, for example, water depth and wave conditions influence the operation of dredgers. Seabed conditions can vary widely and water depth and type of materials influence both settlement and stability. A low-energy environment is advantageous as fill material can be placed on the seabed and not easily scoured by wave action. The availability of power, water, and communication services at the site can influence the design of reclamation.

Much of the early reclamation efforts often capitalized on the availability of fills such as quarry waste, urban refuse, excavation spoil, building demolition spoil, dredged material from ports and navigation channels, or the byproduct of some dredging operation. It was also carried out in connection with shallow tidal, marshy, or mangrove areas. Usually, for specific reclamation projects, materials cut from hills or seabed sand have been the most common fill. But with increasing scarcity of such fill materials, alternative sources, such as marine clays were dredged from the seabed. Marine aggregates, primarily sand and gravel from offshore-submerged sources, are also used in reclamation projects. In 1989, 7.7 million m³ and more than 9 million m³ were used for reclamation works in Denmark and Netherlands, respectively (Bokuniewicz 2000).

Various techniques have been in use to deal with the reclamation of coastal areas with soft foundations. For example, in Japan, the removal of the soft material and replacement by sand was the common method. The lack of sand and environmental problems, such as increased turbidity of water and the problem of disposal of dredged material, saw the end of this method for large-scale reclamation in 1975. This was replaced by the sand drain, which first came into use in 1952. From 1970–83 some 230,000 sand drains, each about 20 m long, were used for seabed foundation improvement in Japan. Driving-type sand compaction was used in 1966 in offshore construction and supplemented by the deep mixing method (DMM) from the mid-1970s (Watari et al. 1994).

For its earlier phases of reclamation, Singapore obtained landfill from the cutting of hills in the construction of public housing where extensive platforms were required. In the east coast reclamation, the fill was excavated by bucket wheel excavators and transported by belt conveyors to a loading jetty for loading to barges and dumped directly into the reclaimed site. Where water depth limited the movement of barges, the fill was unloaded by a reclaimer and conveyor system. Bulldozers and dump trucks then spread, graded, and compacted the reclaimed land (Yong et al. 1991). The reclaimed land required little settlement after compaction.

As materials from land become less readily available, Singapore has to import from neighboring countries or obtain them from marine resources. The typical procedure in large-scale coastal reclamation is as follows (Yong et al. 1991; Chuah and Tan 1995).

1. Seabed stabilization is first carried out in several ways by the excavation of soft material from the seabed, dredging to form a sandkey trench or the installation of sand compaction piles or vertical drains. Sand compaction piles were first used in 1989. A hollow pipe casing is driven into soft material, filled with sand, then compacted, and the casing is withdrawn. Sand piles were placed at close intervals. This method strengthens the clay seabed, obviated dredging, and thus avoids the costs of disposal.
2. A sandkey is formed from transported sand towed in by hopper barges. In some cases, where conditions are suitable, a sand wall is also constructed along the coast to be reclaimed. Sand barges build up a stockpile outside the sand wall.
3. Reclamation is carried out by direct dumping or hydraulic filling using cutter suction dredgers and pumps or trailer hopper suction dredgers. Where sand is sucked from the stockpile, it is spread into the fill area by a floating spreader. If the depth is too shallow, sand is pumped through overland pipes.
4. After filling, compaction is carried out by rollers. Shore trimming and shore protection works (geotextile placing, stone placing, and handpitching) complete the reclamation. The reclaimed land is usually ready for development after 1–5 years.

The above procedure has been applied to the reclamation of a group of islands located about 1 km south of Jurong Industrial Estate to form a single island zoned for petrochemical and chemical industries. The shape of the reclaimed island is to give a maximum land area that could be possibly reclaimed and a coastline with harbor basins that have no adverse conditions on water currents, sedimentation, and navigation (Chuah and Tan 1995).

Ground Treatment

The magnitude and rate of settlement of in situ soils and fills is a major concern in reclamation. The settlement depends on the type of fill, method of placement, and use of reclaimed area (Plant et al. 1998, p. 131). Ground treatment is any process in which the properties of the ground are improved or changed. Various methods are used to speedup settlement, each has its own merits.

Surcharging or preloading is the process in which the ground surface is loaded with additional mass. A stockpile

of material is required and the material is moved from location to location until the process is completed (rolling surcharge). This procedure avoids importing vast quantities of additional material. Where rockfill is used, surcharging is an effective means of reducing future settlement in rockfilled areas (Plant et al. 1998, p. 417) and less disruptive to the geotextile over rockfill. Fine-grained dredged materials take up to 2–3 years or more to consolidate under selfweight before surcharging is used. Although surcharging avoids the costs of handling large quantities of surcharge fill material and its eventual disposal, the whole process can take up to a decade or more for a 200-acre site (Thevanayagam et al. 1994).

Vibrocompaction refers to ground treatment in which heavy vibrators are inserted into loose granular soils and then withdrawn leaving a column of compacted soil in the ground. This reduces creep and vibration-induced settlement of the ground surface during follow-up construction activities. It is a fast method of achieving a high degree of compaction for a specific material. In dynamic compaction, the ground is compacted by high-energy impacts using a tampering weight. It can be used for all fill types but its effective depth is limited to about 10 m.

To facilitate rapid settlement associated with very soft clayey material, water conduits have to be installed for easy dissipation of water. In the past, this was achieved by vertical columns of sand drains installed at a grid of about 1–3 m. In recent years, sand drains are being replaced by wick drains or band-shaped drains, which make installing faster and easier. In the reclamation of a seabed underlain by thick deposits of soft clay, soil improvement is required. One of the most popular procedures is to combine prefabricated band-shaped drains with preloading.

In the reclamation for Hong Kong International Airport at Chek Lap Kok, seabed conditions varied widely in terms of water depth, thickness of marine mud to be dredged, and underlying compressible strata. These had a strong influence on both settlement and stability of the reclaimed land. The time taken for settlement was difficult to determine as the settlement of fill after construction depends on the type of fill, method of placement, and use of filled area. Instruments monitored settlement and allowed analytical settlement models to be calibrated and updated as reclamation and follow-up works progressed. An extra 0.5 m was provided for future settlement. For ground treatment, surcharge, vibrocompaction, and dynamic compaction were used (Plant et al. 1998).

Coastal Protection

Land gained from sea has to be protected from erosion by waves and currents. Depending on coastal conditions, various types of halophytes can be used to protect reclaimed land. For

example, marram grass (*Ammophila*) is used in Europe (Cole and Knights 1979). In Singapore, idle reclaimed land along Changi Coast Road has been effectively colonized by beach vegetation, such as *Ipomoea pes-caprae* and trees such as *Casuarina equisetifolia*.

More often, reclaimed land is protected by shore protection works. In accordance with current international practice, seawall construction allows for some risk of damage under extreme conditions. A high factor of safety will result in a disproportionate effect on costs. As the philosophy on coastal protection changes, this is also reflected in the protection of reclaimed land. For example, in Japan, coastal protection began with seawalls, followed by groins and then detached breakwaters (Hsu et al. 2000).

Although Singapore is not in a high wave-energy environment, various types of coastal protection works were required for the reclaimed land depending on several variables, such as the usage of reclaimed land, ease and speed of construction, site constraints, cost of construction, and predominant wave approach. Three major groups of shore protection works for reclaimed land in Singapore can be identified.

1. Reclaimed land with beaches. Initially, a seawall was used but this was superseded by a series of breakwaters acting as headlands between which beaches can be developed. These were used in the east coast because of the low wave energy, predominant wave approach from the southeast, and net littoral drift to the west. Beaches were formed in J-shaped bays with their upcoast curves in the east and downcoast straight sectors in the west.
2. Reclaimed land for port purposes. A sand key provides the broad base for the seawall to be constructed and withstand loads, for example, in Jurong. Sheet piles are used for deep-water frontage where wharf construction is required.
3. Reclaimed land retained behind a marine retaining wall. Various types of structures are used. Generally, stone bunds are constructed with stones large enough to withstand dynamic lifting and to absorb forces of waves on their outer face. Stone bunds were constructed around some islands with gaps for beaches to form (Wong 1985). Various prefabricated structures were also used. Caissons or huge reinforced concrete boxes were towed to sea by tugboats, then filled with sand and positioned on the seabed. These are costly and used where a rocky seabed is available. Depending on their size to be used, L-blocks of 30–70 tons require a compacted rubble foundation and are positioned by cranes. These were used in the reclamation of Marina Bay and Tanjung Rhu.

Depending on the size of area to be reclaimed, the condition of the seabed which may require stabilization, and the necessity of shore protection to protect reclaimed land from erosion, the following approaches can be identified for

reclamation in Singapore. For small areas, such as offshore islands, site preparation is followed by seabed stabilization for the construction of breakwaters to prevent erosion from currents and waves and followed by the filling of sand. For large areas extending from the mainland, site preparation is followed by seabed stabilization, filling of sand, and shore protection works. For large swampy areas, settlement is a problem and additional time and costs are required for shore protection measures. Site preparation is thus followed by sand filling and shore protection works.

New Developments

Reclamation often involves work on land and sea, which is collectively referred to as land operations and marine operations, respectively. The land operation is basically civil engineering and quarrying operations and the scale of earthworks is typical of large mining operations. The marine operation usually involves dredging and coastal protection. Dredging is defined as “underwater excavation of seabed material, transportation of the materials to a discharge area, and subsequent discharging of dredged material” (Plant et al. 1998, p. 266). In recent years, reclamation has seen new developments in both land and marine operations.

Modern reclamation relies on some heavy specialist equipment to recover (dredge) from the burrow area, transport, and place the material over the reclaimed site. Cutter-suction dredgers and trailing-suction hopper dredgers (also called trailer dredgers) have been developed to dredge a greater volume of materials in a short time and at lower costs, often in deeper waters. A trailing-suction hopper dredger is a selfproposed ocean-going vessel fitted with special dredging equipment. The hopper capacity of these dredgers has doubled from 10,000 m³ in the 1980s to 23,000 m³ and set to treble to 33,000 m³ in 2000 (Riddell 2000). A cutter-suction dredger differs from a trailing-suction hopper dredger in that the former is effectively stationary during dredging, has more control over the dredging process and is also capable of dredging harder material. Depending on water depth, a cutter-suction dredger trails as dredging progresses. The dredged material is projected through a nozzle at its bow in a process called “rainbowing” as land is reclaimed. Hopper barges are for direct dumping of material and are confined to sheltered waters.

The increasing use of geosynthetics in coastal and harbor engineering has also found its place in reclamation. Geosynthetics is a generic name given to various materials that are synthesized for use with geological materials to improve or modify their behavior. One major success was the construction of storm-resistant structures over soft soils at the coast in the Netherlands (Rao and Sarkar 1998). In Japan,

geosynthetics have been commonly used for more than three decades in land reclamation involving a soft clay foundation. They are mainly used for surface stabilization to increase bearing capacity and to reinforce the base of fill and for foundation improvement to replace sand drains in facilitating drainage (Akagi 1998). Probably the world’s largest sewn single sheet of geotextile used for reclamation took place in Singapore in one phase of the reclamation for Changi Airport. A 180-ha pond, 2000 m in length and 750 m and 1050 m in width at both ends, created earlier by the borrowing of sand contained slurry-like material 3 to 20 m thick. The initial spreading of sand over the pond failed as mud burst through the sand cap. Remedial measures started with the removal of some of the exposed slurry. Then, the geotextile sheet of 1,060,000 m² sewn from 5 × 90 m² rolls was laid across the silt pond to strengthen the foundation soil. This was supplemented by prefabricated band-shaped vertical drains to accelerate the consolidation time by shortening the drainage path (Na et al. 1998). This example shows that high-strength geotextile can strengthen foundation soil that is extremely soft.

The technique of vacuum consolidation is being used in reclamation. This is a process in which a vacuum is applied to a soil mass to produce a negative pore pressure, leading to increased effective stress that leads to consolidation. Although the technique was known in the early 1950s, it was not widely used until the 1980s because of high costs and implementation difficulties. Its application was made possible by recent technological advances in geotextiles and efficient and cost-effective prefabricated vertical drains (wick drains) (Shang et al. 1998). The technique has been field-tested for on-land vacuum consolidation in countries such as the Netherlands, France, Malaysia, Sweden, Japan, and China. It shows considerable promise as an economically viable method to replace or supplement surcharge fill and the potential to strengthen weak sediments on the seabed adjacent to or beneath water, or consolidate fine-grained hydraulic fills during construction. The technique can also be used with prefabricated horizontal drains and selective placement of dredged materials for new land reclamation (Thevanayagam et al. 1994). It was applied to 480,000 m² of reclaimed land in Tianjin Harbor, China, where individual sites treated with vacuum ranged from 5000 to 30,000 m², illustrating that it was especially attractive for hydraulic fills and in reclamation sites with a shortage of surcharge fills but that have an easy access to a power supply. Currently, research on the method is on numerical modeling of the consolidation process, the application of geotextiles over larger areas, and the development of high-efficiency vacuum equipment (Shang et al. 1998).

During earthquakes, especially in Japan, some reclaimed land can undergo a complex phenomenon called liquefaction in which loose sandy deposits change into a liquid state.

The liquefaction sites are related to the age of reclamation, the methods used, and type of material. The remediation methods depend on large-scale or localized remediation and are implemented during or after completion of land reclamation. With wide experience in this area, Japan has produced a handbook on remediation measures (Port and Harbor Research Institute 1997) that can be used in other seismically active regions where reclamation has been carried out.

Compared with the past, environmental considerations are given serious attention in modern reclamation and associated works (Bates 1994). The main areas of impact are at the dredging site, the transportation route, and the reclamation site. The potential adverse effects of dredging include the release of contaminants into the water, increased turbidity, disturbance to the seabed, erosion, noise, oil spillage, blanket cover, and consequent loss of habitat. To some extent, the impacts resulting from dredging and dumping may be overcome by changing the type of dredger and method of dumping. With the concern on the impacts on water quality and ecology, there is a need to develop methods to minimize environmental impact, such as modeling to examine environmental impacts of reclamation projects.

Conclusion

Reclamation continues to be a significant means in providing land to meet the needs of expanding population and economic development, especially in small countries and coastal cities where land is scarce. It brings about a complete change in the coastal environment. Its implementation involves the input of various disciplines, for example, physical geography, geology, soil mechanics, loose boundary hydraulics, land drainage, coastal engineering in the planning and design of reclamation, and ecology, with the increasing concern on loss of natural habitats. New developments have to try to overcome the technical constraints, make reclamation low cost, take advantage of available soft material and other fills, such as incinerator ash, and to reclaim further into increasing depths.

Cross-References

- [Artificial Islands](#)
- [Beach Drainage](#)
- [Bioengineered Shore Protection](#)
- [Dredging of Coastal Environments](#)
- [Geotextile Applications](#)
- [Polders](#)
- [Shore Protection Structures](#)
- [Wetland Restoration](#)

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Reefs, Non-coral

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Although most modern reefs are communities of coral and coralline algae that live in clear, well-lit tropical and subtropical waters, there are many different groups of reef-forming organisms that are found on living and ancient reefs. The modern non-coral reefs thrive in a wide range of environments extending from sponge reefs in the arctic to non-photosymbiotic algae and *Halimeda* reefs found near methane seeping faults at depths of 600 m (Wood 1999).

Definitions of Reefs

The word, reef, is derived from the Norwegian word, *rif*, which means rib. In nautical terms, reef refers to a narrow chain of rocks, shingle or sand lying at or near to the surface of the water. When early sailing ships explored the tropical waters of the South Pacific, they encountered ring-like reefs of coral that enclosed a lagoon which they called “atoll” after “atolu,” the Malayalam name for the Maldives Islands. In the more restrictive modern use of the word, reef denotes a rigid, wave-resistant framework constructed by large skeletal organisms (Ladd 1944). While living coral reefs on atolls are wave resistant and contain a framework of corals and algae, boreholes drilled beneath the reefs consist of rubble, sediment, and voids (Hubbard et al. 1990). Many ancient carbonate buildups, that are referred to as reefs, show that the original coral framework is almost completely destroyed by deep burial and diagenesis. A broader definition of reef, which would encompass both modern and ancient non-coral reefs, has been proposed by Rachel Wood: “a reef is a discrete carbonate structure formed by in-situ organic components that develops topographic relief upon the Seafloor” (Wood 1999, p. 5).

Reef-Forming Organism on Non-coral Reefs

The earliest recognized reefs are composed of stromatolites which were found on Phanerozoic carbonate platforms dating back to 2.5 Ga. Stromatolites are finely laminated microbialites produced by photosynthetic blue-green algae (cyanobacteria) that form a range of morphologies including domes, columns, and mounds (Reitner 1993). Ancient stromatolite reefs were constructed on preexisting carbonate ramps and rimmed shelves (Grotzinger 1989). Living stromatolites are found in intertidal zones on Lizard Island in the

Great Barrier Reef, in Shark Bay, Australia, and in submerged tidal channels on Lee Stocking Island in the Bahamas. Thrombolites, which also form reefs, are non-laminated microbial structures which often have a mottled or bioturbated appearance.

Archeocyathids were the first metazoan reef-forming organisms and are found in Lower Cambrian limestones. The archeocyathids are large sponges with double-walled inverted conical calcareous skeletons (Debrenne and Zhuravlev 1994). The first bryozoan reefs appeared in deep cold waters perhaps related to the presence of microbial mounds in the Lower Ordovician (Pratt 1989). Stenolaemate bryozoa colonies of clonally calcified chambers, which formed reefs in the Lower Ordovician and died out during the Permian, reappeared as gymnolaemate bryozoans in the Jurassic and expanded during the Cretaceous and Eocene. Phylloid algae are calcified algae of platy, cup, and encrusting leaf-like forms that inhabited many late Paleozoic reefs.

Rudistid reefs are common in the Jurassic to Cretaceous limestones. Rudists are heavily calcified, heterodont bivalves in which the hinge and ligament have been modified forming a complete uncoiling of both valves (Skelton 1991). The large lower (right) valve is conical, cylindrical, or coiled and the upper (left) valve is flattened. Most rudists were semi-infauna, soft sediment dwellers, but they often colonized storm-generated debris forming large rudistid reefs.

Large colonies of the common oyster, *Crassostrea virginica*, are found in intertidal to subtidal environments including sounds and estuaries where the salinity is between 5 and 30 ppt. The oyster spat becomes cemented to old oyster shells and forms mounds of oysters which are commonly known as oyster reefs. When the buried oyster beds are exposed as fossils, they take on a reef-like form, but do not resemble the modern coral reefs.

Conclusions

Most living reefs are composed of coral and coralline algae and consist of a wave-resistant framework constructed by large skeletal organisms. A broader definition of a reef is “a discrete carbonate structure formed by in-situ organic components that develops topographic relief upon the sea floor” (Wood 1999, p. 5). A wide variety of non-coral organisms including coralline algae, stromatolites, archeocyathids, bryozoans, rudists, and oysters formed reefs in the geologic past and many are still forming reefs today.

Cross-References

- [Algal Rims](#)
- [Australia, Coastal Ecology](#)

- Bioherms and Biostromes
- Caribbean Islands, Coastal Ecology and Geomorphology
- Coral Reef Coasts
- Coral Reefs

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Reflective Beaches

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Definition and Classification

Reflective beaches are systems where there is minimal wave-energy dissipation by breaking, and therefore most incident wave energy is reflected back offshore by the nearshore morphology. In cases of strong reflection, individual reflected waves can be seen propagating away from the foreshore. Guza (1974), in his study of beach cusp formation, was apparently the first to use the term reflective beach. He

distinguished between reflective and dissipative beaches (*q.v.*) using the surf-scaling parameter, ε :

$$\varepsilon = \frac{\alpha \omega^2}{g \tan^2 \beta}$$

where α is the wave amplitude at breaking, ω is the wave radian frequency ($\omega = 2\pi/T$, where T is wave period), g is the gravity constant, and β is the beach slope in degrees. The proportion of incident wave energy that is reflected from the beach increases as ε decreases. For beaches where ε is larger than 20, most energy is dissipated by the turbulence associated with wave breaking. Where ε is less than about 2.5, most wave energy is reflected off the foreshore, and such beaches are designated as reflective. Thus, Guza (1974) used the relative degree of reflection or dissipation of incident waves as a rationale for the classification of beaches. This approach was subsequently subsumed under the rubric of nearshore morphodynamics.

Nearshore Morphodynamics

The concept of nearshore morphodynamics was developed to characterize systems where form and process are closely coupled through feedback mechanisms (Wright and Thom 1977; Short and Jackson 2013). On beaches, waves (*q.v.*) interact strongly with sediments and morphology, and the form of wave breaking is one manifestation of these interactions. For a given wave steepness, H/L (where H is wave height and L is wave length), the breaker type will change as the nearshore slope changes. On a very low gradient slope, spilling breakers should occur. As the gradient increases, there should be a progression through plunging and collapsing breakers. Finally, on very steep beaches, surging breakers should occur (Galvin 1968). For a constant nearshore slope, the same sequence of breaker types will occur as wave steepness decreases. Breaker type is closely associated with the expenditure of wave energy in the nearshore (e.g., reflection or dissipation) and the development of nearshore morphology. The morphology, in turn, controls breaker type. These relationships are the underlying bases for the concept of nearshore morphodynamics (see summary by Wright and Short 1984). The recognition of characteristic sets of dynamic relationships provides the basis for using morphodynamic regimes (or states) as a means for classifying beach types. For example, collapsing or surging breakers occur on reflective beaches. This contrasts with dissipative beaches (*q.v.*), where spilling breakers are common. Plunging waves tend to occur in the intermediate beach states of the morphodynamic model (i.e., systems where $20 \geq \varepsilon \geq 2.5$), where neither reflection nor dissipation dominates the nearshore energy response.

Reflective Beaches, Fig. 1 A reflective gravel beach near Malin Head, Co. Donegal, Ireland. Note steep foreshore and cusps. Maximum height of the collapsing breakers is less than 0.5 m, with a period of about 6 s



Characteristics of Reflective Beaches

In cross-section, morphodynamically reflective beaches display the classic form of “swell” or “summer” beach profiles (e.g., Sonu and Van Beek 1971). According to Wright and Short (1984), other distinguishing characteristics include steep nearshore and beach slopes ($\tan \beta$ between about 0.10 and 0.20), and, typically, relatively coarse sediment sizes. Coarse clastic beaches, therefore, tend to be reflective. The subaerial beach tends to be narrow with a pronounced step at the foot of the foreshore. Low-energy reflective beaches are approximately two-dimensional alongshore. Higher-energy systems frequently include well-developed beach cusps on the foreshore. These cusp systems are presumed to be caused by low-mode, subharmonic edge waves, although other periodic motions may form cusps. Incident wave energy is a maximum at or near the beach face. The classic reflective system displays only one coincident set of breakers (Fig. 1), and substantial energy remains at the landward extremity of uprush. On meso- and macrotidal beaches, the nearshore system may be reflective only at higher tidal stages and dissipative at low tide (Short 1991; Masselink and Hegge 1995). Short and Hesp (1982) have linked the reflective beach state to the formation of small foredunes that are eroded frequently. This linkage is a key concept in the development of beach-dune interaction models (e.g., Sherman and Bauer 1993). Coastal ecologists have linked beach states with species richness and abundance (Defeo and McLachlan 2011),

finding both to be minimized in reflective systems relative to comparable dissipative beaches.

Cross-References

- Bars
- Beach Features
- Beach Processes
- Dissipative Beaches
- Rhythmic Patterns
- Sandy Coasts
- Surf Zone Processes
- Waves

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Remote Sensing of Coastal Environments

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Coastal ecosystems are transitional environments that are sensitively balanced between open water and upland landscapes. Worldwide, they exhibit extreme variations in areal extent, spatial complexity, and temporal variability. Sustaining these ecosystems requires the ability to monitor their biophysical features and controlling processes at high spatial and temporal resolutions but within a holistic context. Remote sensing is the only tool that can economically measure these features and processes over large areas at appropriate resolutions. Consequently, it offers the only holistic approach to understanding the variable forces shaping the dynamic coastal landscape. Remote sensing must be able to adjust to these spatially and temporally changing conditions and also be able to discriminate subtle differences in these systems. As a result, remote sensing of coastal ecosystems is a complex undertaking that needs to incorporate not only the ability to define the observable hydrologic and vegetation features, but also the scale of measurement.

Historical Development

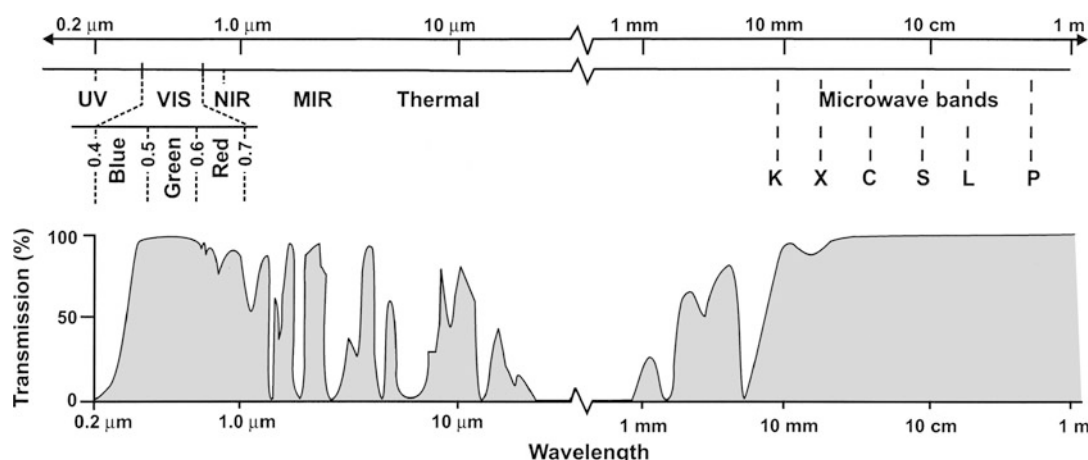
Since the 1960s, remote sensing has been used to describe a new field of information collection that includes aircraft and satellite platforms carrying cameras to electro-optical and antenna sensor systems (Jensen 2000). Up to that time, camera systems dominated image collection and photographic media dominated the storage of the spatially varying visible (VIS) and near-infrared (NIR) radiation spectral intensities reflected from the earth to aircraft platforms. Beginning in the 1960s, electronic sensor systems were increasingly used for

collection and storage of earth's reflected radiation, and satellites were posed as an alternative to aircraft platforms. Advances in electronic sensors and satellite platforms were accompanied by an increased interest and use of radiant energy not only from the VIS and NIR wavelength regions but also from the thermal and microwave regions. In 1983, the American Society of Photogrammetry and Remote Sensing adopted a formal definition of remote sensing as “the measurement or acquisition of information of some property of an object or phenomenon, by a recording device that is not in physical or intimate contact with the object or phenomenon under study” (Jensen 2000, p. 3). Although, others extended this definition to encompass the new technologies established for data collection, all definitions implicitly suggest that the property measured should describe a feature occupying a finite volume at a certain spatial and temporal position.

Remote sensing can include mapping of the earth's magnetic and gravitational fields and monitoring activities based on mechanical vibrations such as marine profiling by sonar and seismic exploration (Slater 1980; Horler and Barber 1981), or it can be applied to fields as diverse as cosmology to medical imaging. Historically, however, remote sensing has been used primarily to describe the collection of information transmitted in the form of electromagnetic radiant energy from an aircraft or satellite that is relevant to the earth's natural resources. Following this description, most remote sensing applications can be generally described by considering, (1) the nature of the probing signal, (2) the characteristics of the sensor and sensor platform, and (3) the interaction of the signal with the target.

Electromagnetic Spectrum

The electromagnetic spectrum describes the distribution of energy per wavelength (λ) or within an interval of consecutive wavelengths referred to as spectral bands ($\lambda\Delta$). Standardized to the speed of light in a vacuum, the electromagnetic spectrum is used to categorize general similarities of electromagnetic radiation in terms of changes in wavelength. The spectrum begins at the short wavelength cosmic rays and extends to the long wavelength radio waves (Fig. 1). Ultraviolet radiation below 0.3 μm is removed by atmospheric ozone absorption, and between 0.3 and 0.4 μm atmospheric scattering reduces image contrast to levels generally unacceptable for satellite remote sensing applications (Slater 1980). From about 0.4 μm to about 0.7 μm (VIS) little absorption occurs in clear and unpolluted atmosphere, although scattering is higher in this region than at longer wavelengths. In practice, the VIS is subdivided into the blue, green, and red regions. Above 0.7 μm , remote sensing applications are concentrated in atmospheric transmitting regions or windows bridging strong absorption bands primarily related to water vapor, ozone, and carbon dioxide. The region from 0.7 to 1.3 μm defines the NIR region, and the



Remote Sensing of Coastal Environments, Fig. 1 The upper figure shows the general locations and ranges of various wavelength regions. Note the microwave region is used by passive microwave (passive measurement of emitted energy) and radar (active) remote sensing systems.

The lower figure depicts atmospheric transmission. Satellite remote sensing applications are normally carried out in spectral regions of high transmission (shaded) and avoid regions of high blockage

combined VIS and NIR regions are commonly referred to as the VNIR. The middle infrared (MIR) region from 1.3 to 8.0 μm is sometimes partitioned into a shortwave infrared (SWIR) region from 1.3 to 2.5 μm dominated by reflectance and a region from 2.5 to 8.0 μm dominated by emission. The MIR contains high atmospheric water vapor and carbon dioxide absorption bands and includes an atmospheric window from about 3 to 4 μm and a region of strong absorption between 5 and 8 μm. Ozone absorption between 9 and 10 μm interrupts the thermal region extending from 8 to 14 μm. Poor atmospheric transmission and the lack of sensitive detectors and instrumentation prohibit remote sensing applications between around 14 μm and 1 mm. Starting at 1 mm and extending up to 1 m, the microwave region is subdivided into regions from the shortest (*K*) to the *K* longest (*P*) wavelength. Beyond the microwave region begins the region of television and radio frequencies. The electromagnetic spectrum does not end abruptly, but frequencies less than 3 kHz corresponding to wavelengths longer than 10^5 m are not used. Mechanical vibrations including sound and seismic waves begin at frequencies below 20 kHz. Marine profiling by sonar, the audio analog to radar, is in the 200-Hz range (Slater 1980).

Platforms and Sensors

Aircraft color infrared photography is useful in providing detailed biophysical, high-quality information about coastal ecosystems; unfortunately, turnaround for new map production is relatively slow. Satellite remote sensing provides holistic but detailed information on a regional as well as repetitive basis, and it is the only feasible approach to successfully overcome many intractable problems related to mapping

and monitoring of coastal ecosystems. Satellite remote sensors are increasing in number, in type, and in operational usefulness, and will provide the basis for integrated remote sensing applications (Lillesand and Kiefer 1994; Jensen 2000). Visible to thermal sensors have the longest history and have shown promise in mapping and monitoring coastal wetland type, health, biomass, and water quality. Since the late 1970s, microwave has gained importance in wetland mapping (Lewis et al. 1998). Microwave sensors extend the past capabilities of visible to thermal sensors in mapping coastal ecosystems by adding the potential for higher canopy penetration, more detailed canopy orientation and density information, and 24-hr-a-day collections nearly independent of weather conditions. Aircraft data will continue to provide calibration surveys, algorithm verification, testing of new sensor systems, specialized sensor collections, and in some cases local disaster response.

Remote sensing sensors measure radiant flux over bands defined by a range of wavelengths (e.g., VIS). A radiometer is a remote sensing instrument that measures radiant flux at any distance over a single band and a spectroradiometer measures over multiple bands. The specification of the instrument determines the spectral coverage, spectral, angular (or spatial or ground resolution), and radiometric resolutions. These factors clearly control the type and value of the data obtained.

Imaging and nonimaging sensors used in remote sensing use either electro-optical (VIS to thermal) or antenna detectors (microwave). Normally, nonimaging sensors are radiometers that collect accurate data over wide spectral regions where the spatial and spectral aspects are less important (Elachi 1987). The signals returned to the sensor represent the scene spatial characteristics as rows and columns of pixels (discrete picture elements) within an image. Resampling is

sometimes necessary to construct a continuous image representation of the scene (orbital and sensor characteristics), and almost always necessary to georeference the image to an earth coordinate system (e.g., latitude/longitude). Resampling tends to reduce confidence in the pixel location and blur the information contained in spatially adjacent pixels by adding a component of within-image spatial covariation. The pixel dimensions and location on the image represent an estimate of the spatial dimensions (spatially averaged ground area) and scene location that contributed to the pixel value.

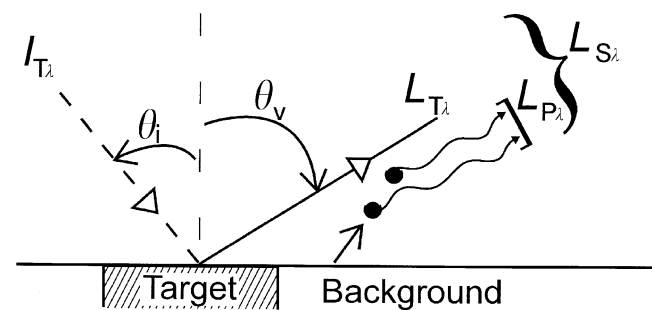
Sensors commonly used can be separated into two types, active and passive. Active sensor systems both transmit and receive reflected or scattered radiant flux. The light detection and ranging (LIDAR) and radio detection and ranging (RADAR) are common active sensor systems. These systems track the time difference between the transmission of the emitted pulses of energy and the arrival of the scattered return at the sensor. The distance to or range of the target is then directly obtained from the time difference. Active systems can also control the nature of the radiation used to probe the target. For instance, the wavelength placement, bandwidth ($\lambda\Delta$), polarization, and the angle of incidence can be controlled by the sensor and sensor platform. And because the energy source is part of the system, active systems can operate day and night and in the case of radar during most weather conditions. These features allow a greater control over the application of remote sensing techniques, and their use in monitoring earth's resources is increasing.

Passive remote sensing primarily uses the sun as the source of electromagnetic energy. Planck's blackbody law describes the sun's energy distribution with respect to frequency and wavelength. A perfect blackbody absorbs and re-emits all electromagnetic radiation impingement upon it, while a partial emitter (gray body, e.g., water) is spectrally similar but of lower amplitude, and a selective emitter (e.g., quartz) is spectrally selective compared to the general shape of a blackbody. The sun's emitted radiation closely approximates a blackbody at 6,000 K where energy emissions are mostly contained between 0.3 and 2.5 μm , peaking near 0.55 μm . Solar radiation transferred through the earth's atmosphere and scattered and reflected by its surface is generally referred to as *solar reflected radiant flux* (Slater 1980). In addition, solar heating of the earth's atmosphere and surface produces a secondary source of energy that has a distribution similar to radiation emitted from a blackbody at approximately 300 K. Energy emitted by the earth starts around 2.5 μm and peaks near 10 μm . Radiation emitted by the earth's atmosphere and surface is referred to as *self-emitted thermal radiant flux* (Slater 1980). Radiant flux below about 2.5 μm represents solar-reflected radiant flux while above 6.0 μm it represents self-emitted thermal radiant flux. Between 2.5 and 6.0 μm , the relative amounts of each flux depend on the target reflectance,

emissivity, temperature, and atmospheric transmittance. Lowered atmospheric transmittance between 2.5 and 5.5 μm normally results in the self-emitted flux dominating this region even when the surface has a high reflectance. Even though the self-emitted thermal radiant flux is very low at microwave wavelengths, the atmosphere is nearly transparent permitting successful application of passive microwave remote sensing.

Target Interactions

Surface irradiance ($I_{T\lambda}$) (e.g., solar in passive, instrument in active) interactions with earth's features can be partitioned following Kirchhoff's law (a restatement of the conservation of energy) as the proportion of $I_{T\lambda}$ reflected ($\rho(\lambda)$), transmitted ($\tau(\lambda)$) and absorbed ($\alpha(\lambda)$), that is, $\rho(\lambda) + \tau(\lambda) + \alpha(\lambda) = 1$. These interactions are commonly described as finite volume or surface averages of discrete elements, such as algal cells in water, leaves in a canopy, pebbles on the soil surface, and aerosol particles in the atmosphere. If the aggregate properties associated with these elements are independent of changes in $I_{T\lambda}$ and the method of measurement, they are referred to as inherent, the desired quantity to extract from the target radiance ($L_{T\lambda}$) (Bukata et al. 1995). Other definitions include the specific conditions, particularly the view (θ_v) and local incident (θ_i) angles (Fig. 2), within the definition of inherent optical properties. Under such measurement conditions, inherent optical properties whether describing discrete scatters or averages, reflect, transmit, and absorb the same fraction of $I_{T\lambda}$ unless the material's inherent properties change. Unless the sensor is within the target volume (transmittance), a remote sensing sensor only measures the net result of the reflectance, transmittance, and absorption summed over the



Remote Sensing of Coastal Environments, Fig. 2 $I_{T\lambda}$ depicts solar irradiance at a sun zenith angle of θ_i on a horizontal surface target, but the same depiction can refer to radiance from an active source at an incidence angle of θ_i (in this case θ_i and θ_v may be equal although the direction of the incident and reflected (backscattered) fluxes would be opposite). $L_{T\lambda}$ depicts the reflected or scattered radiance from the target at θ_v , the sensor view zenith angle. $L_{P\lambda}$ depicts nontarget radiance added to $L_{T\lambda}$ from atmospheric scattering (* being the scatter center) and from areas surrounding the target (background). $L_{S\lambda}$ is the radiance at the sensor

target (pixel) (including depth) and generalized to a single source and view (normally ϑ_v and ϑ_i) as r_λ . The resultant target reflectance (r_λ) represents the measured fraction of $I_{T\lambda}$ reflected ($I_{R\lambda}$) ($r_\lambda = I_{R\lambda}/I_{T\lambda}$, range=0–1). The source of r_λ (and ultimately $L_{T\lambda}$) is surface and volume scattering, partitioned based on the transmittance depth (Whitt et al. 1990).

VNIR and MIR

If the conditions of the measurement and $I_{T\lambda}$ are clearly specified, r_λ is related to the averaged target properties through the averaged inherent optical properties ($\rho(\lambda)$, $\tau(\lambda)$, and $\alpha(\lambda)$) throughout the surface or volume. However, the $L_{T\lambda}$ is commonly recorded but not irradiance ($I_{T\lambda}$), thus, a factor similar to r_λ and tied to $I_{T\lambda}$ by a geometric distribution related to $L_{T\lambda}$ is needed. In VNIR and MIR, the bidirectional (ϑ_v and ϑ_i , bistatic) reflectance distribution function ($BRDF_\lambda$) describes the fraction of $I_{T\lambda}$ reflected over a solid angle (Ω) at ϑ_v , or the surface distribution of $L_{T\lambda}$ ($BRDF_\lambda = L_{T\lambda}/I_{T\lambda}$). $BRDF_\lambda$ is a function of the incident (φ_i) and view (φ_v) azimuths, ϑ_i , ϑ_v , and Ω_s subtended by the source at a point on the surface, and ϑ_v subtended by the entrance pupil of the sensor at the surface, that is, $BRDF_\lambda(\vartheta_i, \vartheta_v, \varphi_v, \varphi_i, \Omega_s, \omega_v)$ (Jensen 2000).

Emitted Energy

In thermal radiometry, the target or source emissions depend on the target contact kinetic temperature (KT) and the emissivity ($\varepsilon(\lambda)$). The $\varepsilon(\lambda)$ of a target equals $1 - \rho(\lambda)$ (when $\varepsilon(\lambda) = \alpha(\lambda)$, all transmitted incident Φ_λ is absorbed) and is the ratio of the emission spectral characteristics to a blackbody at the same temperature. The $\varepsilon(\lambda)$ relates KT to the self-emitted Φ_λ ($T_R(\lambda)$) from the target (i.e., $T_R(\lambda) = \varepsilon(\lambda)^{1/4} \cdot KT$). Because of this relationship, materials with equal KT but different $\varepsilon(\lambda)$ s will have different $T_R(\lambda)$ s or in terms of the sensor, $L_{T\lambda}$ s. Passive microwave intensity is the product of $\varepsilon(\lambda)$ and KT and is usually reported as brightness temperature. The $\varepsilon(\lambda)$ determines the energy from the effective emitting layer that is transferred across the soil surface, and it is dominated by surface soil moisture, and soil moisture dampens thermal and microwave emissions. In nonvegetated areas (e.g., deserts, oceans), atmospheric influences (including clouds) pose a problem to retrieval of the $T_R(\lambda)$ and therefore KT based on thermal radiometry. In vegetated areas, the vegetation adds to and attenuates the soil emissions; thus, $T_R(\lambda)$ detected at the sensor as $L_{T\lambda}$ contains emitted information proportional to both the vegetation and the soil layer. In passive microwave, the single scattering albedo and the optical depth describe microwave interactions and emissions from the overlying vegetation layer. In most cases, the effects of the single scattering albedo appear small and can be incorporated into the optical depth or set to a constant value.

Radar

The scattering properties of discrete targets (in isolation) are described by the radar cross-section (RCS, m^{-2}). As in optical cross-sections, the RCS symbolizes the interaction cross-section or the target backscatter reflectivity, not the actual target area (Raney 1998). Flux reflected or scattered to the sensor is the product of the incident radiant flux and RCS normalized by propagation losses and the area of the receiving antenna (Zebker et al. 1990). In typical resource applications, the radar signal recorded per pixel is the coherent summation of reflected (backscattered) energy from all scatterers (relative to the wavelength) in the distributed or diffuse target back to the sensor (Massonnet and Feigl 1998). As such, the recorded flux depends on the target or pixel area. The backscatter coefficient ($\sigma^\circ(R, A)$ at a specific range (R) and azimuth (A) location) generated from the calibrated return and normalized by the target area (corrected for the local incidence angle) (where, $\sigma^\circ(R, A) \Delta R \cdot \Delta A / \sin \vartheta_i(R)$ is equivalent to RCS) represents the measured fraction of incident Φ_λ backscattered ($\Phi_{b\lambda}$) from the target, $\sigma^\circ = \Phi_{b\lambda} / \Phi_\lambda$ (Raney 1998). As in reporting BRDF (albeit at one angle, monostatic), σ° , at a specific wavelength, incident direction and polarization, is most closely related to the size, shape, orientation, and composition (primarily water content) properties of the diffuse target. Changes in σ° reflect the variability of these diffuse targets to send the incident energy back to the sensor (Massonnet and Feigl 1998). A positive $\log(\sigma^\circ)$ in decibels implies focusing energy toward the sensor, while a negative number implies focusing energy away from the sensor (Elachi 1987).

Synthetic aperture processing (focusing) creating synthetic aperture radar (SAR) (NASA 1989) is used to improve the radar's spatial resolution. Within the processing, signal return variability is related to successive observations of the same area but from slightly different positions and somewhat different fine details in neighboring pixels with the same RCS (grainy appearance of image) (Elachi 1987). Increasing the number of looks (statistical averages of the radar returns) reduces these effects but decreases the effective spatial resolution. Contrary to the total solar irradiance, polarimetric radar is capable of synthesizing well-defined polarization states represented in the linear case by horizontal and vertical orientations. After standard processing and image construction, phase information related to the distance between the sensor and the target can be linked to the polarimetric return (Elachi 1987; Zebker et al. 1990). An interferogram is the resulting phase difference between two SAR images collected either from two antennae (bistatic) from a single platform or one antenna (monostatic) at two different times (Massonnet and Feigl 1998). In the latter case, the direction of observation and wavelengths must be identical, and in practice, the input images are collected from the same satellite in the same orbital configuration and

focusing or synthetic aperture processing (SAR) of the original image data are identical.

Both imaging and non-imaging radar sensors are commonly used in today's remote sensing applications. Altimeters use radar's ranging capabilities to measure the surface topography profile to centimeter-level precision at relatively high pixel spatial resolutions. This level of precision requires precise measurement of time and information extracted from the shape and slope of the returned pulse. Scatterometers assess the average scattering properties over large areas and within narrow spectral bands and provide directional capability by including more than one antenna (Elachi 1987). They provide high precision backscatter measurements that cover large areas but at low pixel spatial resolutions (Cavanie and Gohin 1995). Imaging SAR sensors are most often used for resource mapping because of their ability to provide high pixel spatial resolution data at multiple wavelengths, polarizations, and incident angles.

In all cases, objects of comparable size to the microwave wavelengths (mainly 2–30 cm in satellite imaging of earth's resources) most strongly influence the microwave scatter. At a constant incidence angle, canopy components such as leaves and stems normally interact with microwave wavelengths from about 2 to 6 cm, and trunks and limbs at longer wavelengths (10–30 cm). Longer wavelengths (>30 cm) provide more information about the surface properties but little about the canopy volume. Use of cross polarizations (e.g., horizontal send and vertical return, HV) and higher incidence angles, however, enhance volume scattering relative to lower (more vertical) incidence angles and like polarizations (e.g., VV, HH).

Applications of Remote Sensing in Coastal Ecosystems

Polarization, angles of incidence and view, proportions of direct and scattered irradiance, and surface roughness all work to modify and build a directional reflectance character that becomes the target radiance ($L_{T\lambda}$). $L_{T\lambda}$ can be further altered from the surface to the satellite ($L_{S\lambda}$) by attenuation and addition of path radiance ($L_{P\lambda}$) (Fig. 2), especially in the VIS but also in the NIR, MIR, and thermal regions and further modulated from the sensor to its representation on the image. In short, uncovering the relationship between $L_{S\lambda}$ output as a pixel in a grid-based image representation and the inherent properties of the target is often highly complex, and many times may be impossible to fully determine. Accountability can be built into the analysis by linking the image data to site-specific measurements through physical-based models. Greater accountability can diminish the reliance on gathering site-specific data and ultimately advance the operational and

accurate representation of the temporal and spatial distributions of biophysical features in the scene.

The generation of the bidirectional reflectance distribution function (BRDF $_{\lambda}$) or radar backscatter coefficient (σ°) is required when target inherent properties are sought or when the spectral contrast between the target ($L_{T\lambda}$) and its surroundings at the sensor prevents successful classification of the target. Classifications can be improved by generation of BRDF $_{\lambda}$ or σ° ; for example, biomass estimates are improved when generated from $L_{T\lambda}$ (top of canopy after atmospheric correction) versus $L_{S\lambda}$ (top of the atmosphere). Inferential relationships have more promise of extension over space and over time and the detection and mapping of subtle scene features when based on BRDF $_{\lambda}$ or σ° . The consistent success of these detection and monitoring methods based on remote sensing data requires a rudimentary estimation of the relative extent the target material reflects, transmits, and absorbs surface irradiance ($I_{T\lambda}$) under various geometric and $I_{T\lambda}$ conditions.

Classification

Most often, classification has relied on the differential interactions and responses of VNIR and MIR (e.g., spectral signature or $L_{T\lambda}$) to changes within the coastal scene to provide spectral classes uniquely linked to the type and state of coastal features. These classifications generally use simple image-based parametric statistical models (e.g., clustering techniques, principal component analysis, canonical correlation, discriminant analysis) that do not require detailed information about the plant, canopy, or water optical properties. The developed relationships are commonly limited to conditions existent only during the data collection, and are not necessarily extendable temporally or spatially. Integrating data from multiple remote sensors (VIS to microwave, hybrid models) has successfully improved the spatial detail of coastal vegetation classifications (Ramsey et al. 1998), as have newer classifications based on nonparametric statistical models (e.g., neural network analysis) that allow greater control in linking the classification to the biophysical characteristics.

Classification of the radar images can follow the commonly used procedures of VNIR and MIR classifications (Ramsey et al. 1998), while also enhancing the use of neighborhood information or image texture in the classification. Radar image texture is a combination of the system and scene features. System texture (speckle) can be diminished by increasing the number of looks or can be estimated and removed by averaging the radar return over undisturbed water bodies. The improved or corrected measure is a more accurate representation of scene texture and a more pertinent input into the classification process. A more direct method is developed on rule-based logic (van Zyl 1989; Dobson et al. 1995). In one method, returns associated with single or

combinations of SAR sensors of different wavelengths and polarizations from sites with known structural characteristics are used to generate a progressive classification hierarchy. In another, rules based on predicted polarimetric return signatures from different features are used to classify targets within the scene.

Biophysical Features

Vegetation type classifications are not always the primary objective of the remote sensing application. Often the objective is to directly link the remote sensing data to biophysical variables that describe the coastal ecosystem. Of the biophysical variables, leaf area index (LAI) is probably the most sought, and it can be related to wetted or total biomass and in certain instances primary productivity and even CO₂ exchange.

In the VNIR, the ability to map canopy LAI and productivity is based on two facts (Horler and Barber 1981; Smith and Morgan 1981). First, although biomass and yield depend ultimately on light absorption, the usually close correlation between leaf reflectance and transmittance provides a basis for remote sensing reflectance measurements to be successfully applied in agronomic applications. Second, there is a striking attenuation difference between the red and NIR. NIR is not significantly absorbed and is nearly equivalent to above-canopy flux while the red is strongly absorbed and is therefore nearly equivalent to flux transmitted between the canopy leaves. Thus, the degree of canopy shading is related to both the LAI and the differential interactions of red and NIR as described numerically by a vegetation index (VI). In the VNIR, two types of VI are commonly used to transform remote sensing data into estimates of LAI: those based on ratio transforms and those based on orthogonal transforms. These VIs can be modified and possibly improved by altering the required input bands, or by precorrecting the image data to account for atmospheric influence, but variability remains related to the canopy BRDF_λ, or primarily to canopy structure.

Because the three-dimensional (3-D) distribution of water within the canopy has the greatest influence on microwave interactions, and because wetted biomass is related to vegetation water content, passive microwave and radar remote sensing are sensitive to LAI, and in turn, biomass variations. Similar to VNIR to MIR ratios and differences, biophysical variables such as LAI can be related to radar copolarized (e.g., HH, VV) and cross-polarized (e.g., VH) ratios and differences (Wegmuller and Werner 1997). The vegetation biomass distribution (canopy structure) and quantity (water content) are also related to canopy optical depth that in turn is related to emission variability at microwave wavelengths (Wigneron et al. 1995).

Vertical Canopy Profiling

Vertical canopy profiling is an indirect result of using multi-spectral remote sensing systems. Canopy profiling can be used to estimate canopy architecture, an indicator of species variety, phenological stage, and present and past vigor (Malet 1996). In general, the longer wavelengths transmit further into fully formed vegetation canopies relative to shorter wavelengths. Canopy structure, or the spatial distribution and orientation of the canopy elements, however, also influences canopy penetration and must be removed or accounted for before the variable return can be used to describe the canopy architecture. Typically, NIR to MIR wavelengths transmit from 8 to 10 leaf layers (equivalent LAIs) into the canopy and VIS wavelengths from 2 to 3 LAIs. Although there are notable exceptions, active sensors offer a greater ability to profile the response from various depths within the vegetation canopy as compared to passive sensors. The use of shorter to longer radar wavelengths, multiple incident angles probing with a single band, and multiple polarizations can offer variable depths of transmittance although the analogy becomes less straightforward in canopies with convoluted branching (Ramsey 1998). LIDAR offers the most direct canopy profiling. By using the allometric relationship between tree height and diameter-at-breast-height, LIDAR can provide volumetric representation of the canopy structure.

Vegetation Stress

One of the greatest challenges to coastal remote sensing is detecting plant stress before irreversible losses occur due to changes in inundation, flushing water salinity, and other external forces as a result of sea-level rise and shoreline alteration and protection. Although broadband VNIR to MIR remote sensing applications have detected broad indicators of vegetation stress, hyperspectral systems have identified specific spectral features related to stress from metal contamination to deficient foliar water content (Card et al. 1988). Radar is sensitive to vegetation stress through changes in water content, and because optical depth is linked to water content, passive microwave and thermal radiometry are closely related to plant stress. Chlorophyll *a* fluorescence can also be used to assess vegetation stress by providing estimates of photosynthetic capacity (Carter et al. 1996). A passive technique using Fraunhofer line radiometers (FLR) detects the absorbed photosynthetic radiant flux (about 3%) re-emitted as fluorescence by taking advantage of the relatively strong Fraunhofer absorption lines in the solar irradiance (Horler and Barber 1981; Carter et al. 1996). One of the strongest Fraunhofer lines is located in the chlorophyll fluorescence peak providing a convenient method for identifying plants suffering from metal toxicity and water stress. Active laser-induced fluorescence (LIF) sensors also provide the ability to assess the fluorescent properties of the leaf pigments (e.g., chlorophyll). Both

passive FLR and active LIF offer new capabilities to isolate the alteration or change in dominance of specific pigments as an early indicator of vegetation stress.

Thermal Radiometry

In dry environments (e.g., nearly constant emissivity (ϵ_λ)), the rate of temperature change in response to variable surface solar irradiance (I_T) can be used to uniquely identify the target material (e.g., soil composition, mineral) (Lillesand and Kiefer 1994). To characterize the rate materials respond to temperature changes (thermal inertia), radiant temperature ($T_R(\lambda)$) measurements are collected in the early morning and the afternoon. This temperature difference indicates the variable heat capacity of the different materials and can be used to map landcover variation. Heat capacity mapping has been applied mostly in nonvegetated regions for identifying geologic materials. Intense heat sources, however, can often be directly observed. Fires can exhibit self-emitted thermal fluxes down to about 3.0 μm , and some volcanic lava flows as low as the NIR (0.7–1.3 μm).

Area Mixtures

In any remote sensing application, all pixels are weighted mixtures of different scene features, even when the pixel nearly matches the target feature's mean spatial extent (e.g., adjacency effects, boundary pixel landcover mixtures). Mixture models are used to extract the occurrence of specific scene features from composite mixtures (e.g., trees, water, bare ground). In a linear mixture model, weighted combinations of specific scene features (e.g., spectral endmembers) are combined to completely reconstruct every spectral signature ($L_{S\lambda}$ or $L_{T\lambda}$) as represented on the image. This reconstruction allows target features to be detected and the percent occurrence in each image pixel to be determined. In nonlinear endmember analysis, the interaction between target features is included. Mixture models can be based on broadband sensor data; but most successful applications are based on hyperspectral sensor data (Adams et al. 1986).

Soil Moisture Content

The NIR to MIR regions are used to estimate soil moisture where the increase in soil moisture (as in the presence of standing water) dampens the return to the sensor. Successful studies have relied more on direct determinations with thermal radiometry and passive microwave and radar (Idso et al. 1978; Ulaby et al. 1983; Shutko 1992; Kostov and Jackson 1993; Chanzy et al. 1995; van de Griend et al. 1996). As noted, decreases in moisture content are associated normally with decreases in the radiant temperature ($T_R(\lambda)$). In the thermal region, emissions are constrained to within 50 μm of the surface and this shallow depth can exhibit rapid temperature variations. Even though monitoring diurnal temperature variations in the thermal region may improve the

moisture content estimation, microwave is the only remote sensing platform and technique that can provide soil moisture with reasonable precision and consistency.

Both radar and passive microwave sensors are sensitive to non-bound water, and in general, sandy soils hold less bound water than clays. Normally, microwave is sensitive to soil moisture content within the top 2–5 cm of the soil depth (Chanzy et al. 1995). In exceptional conditions, radar can detect changes in moisture content at depths greater than 1 m, and passive microwave returns have been related to moisture content and groundwater at depths exceeding 1 m (Reutov and Shutko 1992). Up to saturation, soil moisture acts to enhance the radar return at any given soil surface roughness height; in some cases, surface roughness variability can severely hamper the ability of radar to estimate changes in soil moisture (Ramsey 1998). Radar is sensitive to soil moisture under short vegetation canopies; however, where moderately dense, the detection of soil moisture depends on the relative strength of the vegetation canopy and the incident flux interaction (Dobson et al. 1995). Surface roughness increases also enhance passive microwave emissions and can enhance soil emissivity differences between wavelengths. For most natural surfaces, however, roughness is not a serious limitation, and in wet soils, emissivity differences may be minor (Wang et al. 1987; Engman and Chauhan 1995). In most soils, passive microwave emissions are practically independent of soil type, salinity, bulk density, and temperature variability (Shutko 1992; Engman and Chauhan 1995), but overlying vegetation attenuation increases as water content increases (van de Griend et al. 1996). In both passive microwave and radar, in general, the transmission through the vegetation canopy increases with increasing wavelength. Thus, longer wavelengths, HH polarization, and steeper incident angles are preferred for sensing soil moisture through a vegetation canopy.

Shoreline Placement

The delineation of land and water is important in coastal classifications, land loss, and shoreline displacement. Accurate mapping of coastal shoreline (defined as the high water line or wet-dry boundary) placement requires not only high spatial resolution sensors but also the spectral ability to provide contrast between open water and the regional nearshore material. Historically, optical sensors (especially photographic) have provided image data used for shoreline mapping, but more recently, radar image data have been used to construct shorelines (Lee 1990; Ramsey 1995). Coastline detection and automated tracing algorithms are being developed to provide dynamic shoreline construction. In the case of radar, scatter from roughened open water at times can limit the ability to differentiate land and water areas. In addition, a shoreline position is dynamic, especially in coastal regions experiencing high tidal ranges. Consequently, to truly

represent the shoreline position, the variation in the tidal height relative to mean high tide and the occurrence of influencing forces such as wind set-up or set-down or abnormal river runoff must be accounted for at the time of the measurement.

Flood Monitoring

Remote sensing can detect flooding under vegetation. VNIR to MIR have been used, but microwave remote sensing offers the greatest potential for the instantaneous and consistent determination of flood extent. Within the microwave region, the most extensive history of flood detection under vegetation has been associated with radar (Ramsey 1998). The radar return from flooded forest is usually enhanced compared to returns from nonflooded forests. The enhancement is related to the double bounce mechanism where the signal penetrating the canopy is reflected off the water surface and subsequently reflected back toward the sensor by a second reflection off a tree trunk (Hess et al. 1990). In contrast, diminished returns from flooded relative to nonflooded coastal marshes have been observed (e.g., Ramsey 1995). The marsh grasses may calm the water surface accentuating specular reflection but without the grasses providing the double bounce (Ormsby et al. 1985). As in soil moisture mapping, flood detection can occur only if transmitted through the canopy; thus, longer wavelengths, HH polarization, and steeper incident angles are preferred.

Topography

Coastal topography controls the hydrology of the coastal wetlands, and thereby the distribution and health of the coastal vegetation. Offshore coastal bathymetry is a result of the dynamic forces of local erosion and sedimentation and littoral drift. Mapping and monitoring the onshore topography and offshore bathymetry is of vital importance to the coastal engineer and resource manager. Historically and currently, indirect methods based on the simultaneous viewing of overlapping images (parallax) are commonly used to generate topographic information. Methods based on passive optical remote sensing also have been used to map coastal bathymetry, but these methods are limited by severe and variable attenuation by the water column materials (Ji et al. 1992; Lyon et al. 1992). Besides audio-mechanical systems (sonar), radar and LIDAR systems offer a direct and more consistent approach to surveying coastal topography and bathymetry.

After processing and most orbital contributions have been eliminated from the interferogram (radar phase difference image), slight remaining differences in the point of view of the radar sensor yield fringes that follow the topography (Massonnet and Feigl 1998). These topographic contours can be used to generate a digital elevation model (DEM). Additionally corrected for elevation and local elevation

gradient (slope) spatial variances, the interferogram can be related to finer resolution topographic changes from deformation (e.g., coastal volcanoes, surface subsidence, erosion, rebound, or deposition). In monostatic systems, success of this technique requires all scatters comprising the target (e.g., overlying vegetation, moisture content, inundation) remain unchanged between the time of the two radar image collections (Massonnet and Feigl 1998). While successful application is problematic in vegetated environments, absolute stability, and thereby success, in dynamic and highly vegetated coastal environments is less probable. Frequent and variable flooding and the associated changes in soil and vegetation moisture contents add complexities that appear as random speckles in the interferogram. In coastal areas, these complexities may limit the absolute elevation and elevation change resolutions attainable with interferograms generated from monostatic systems.

Airborne laser altimetry (ALS) is the simplest application of LIDAR remote sensing. ALS surveys are primarily performed at 700–1,000 m above ground level in order to eliminate most atmospheric attenuation of the signal. As in radar systems, when properly calibrated to a stable platform, the time between the emitted and detected pulses is directly related to the range, and thereby to changes in the surface elevation. Vegetation interferes with the laser pulse and complicates conversion of the ALS image into a topographic surface. Reflectance of solar illumination into the sensor field of view within the laser operational bandwidth also corrupts the ALS signal. A correction for vegetation interference and contamination uses data collected near in time along multiple transects to develop a topographic precision estimate.

Bathymetry

ALS is also used to develop coastal bathymetry maps. Most current ALS systems use two wavelengths: a green band for high water penetration to the bottom and a NIR band with little to no water penetration and almost total reflectance from the surface. Use of the two-band system helps diminish errors resulting from platform altitude variation so that the time delay difference between the two return pulses is a direct bathymetric measure. Increased turbidity, however, results in lower spatial resolutions and increased water volume backscattering of the emitted pulse creates false echoes in the record. The low altitude and fairly narrow coverage of current ALS systems hamper the operational feasibility of these systems in regional assessments. Even with these limitations, of all the electromagnetic systems, the ALS systems may provide the only feasible mechanism for rapid and consistent detailed mapping of coastal bathymetry and wetland topography.

Water Quality and Submerged Aquatic Vegetation

Remote sensing of estuarine and coastal waters is primarily concerned with mapping the type, concentration, distribution,

and dispersion of materials suspended and dissolved in the water (water quality) (Morel and Gordon 1980; Ramsey and Jensen 1990), and the type and distribution of submerged aquatic vegetation and bottom cover (e.g., rock, mud, sand, shell). To accurately map the water quality and bottom type, the radiant flux depth-intensity and spectral distribution must be estimated (Kirk 1980). In a well-mixed water column, the underwater radiation environment is determined by reflection and refraction at the water-air interface (surface), absorption and scattering within the water body, and reflection from the bottom. Surface reflections of $I_{T\lambda}$, upwelling restrictions (due to refraction at the surface), added atmospheric path and background fluxes, and atmospheric attenuation result in the upwelling water volume flux (below the water surface) transferred through the surface ($L_{T\lambda}$) normally comprising only 3–5% of the sensor signal ($L_{S\lambda}$). Corrected for water surface reflection, transmission, and atmospheric influences, $L_{T\lambda}$ can be related to the inherent bulk absorption and scattering properties of the water and water materials and bottom reflectance in the shallow waters (Morel and Gordon 1980; Carder and Steward 1985).

Excluding bottom reflections, the inherent bulk absorption ($\alpha(\lambda)$) and backscatter ($b_b(\lambda)$) coefficients can be related to $L_{T\lambda}$ as $L_{T\lambda}/I_{T\lambda} = BRDF_{\lambda} = C_F \cdot b_b(\lambda)/\alpha(\lambda)$, where backscattering is scattering into the hemisphere trailing the incident flux, and C_F incorporates the ratio of two subsurface upwelling fluxes and the water refractive index (Carder and Steward 1985; Bukata et al. 1995). In optically complex coastal waters, bulk $b_b(\lambda)$ and $\alpha(\lambda)$ are commonly related to bulk descriptors of biomass (e.g., chlorophyll-a [Chl]), suspended materials (SMs) (e.g., detritus, suspended inorganic particles), and dissolved organic carbon (DOC) (Bukata et al. 1995). Use of these bulk descriptors normally provides a good and stable estimation of water quality with respect to location and time in coastal environments. The inherent bulk properties are related to the specific optical coefficients (i as $b_b(\lambda) = \sum C_i (b_b)_i(\lambda)$ and $\alpha(\lambda) = \sum C_i \alpha_i(\lambda)$, where the subscript (i) represents one water component and C_i refers to the components concentration (e.g., $\alpha(\lambda) = a_w + C_{Chl}\alpha_{Chl}(\lambda) + C_{SM}\alpha_{SM}(\lambda) + C_{DOC}\alpha_{DOC}(\lambda)$, where a_w is absorption due to the water).

LIDAR systems can be used to stimulate fluorescence in chlorophyll pigments associated with phytoplankton, plants and corals, and fluorescent DOC (i.e., Gelbstoff) (Measures 1984). As in passive optical remote sensing of water quality, the signal returned to the sensor can be corrupted by the atmosphere and by addition of reflectance from the bottom. The selection of the excitation (send) and emission (return) wavelengths is either fixed by the LIDAR system or optimized by laboratory measurements. Chlorophyll is normally determined by excitation in the low red and measuring the emission in the high red wavelength regions. Gelbstoff is linearly related to natural fluorescence (Otto 1967) that has

been used as a conservative tracer of riverine and ocean waters mixing. Natural fluorescence is normally determined by excitation in the ultraviolet and by measuring the fluorescence in the blue. As in canopy profiling, the fluorescent return can be scattered or self-absorbed before reaching the water surface, causing ambiguity in mapping the concentration of the fluorescent material. Raman intensity variability can be measured simultaneously with the fluorescent return and used to remove the effect of self-absorption, thereby creating spatially comparable fluorescent images (Bristow et al. 1981).

Bottom Reflectance

Seagrasses and bottom type (mud, sand, shell) mapping is an important aspect of coastal monitoring. In this case, the overlying water column attenuates the signal to and from the bottom. In both passive VNIR and LIDAR mapping, the same problems apply as in water quality monitoring; however, in this case, the overlying water column signal must be removed from or de-emphasized in the return signal (Ji et al. 1992; Lyon et al. 1992). Increasing water turbidity and absorption can severely restrict the ability of either method (passive VNIR and LIDAR) to accurately map bottom reflectance variations as does low contrast between the different bottom materials.

Surface Films and Salinity

Observation of surface water features, especially surface films, has long been recognized as an indirect method of mapping convergence zones, mixing zones, and internal wave fields in optical oceanography (Klema 1980). The natural and extracted oils and similar substances also can be observed by SAR systems because they tend to dampen the creation of surface waves, smoothing the water surface, and attenuating the SAR returns. This differential dampening enables SAR sensors to map and monitor surface spills, and because of its nearly all weather capabilities, SAR provides capabilities many times superior to VNIR and MIR. Laser Induced Fluorescence (ultraviolet excitation, visible emission) has also been used to detect and classify oil slicks and oil film thickness (Measures 1984).

Of the possible conservative tracers of water mass mixing, salinity is the most notable. Salinity is not directly measurable with VIS to thermal systems (ignoring extremely slight dependencies), although salinity changes have been inferred from changes in other water properties, such as fluorescence and suspended particle concentrations. A more direct measure is based on the definition of salinity as the concentration of dissolved cations and anions. Changes in these concentrations change the water's ionic

strength, leading to changes in the dielectric properties of the water that are most apparent at microwave wavelengths. These changes are best observed as changes in the emissivity, although to accurately observed changes, microwave emissions must be corrected for water temperature and surface roughness variations. To accomplish this correction, microwave measurements are collected at two wavelengths, obtaining a direct method to map changes in water salinity (Shutko 1985).

Sea Surface Temperature

One of the earliest uses of radiometry was to map the sea surface temperature (SST) and thereby map different water masses and physical dynamics (e.g., frontal convergence, upwelling). Along the same line, water temperature mapping of heated effluents into rivers, estuaries, and coastal oceans is used to monitor compliance of discharges. In each of these radiometric applications, the radiant temperature ($T_{R\lambda}$) is related to the 3–5 μm thick surface skin by converting via emissivity ($\epsilon(\lambda, \text{water})$) to the skin kinetic temperature (KT). On average, surface skin temperatures are about 0.3 K cooler than bulk temperatures, but differences can range from about +1 to –1 K (Emery et al. 1995). Depending on the water stability or the amount of mixing, the bulk temperature can represent the well-mixed surface layer or the temperature gradient depth. Often the skin $T_{R\lambda}$ (radiometer measurements) is related to the bulk KT or SST by breaking the surface skin with buckets of water (i.e., bucket temperature). Ship intakes and moored buoys offer bulk temperature measurements at variable depths. These measurements are used to correct atmospheric influences (especially water vapor) and consequently directly relate the sensor signal ($L_{S\lambda}$) to the SST, aggregating skin effects into atmospheric correction (Minnett 1995). A separate type of atmospheric correction relates $L_{S\lambda}$ to atmospheric attenuation and thereby to the skin KT. Atmospheric effects are inferred from spectral relationships based on multiple band measurements (e.g., $\text{SST} = a_0 + a_\lambda \text{KT}_\lambda$, where λ refers to one or a combination of bands) (Emery et al. 1995; Minnett 1995). Alternatively, measurements of the same target but at different sensor view angles provide a direct measurement of atmospheric influences ($\text{SST} = b_0 + b_{\lambda, N} \text{KT}_N + b_{\lambda, S} \text{KT}_S$, SKTS, where N is nadir, S is oblique views, and λ refers to one band or a band combination). Based on current correction techniques, skin and bulk surface water temperatures can be estimated to less than 0.5 °C. Even though less influenced by atmospheric conditions, the coarse spatial resolution associated with passive microwave systems makes them less preferred than thermal radiometry. Further, even at longer wavelengths, passive microwave measurements have shown dependence on surface roughness as a function of wind speed (Shutko 1985; Trokhimovski et al. 1995).

Sea Ice

Detection and monitoring the distribution and type (first year and multiyear ice) of coastal (Arctic) sea ice is important in marine mammal ecology, climate processes, and early detection of global warming (Piwowar and LeDrew 1996). Operational methods until recently have been applicable only to broadscale spatial inventories. More recently, optical, passive microwave, and radar sensors have provided higher spatial resolutions; however, increasing spatial resolution beyond 1 km constrains the ability to operationally monitor global or hemispheric regions. The primary factor controlling the remote sensing of sea ice is emissivity ($\epsilon(\lambda)$), and the major factor controlling emissivity is salinity (Comiso 1995). In thermal radiometry and passive microwave, the effective emissivity decreases from the cold saline first-year to the cold desalinated multiyear ice. Emissivity also decreases from first-year to multiyear ice due to the decreased density and increased surface roughness resulting in increased surface scatter. Added to this overall change, emissivity varies with wavelength and polarization. First-year ice is nearly independent of wavelength and polarization while emissivities of both VV and HH polarizations associated with multiyear ice increase with wavelength (Comiso 1995). SAR returns tend to increase from first-year to multiyear ice (Drinkwater 1995). In radar imaging, shorter wavelengths scattered from the ice surface primarily respond to dielectric differences (salinity) and roughness, while longer wavelengths (>5 cm) penetrate the ice (multiyear >> first-year) and are returned via volume scattering. Use of longer wavelengths normally results in a higher return from the relatively lower salinity and density multiyear ice than other ice types (Drinkwater 1995). Of the three sensors, optical sensors are constrained by persistent clouds and darkness in Arctic regions, and radar returns from roughened multiyear ice surfaces are more prone to confusion with other types of sea ice than are emissions sensed by passive microwave sensors (Hall 1998). Integrated approaches seem to provide the best results and the added benefit of comparison and validation.

Wind Speed and Surface Waves

Short gravity (>1.7 cm) and small capillary (<1.7 cm) waves are the surface water features observed with operational radar systems (2–30 cm wavelengths) (Elachi 1987). Increasing near-surface wind speed increases the amplitude of these waves, intensifying surface roughness that in turn promotes increasing slope (specular or facet) reflections and point scatter (or Bragg). As wind speed is related to surface roughness, altimeter and scatterometer sensors can provide estimates of wind speed (U). Scatterometers provide regional coverage, but at coarse pixel resolutions, while altimeter measurements cover narrow swaths. In the case of altimeters, σ_0 is a result of near nadir specular reflections (facet) that decrease as surface roughness increases ($\sigma^\circ \propto U^{-x}$) (Elachi 1987;

Dobson 1995). Although designed for measuring open ocean winds, scatterometers have also been found useful in measuring winds in coastal and enclosed seas. σ_0 derived from scatterometer measurements increases with wind speed increases (at $>25^\circ$ incident angle) as $\sigma_0 \propto U^x$ (x is dependent on wavelength) and are principally a result of Bragg scattering (Topliss and Guymer 1995). Scatterometers also include multiple azimuths, providing the ability to estimate the wind direction. Wind speed accuracies derived from scatterometers are about 2 m/s with a directional tolerance of 20° (Topliss and Guymer 1995). Compared to open ocean measurements, altimeter and scatterometer coastal measurements are more difficult to explain based on dynamic processes and are generally plagued by three types of problems: (1) contamination from land-water mixing, (2) varying wind-radar relationships, and (3) substantial influences of temporal and spatial variations in SST resulting in incorrect estimates of σ_0 .

Currents and Waves

Indirect observation of convergence and divergence zones associated with currents and possibly internal waves through varying surface features is used in optical remote sensing of ocean features (Klemas 1980). Radar systems, however, offer a nearly unimpeded source of mapping surface features related to ocean dynamics. Short gravity and capillary waves created by wind stress and mechanically (independent of wind stress) are modified (local slope and growth) by long-period gravity or internal waves and variable currents (fronts, eddies, upwelling, tidal circulation) and bottom topography (Topliss and Guymer 1995). These spatially and periodic modulated small wave fields (bands of roughness) are detectable with altimeters, scatterometers, and SAR imaging. Dependent on the angle of incidence and wavelength, SAR returns can be dominated by either a mixture of Bragg scattering or specular reflections. SAR images are used to define wave direction and length and to map the location of convergence zones and currents.

Water Surface Topography

Instantaneous sea surface height (S_0) observed by an altimeter is the sum of the geoid (N , a level surface of the earth's gravity field associated N with a motionless ocean surface regarded as time invariant), the permanent dynamic topography (ζ_0 , related to ocean circulation, $N+\zeta_0$ =mean sea level), the variable topography (ζ_v , e.g., ocean tides and waves and swells), orbital and propagation errors (e.g., sensor attitude corrections, barometric correction), and sensor noise ($S_0=N+\zeta_0+\zeta_v$ +error+noise) (Le Traon 1995). Wave heights from altimeter measurements are related to the shape or rise time of the returned pulse (Dobson and Monaldo 1995). Increasing wave heights increase the slope or rise time. Large surface-wave heights or swell heights are estimated by differencing

wave energy ($\propto(\text{wave height})^2$) and wave energy associated with wave heights estimated from altimeter wind speeds. In principle, skewness of the generated wave height probability distribution can also be used to estimate the dominant wavelength in unimodal seas (Dobson and Monaldo 1995).

The low-frequency harmonic rise and fall of the coastal tides can be observed with satellite altimetry with a precision of about 3–5 cm (Han 1995). Limited sampling frequency associated with altimeter data, however, leads to aliasing shorter tidal periods into longer tidal periods causing ambiguities in tidal period evaluations. Conversely, extraction of tidal fluctuations from the surface height variability is necessary to recover long-term height variability (e.g., annual cycles). As in other coastal applications, increased problems are created by the high temporal and spatial variability in the surface height due to basin morphology (shape and shallow depths (≈ 100 m)), variable river runoff (buoyancy), solar heating (e.g., SST), wind stress (e.g., Ekman drift), and the surface expression of subsurface features (e.g., sea mounds or submarines).

Cross-References

- [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
- [Coasts, Coastlines, Shores, and Shorelines](#)
- [Mangroves, Remote Sensing](#)
- [Nearshore Geomorphological Mapping](#)
- [Photogrammetry](#)
- [Remote Sensing: Wetlands Classification](#)
- [Shoreline and Coastal Terrain Mapping](#)
- [Synthetic Aperture Radar Systems](#)

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Remote Sensing: Wetlands Classification

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Coastal wetlands are a highly productive and critical habitat for a number of plants, fish, shellfish, waterfowl, and other wildlife. Wetlands also provide flood damage protection, protection from storm and wave damage, water quality improvement through filtering of agricultural and industrial waste, and recharge of aquifers. After years of degradation due to dredge and fill operations, impoundments, urban development subsidence/erosion, toxic pollutants, eutrophication, and sea-level rise, wetlands have finally begun to receive public attention and protection (Daiber 1986). Heightened awareness of the value of wetlands has resulted in the need to better understand their function and importance and find ways to manage them more effectively. To accomplish this, at least two types of data are required: (1) information on the present distribution and abundance of wetlands; and (2) information on the trends of wetland losses and gains.

Coastal wetlands can be conveniently divided into four major types: salt marshes, coastal fresh marshes, coastal forested and scrub-shrub wetlands, and tidal flats (Field et al. 1991). Each of these wetland types has different hydrologic requirements and is dominated by a different type of vegetative cover. As a result, their spectral signatures and detectability by remote sensors differ significantly. For instance, salt marshes are widely distributed along the Atlantic and Gulf coasts and are dominated by smooth cordgrass (*Spartina alterniflora*) and frequently include other grasses, such as salt hay (*Spartina patens*) and big cordgrass (*Spartina*

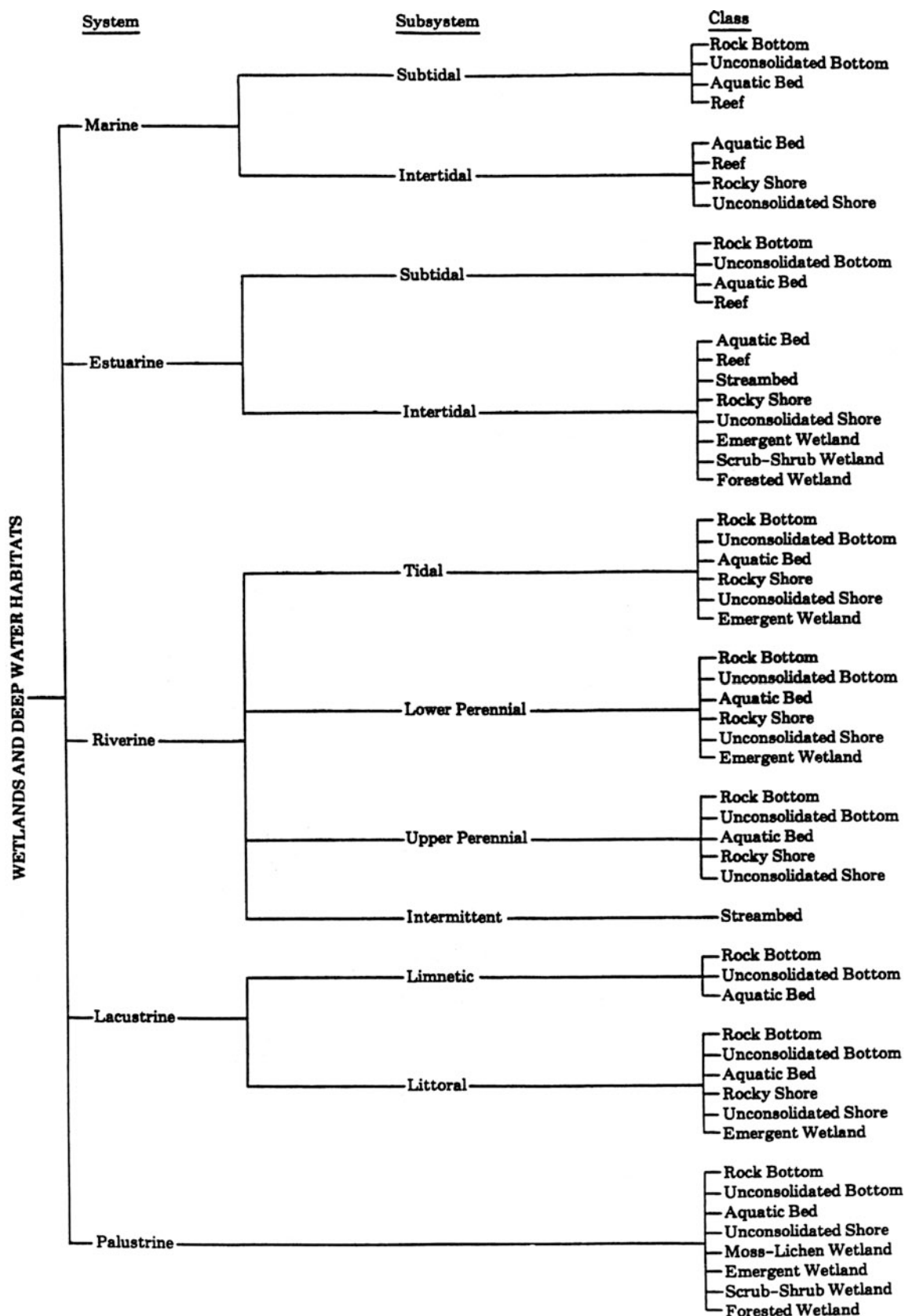
cynosuroides). The relative purity and size of salt marshes makes it possible to map them from aircraft and satellites.

Freshwater marshes are relatively diverse and have a more mixed vegetative cover producing a more complex, composite spectral signature. Forested and scrub-shrub wetlands, characterized as woody communities, are regularly inundated and saturated during the growing season. Wooded wetlands resemble spectrally wooded uplands and are therefore difficult to distinguish from wooded upland areas. Combining RADAR data with Landsat TM helps to distinguish wooded uplands from wetlands by providing soil moisture conditions beneath the tree canopy.

Wetlands Mapping

Most of the major wetlands mapping programs, conducted by the United States Geological Survey (USGS), the National Oceanic and Atmospheric Administration (NOAA), and the Environmental Protection Agency (EPA), and other agencies, are described in Kiraly et al. (1990). Traditionally the US Fish and Wildlife Service (FWS) has played a key role, conducting its first nationwide wetlands inventory in 1954, which focused on waterfowl wetlands. In 1974, the FWS established the National Wetlands Inventory Project (NWI) to generate scientific information on the characteristics of US wetlands, including detailed maps and status/trend reports. The maps are available as 7.5 min quads at a scale of 1: 24,000. Most have been digitized, converting them from paper maps to a GIS (Geographic Information System)-compatible digital line graph (DLG) format (Tiner 1985).

The NWI program produced a new classification system and a more rigorous definition of wetlands: wetlands are transitional areas between terrestrial and aquatic systems where the water table is usually at or near the surface or the land is covered by shallow water. Wetlands must also have one or more of the following attributes: (1) at least periodically, the land supports predominantly hydrophytes; (2) the substrate is predominately hydric soil; and (3) the substrate is non-soil and is saturated with water or covered by shallow water at some time during the growing season each year (Cowardin et al. 1979). The classification system developed for the NWI by Cowardin et al. (1979) is hierarchical, progressing from systems (Marine, Estuarine, Riverine, Lacustrine, and Palustrine) and subsystems (Tidal, Subtidal, Intertidal, etc.), at the most general levels, to classes, subclasses, and dominance types (Fig. 1). While suitable for use with field data and aerial photography, this classification system proved too complex for satellite remote sensors, which lacked the spatial, spectral, and temporal resolution required by such detailed mapping efforts. More recently, satellite techniques are being tested for updating the NWI maps and publishing status and trend reports every 10 years.



Remote Sensing: Wetlands Classification, Fig. 1 Classification hierarchy of wetlands and deepwater habitats showing systems, subsystems, and classes. The Palustrine System does not include deepwater habitats (Cowardin et al. 1979, Fish and Wildlife Service)

Most coastal states have used aerial photography to map their wetlands in great detail (e.g., scales of 1: 2,400) in order to satisfy legal or planning requirements.

Remote Sensing

Satellite remote sensing of wetlands was attempted, with limited success, as soon as the first Landsat MSS was launched. SPOT with its 20/10 m resolution and Landsat Thematic Mapper (TM) with its six reflected bands and 30 m spatial resolution significantly improved our ability to map large coastal marshes. Using Landsat TM data, NOAA has initiated the Coastal Change Analysis Program (C-CAP) in order to develop a nationally standardized database on land-cover and habitat change in the coastal regions of the United States. C-CAP inventories coastal submersed habitats, wetland habitats, and adjacent uplands and monitors changes in these habitats on a one-to five-year cycle with a minimum mapping unit of several hectares. This type of information and frequency of detection are required to improve scientific understanding of the linkages of coastal and submersed wetland habitats with adjacent uplands and with the distribution, abundance, and health of living marine resources. Using a rigorous protocol, satellite imagery, aerial photography, and field data are interpreted, classified, analyzed, and integrated with other digital data in a GIS. The resulting land-cover change databases are disseminated in digital form to users (Dobson et al. 1995).

C-CAP developed a classification system, shown in Table 1 to facilitate the use of satellite imagery (Klemas et al. 1993). Two study areas in South Carolina were used to

evaluate a modified C-CAP classification scheme, image classification procedures, change detection algorithm alternatives, and the impact of tidal stage on coastal change detection. The modified C-CAP Classification Scheme worked well and can be adapted for other coastal regions (Jensen et al. 1993). Unsupervised “cluster-busting” techniques coupled with “threshold 3 majority filtering” yielded the most accurate individual date classification maps (86.7–92.3% overall accuracy; Kappa coefficients of 0.85–0.90). The best change detection accuracy was obtained when individual classification maps were majority filtered and subjected to “post-classification comparison” change detection (85.2% overall accuracy; Kappa coefficient of 0.82). The multiple date images selected for coastal change detection had to meet stringent tidal stage and seasonal guidelines (Jensen et al. 1993; Jensen 1996).

Henderson et al. (1999) performed a detailed accuracy assessment of coastal land-cover mapping results obtained for Long Island using Landsat TM data and the C-CAP protocol. Table 2 displays two columns of user accuracies for C-CAP classification categories obtained by Henderson et al. (1999). The first column shows the user accuracies for the classification based on the raw spectral data, while the second column shows the accuracies after the data was recoded and filtered using ancillary verification data sets. Table 2 indicates that originally there were considerable errors in some categories, such as the Palustrine Wooded and Cultivated. However, as shown in Table 2, incorporation of ancillary data layers (e.g., aerial photographs, NWI wetland maps, etc.) increased the user accuracies of most categories into the upper 90% range, with the lowest, “Cultivated,” attaining 86% (Henderson et al. 1999).

Another way to improve the accuracy of wetland classifications derived from satellite imagery is to use multiple-date (multiple season) imagery. Multi-temporal Landsat TM imagery was evaluated for the identification and monitoring of potential jurisdictional wetlands located in the states of Maryland and Delaware (Lunetta and Balogh 1999). A wetland

Remote Sensing: Wetlands Classification, Table 1 C-CAP coastal land-cover classification system^a

Upland	Wetland	Water and submerged land
Developed land	Marine/estuarine rocky shore	Water
Cultivated land	Marine/estuarine unconsolidated shore	Marine/estuarine reef
Grassland	Estuarine emergent wetland	Marine/estuarine aquatic bed
Woody land	Estuarine woody	Riverine aquatic bed
Bare land	Riverine unconsolidated shore	Lacustrine aquatic bed (basin ≥ 20 acres)
Tundra	Lacustrine unconsolidated shore	Palustrine aquatic bed (basin ≤ 20 acres)
Snow/ice	Palustrine unconsolidated shore	
	Palustrine emergent wetland	
	Palustrine woody wetland	

^aOnly the upper two levels are shown in this table. The third, more detailed level has been omitted

Remote Sensing: Wetlands Classification, Table 2 Comparison of user's accuracy by classification category for combined raw spectral images and composite (including ancillary data) imagery

Category	Raw spectral (%)	Composite imagery (%)
Bare	92.67	93.00
Cultivated	70.00	86.00
Developed	96.67	98.00
Grassland	84.47	95.00
Water	100.00	99.00
Palustrine wooded	46.67	97.00
Palustrine wooded	85.33	97.00
Estuarine emergent	81.13	100.00
Wooded	87.33	89.00

Source: From Henderson et al. 1999. Reproduced by permission of Taylor & Francis

map prepared from single-date TM imagery was compared to a hybrid map developed using two dates of imagery. The basic approach was to identify land-cover vegetation types using spring leaf-on imagery, and identify the location and extent of the seasonally saturated soil conditions and areas exhibiting wetland hydrology using spring leaf-off imagery. The accuracy of the wetland maps produced from both single- and multiple-date TM imagery were assessed using reference data derived from aerial photographic interpretations and field observations. Subsequent to the merging of wetland forest and shrub categories, the overall accuracy of the wetland map produced from two dates of imagery was 88% compared to the 69% result from single-date imagery. A Kappa test Z statistic of 5.8 indicated a significant increase in accuracy was achieved using multiple-date TM images. Wetland maps developed from multi-temporal Landsat TM imagery may potentially provide a valuable tool to supplement existing NWI maps for identifying the location and extent of wetlands in northern temperate regions.

Summary and Conclusions

Coastal wetlands are valuable natural assets and must be protected and managed more effectively. To accomplish this, timely information on wetlands distribution, abundance, and trends are required. This information can be provided efficiently by remote sensors on aircraft and satellites. Since many wetlands occur in narrow, elongated patches and have complex spectral signatures, satellite sensors on Landsat and SPOT can provide accurate wetland maps only if multi-temporal images are used or significant amounts of ancillary data employed. Fortunately, technology, cost, and need are converging in ways that are making remote sensing and GIS techniques practical and attractive for wetlands mapping and coastal resource management (Lyon and McCarthy 1995). With the launch of Landsat 7, the cost of TM imagery has dropped by nearly a factor of 10, decreasing the cost of mapping large coastal areas. New satellites, carrying sensors with much finer spatial (1–5 m) and spectral (200 bands) resolutions are being launched and may more accurately map and detect changes in coastal habitat. Advances in the application of GIS are helping to incorporate ancillary data layers to further improve the accuracy of satellite classification of coastal wetlands and land-cover.

Cross-References

- Estuaries
- History, Coastal Ecology
- Monitoring Coastal Ecology
- Photogrammetry
- Remote Sensing of Coastal Environments

- Vegetated Coasts
- Wetlands

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Rhodoliths

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Synonyms

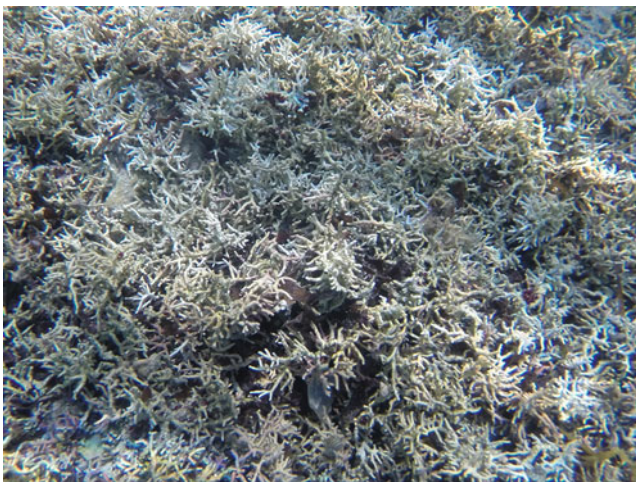
Coralline algal nodules; Maërl; Nulliporae; Prâlines; Rhodolite

Definition

Rhodoliths are unattached coralline red algae that deposit high-magnesium calcite in their cell walls and possess an unsegmented (nongeniculate) structure typically expressed in forms ranging from concentrically crustose laminations to radial branches adapted to spheroidal and ellipsoidal shapes with slow growth over 10s of years to more than 100 years duration.

Introduction

Sunlight is required for photosynthesis by all marine algae, including those in the Division Rhodophyta (from Greek roots for “rose-like” and “plant”). Coralline red algae in the Order Corallinales comprise a distinct subdivision among the red algae, based on a superficial resemblance to corals by many branching forms. The dense nodules and branched forms secreted by the living algae naturally are rose-colored or darker red to purple in complexion, but undergo a loss of pigmentation due to bleaching when exposed to direct sunlight. The term “nulliporae” meaning without pores commonly was applied by early nineteenth-century naturalists to distinguish algal growths from those of colonial corals with pores (or cups) in which polyps with feeding tentacles reside. Coralline red algae are plants not animals, but some species capable of forming rhodoliths also thrive in coral reefs, where they fill a critical role helping to cement and bind together the principal reef-forming corals into wave-resistant structures. Mainly exclusive of corals, rhodoliths thrive in vast banks that range in water depth from a few meters, as shown in Fig. 1, to more than 100 m. Widely known in Europe from the



Rhodoliths, Fig. 1 Rhodolith cover. Drifted rhodoliths with a fruticose morphology (*Neogoniolithon brassica-florida*) in the shallows of South Passage (1.5 m deep) off Dirk Hartog Island, Shark Bay, Edel Land National Park, Western Australia

Mediterranean to the Atlantic shores of France, Ireland, Cornwall, and Scotland, such banks gave rise to the Breton term for maërl beds. Extraction of rhodoliths from the seabed for use in agriculture as fertilizer has a long history in Europe, dating back several centuries.

The term “rhodolith” (from Greek roots for “rose-like” and “rock”) was promoted (Bosence 1983) as the preferred name for the dominant coralline red algae that characteristically nucleate around shell fragments or pebbles and grow as accretions in contemporary maërl beds. Surface features that rhodoliths commonly exhibit are described as varying in appearance from solidly crustose to lumpy (coarsely radiating branches with bulbous tips), fruticose (slender radiating braches with more open space between neighboring branches), and folios (delicate fronds with a bladed appearance). The taxonomic basis for rhodoliths has a history of study dating back to the eighteenth century, but the systematics for these kinds of red algae were reorganized and stabilized by Woelkerling (1988). Aspects of the reproductive cycle of rhodolith-forming red algae are preserved as features called conceptacles in the calcified cell walls and provide some of the most important descriptive criteria in their taxonomy. Such reproductive features are not always found in fossil specimens, however, making those identifications more problematic.

Global Distribution and Biodiversity

Rhodoliths are global but irregular in distribution throughout ocean shelves and along many coastlines around the world, ranging in latitude from parts as distant from the equator as New Zealand in the southern hemisphere to Spitsbergen in the northern hemisphere (Foster 2001). Poleward locations where rhodoliths thrive, such as Alaska’s Prince William Sound (Konar et al. 2006), are better known through ongoing research by the growing ranks of those who study rhodoliths. Ambient water temperature is not a factor in the healthy production of rhodoliths, although seasonality with intervals of more or less sunlight plays a greater role in temperate settings than in tropical settings. Moreover, the direction that sunlight enters the water surface varies from high to low angles across latitudes as a function of distance from the Equator and this factor makes a difference in submarine light intensity. Different species tend to be better adapted to temperate conditions than to tropical conditions. Based on survey compilations (Riosmena-Rodríguez et al. 2017), there exist worldwide about 110 species of rhodoliths formed by coralline red algae arrayed in 15 genera.

Early studies on rhodolith taxonomy and diversity were concentrated around the Mediterranean Sea, but subsequent research has shown major growth in places such as the Brazilian South Atlantic shelf and the coastal zones of the

Eastern Pacific off the Americas, where diversities exceeding 30 species of rhodolith-forming coralline red algae are well documented (Riosmena-Rodríguez et al. 2017). It is reputed that the largest marine deposits of CaCO_3 in the world occur on the tropical Atlantic shelf of Brazil, where vast rhodolith banks stretch over a latitudinal distance between 2°N and 27°S (Amado-Filho et al. 2012a). Rhodolith banks of commensurate size also appear to be active around much of Australia's southern and western continental shelf (Harvey et al. 2016).

Disparity Between Coral Reef and Rhodolith Distributions

In contrast to research on the distribution of coral reefs and diversity of coral faunas, exploration with regard to rhodolith banks still remains in its formative years. During the second voyage of the HMS *Beagle* (1831–1866), Charles Darwin became uniquely positioned to recognize the disparity in distribution between coral reefs mostly absent in the North and South Atlantic oceans in contrast to rhodoliths in many parts of those realms. Likewise, he was impressed by the luxuriance of coral reefs in places like Tahiti in the Pacific Ocean or the Cocos (Keeling) atolls in the Indian Ocean. The rhetorical question arose whether or not sediment from rivers flowing into Africa's Gulf of Guinea impeded coral growth in the Atlantic Ocean. In answer to his own query, Darwin (1842, p. 62) wrote: "But the islands of St. Helena, Ascension, the Cape Verde, St. Paul's, and Fernando Noronha are, also, entirely without reefs, although they lie far out at sea, are composed of the same ancient volcanic rocks, and have the same general form, with those islands in the Pacific, the shores of which are surrounded by gigantic walls of coral-rock." The Rocas Atoll is the only atoll in the South Atlantic Ocean and it has a framework constructed by coralline red algae in association with rich rhodolith beds (Amado-Filho et al. 2016). Likewise, the reefless islands of Fernando de Noronha support rhodolith beds (Amado-Filho et al. 2012b).

Darwin (1844) was among the earliest naturalists to appreciate the occurrence of fossil "nulliporae" as found on Santiago Island in the Cape Verde archipelago, where he associated the transformation of algal limestone to marble due to the heat and compression of overlying basalt flows (Fig. 2). Rhodolith limestone is now recognized as the principal source of carbonate rocks on at least eight of the more than 30 reefless islands scattered throughout the Cape Verde, Canary, Madeira, and Azores archipelagos of Macaronesia in the North Atlantic (Johnson et al. 2017). The distribution of living rhodoliths around many of these same Atlantic islands is still much under investigation. In contrast, evidence for the occurrence of fossil rhodoliths on islands in the Pacific Ocean appears to be slim. Eocene rhodoliths, for example, are

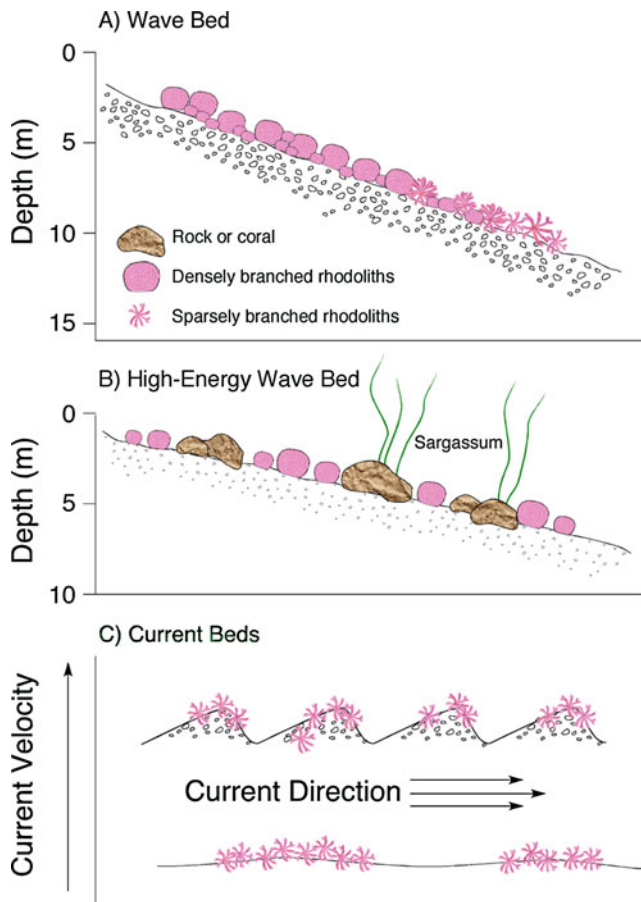


Rhodoliths, Fig. 2 Rhodolith marble, Uplifted coast of Santiago Island near Praia in the Cape Verde archipelago showing Pleistocene rhodoliths altered by an overlying flow of pillow basalt, as found by Charles Darwin in 1832

reported from Eua in the Tonga archipelago (Buchbinder and Halley 1985), but little on the fossil record of rhodoliths has been described in any detail elsewhere in the Pacific and Indian oceans. Darwin's conundrum regarding the lopsided distribution of coral reefs and rhodolith banks around the world remains unresolved. It is becoming increasingly clear, however, that living rhodolith banks rival coral reefs in the production of massive amounts of CaCO_3 .

Role as Umbrella Ecosystem

By nature, living rhodoliths have a propensity to concentrate in dense accumulations, but spread out in beds seldom piled more than two or three rhodoliths thick. Under shallow-water conditions above normal wave base (≤ 10 m), the more spherical forms of nodular and densely branched rhodoliths are subject to agitation by surface waves that promote frequent rotation and equal access to sunlight on the seabed (Fig. 3). Marine invertebrates such as bottom-dwelling arthropods and sea urchins living among the rhodoliths also help to jostle and overturn the nodules. Current beds occur in deeper waters that often extend to depths of 30 or 40 m, where tidal action shifts



Rhodoliths, Fig. 3 Rhodolith hydrodynamics, Water depth as a function of surface waves and bottom currents (Steller et al. 2009)

more delicately branched rhodoliths through asymmetric ripples oriented in the direction of current flow (Fig. 3). Water motion imparted by surface waves and by bottom currents also keeps rhodoliths from being buried by fine sediment that would otherwise block sunlight needed for growth. Under much deeper-water conditions on the Brazilian shelf (≥ 100 m), tilefish are known to heap rhodoliths into mounds and the extreme clarity of the water permits penetration of sunlight to uncommon depths.

In subtropical settings such as Mexico's Gulf of California, a variety of macrofaunal elements live freely among the rhodoliths, including several commercially exploited species of crustacea, gastropods, and bivalves. Notable among the bivalves are bay scallops (*Argopecten ventricosus* and *Lyropecten subnodosus*). The common sea urchin (*Taxapneustes roseus*) acts as a kind of bulldozer plowing its way through the rhodolith field. In particular, the cryptofauna living on or within rhodoliths is especially rich with more than 100 taxa recorded. The most diverse groups among the internal fauna hidden within rhodoliths are the marine worms (polychaetes) and crustaceans (Steller et al. 2009). Other invertebrates often found imprisoned within

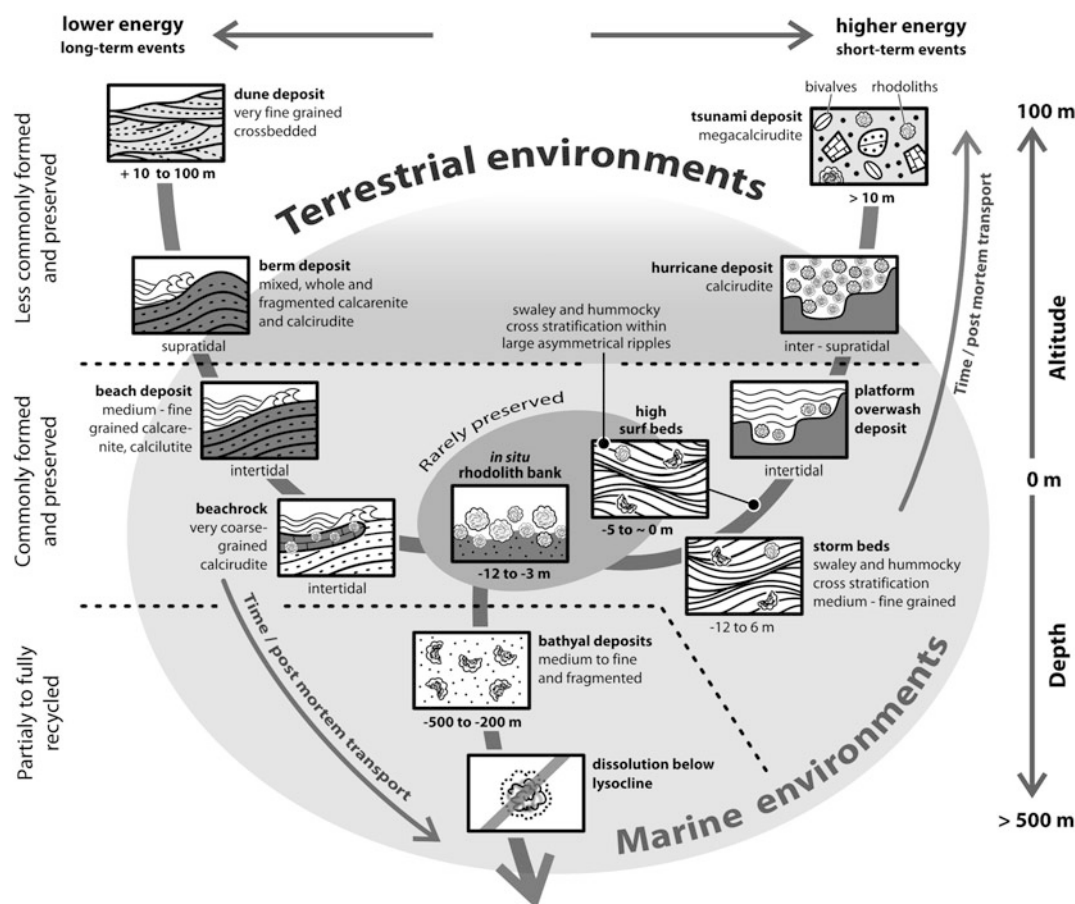
the tight branches of rhodoliths include immature arc shells and mussels, as well as small corals, chitons, barnacles, echinoids, and brittle stars. In addition, a wide variety of fleshy green and brown algae utilize rhodoliths for anchorage in flux on a seasonal basis. A record biota of 238 taxa, including 197 marine invertebrates, 37 algae, and 4 fish, is associated with maërl beds in northern New Zealand (Riosmena-Rodríguez et al. 2017). In this regard, rhodoliths fulfill a major role as an umbrella ecosystem that directly shields or otherwise supports a substantial marine fauna and flora.

Rhodolith Fossil Record

Compared to the fossil rhodoliths of Pleistocene age collected by Charles Darwin in the Cape Verde Islands and still held in the collections of the Sedgwick Museum at Cambridge University, the oldest fossil rhodoliths date to the Early Cretaceous about 140 million years ago (Aguirre et al. 2010). Based on fossils from dozens of places around the world, it is argued that the acme of rhodolith development in terms of dominance of CaCO_3 production occurred from about 20 to 10 million years ago culminating with the Miocene Climatic Optimum, when the role of corals as tropical carbonate producers suffered a temporary imbalance possibly due to changes in ocean circulation triggered by growth of the Antarctic Ice Sheet (Halfar and Mutti 2005).

Post-Mortem Transfer to Coastal Deposits

From the viewpoint of geomorphology in the coastal sciences, rhodolith deposits accrue postmortem on shore under a range of circumstances that entail variations in the intensity and duration of events responding to wind and wave energy (Fig. 4). Because rhodoliths are unattached organic objects with a tendency to grow in spherical to ellipsoidal shapes, they are highly susceptible to transport from place to place on the seabed. Located between 15° and 40° North latitude, islands in the North Atlantic Macaronesian realm provide a virtual laboratory for the study rhodolith taphonomy, or the kinds of processes that occur between death and eventual burial on the way to fossilization (Johnson et al. 2017). The Canary Islands (Fig. 5a), for example, are under the steady influence of the prevailing Northeast Trade Winds. Ocean swells directed to the southwest are generated by winds that commonly register between 5 and 6 on the Beaufort scale, defined as fresh to strong breezes with speeds between 29 and 49 km/h. Living rhodoliths may be studied in maërl beds off the several islands in the Canary archipelago, but also of interest are coastal deposits resulting from the transfer of enormous numbers of rhodoliths to supratidal settings where they undergo bleaching and expire in bulk.



Rhodoliths, Fig. 4 Rhodolith taphonomy, Scheme for the sedimentary recycling of rhodoliths and rhodolith debris as a function of every-day coastal processes as opposed to intermittent storms, hurricanes, and tsunami events

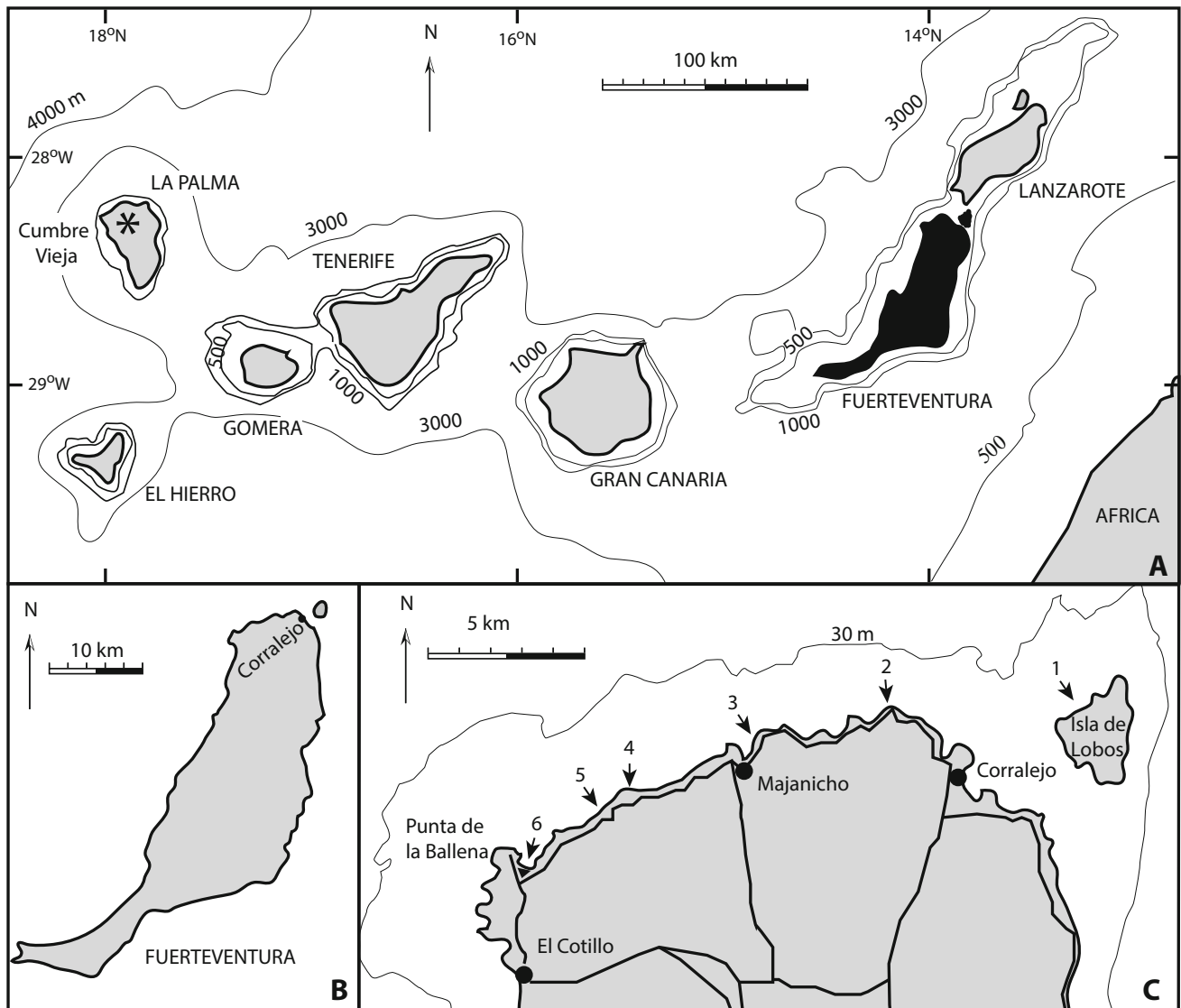
Gradual, Lower-Energy Events

The exposed north shore of Fuerteventura between the town of Corralejo and Punta de la Ballena (Fig. 5b and c) features multiple contemporary rhodolith deposits separated over a short distance. At Caleta del Bajo de Mejillones closest to Corralejo (Fig. 5c, locality 2), rhodoliths comprise an over-wash deposit 120 m long by 10 m wide in the supratidal zone above a basal wave-cut platform. The stranded rhodoliths are small, from 10 to 30 mm in diameter with approximately 5,000 to a square meter. It may be calculated that the deposit's upper surface alone amounts to six million rhodoliths.

Westward near Majanicho, (Fig. 5c, locality 3), another deposit of bleached rhodoliths is similar to that at Caleta del Bajo de Mejillones, but in the form of a berm. Said by local inhabitants to have remained unchanged over recent memory, the berm is 150 m long by 9 m wide and rises 1.4 m above the landward termination of a broad, wave-cut platform (Fig. 6). The berm's volume amounts to 1,350 cubic meters containing an estimated number of popcorn size rhodoliths

(Fig. 7) amounting to 45 million individual pieces. These small rhodoliths from Caleta del Bajo de Mejillones and Majanicho appear to be monospecific in to the genus *Lithothamnion*. Within the bay at Majaicho, traces of brown algae anchored to scattered rhodoliths remain intact. This phenomenon, where the attachment of foliose algae to the rhodoliths acts to catch the wind, is known in Brazil as the *arribadas* or "the arrived." It means that the rhodoliths originated in well-lighted waters where the brown algae thrive, but arrive seasonally on the beaches due to onshore winds that drag the rhodoliths shoreward.

At Playa Hierro (Fig. 5c, locality 4), an exposed beach facing northeast with an area not less than 1,500 m² is composed of coarse rhodolithic sand mixed with coarse grains of basalt eroded from the adjacent wave-cut platform (Fig. 8). Beaches farther west at Caleta la Seba and Caleta de Beatriz (Fig. 5c, localities 4 and 5) are dominated by sand derived from more finely pulverized rhodoliths and smaller basalt grains. Inland south of Caleta del Marrajo (Fig. 5c, locality 6), a dune field covering 40,000 m² is composed of coarse to medium sand (0.25–1.0 mm in diameter) derived



Rhodoliths, Fig. 5 Canary Islands, Maps, (a) For the entire archipelago showing the location of Fuerteventura (black) and the volcano Cumbre Vieja on La Palma (star), (b) Fuerteventura with location of

the town Corralejo, (c) North coast of Fuerteventura marking places (1–6) with contemporary rhodolith shore deposits

from finely reduced rhodolith debris and minor basalt lithics. The persistent trade winds at this locality promote beach deflation, carrying a finer fraction of carbonate sand inland in a southerly direction as confirmed by asymmetrical sand ripples. Over a distance of only 17 km, the transition from platform over-wash deposits to berm, beach, and sand dune deposits (Fig. 4) is well demonstrated across the north coast of Fuerteventura Island (Fig. 5c). Transfer of the smaller, whole rhodoliths to supratidal settings at Caleta del Bajo de Mejillones and Majanicho likely represents minor storm events that struck the coast during high tides when the adjacent rocky tidal flats were flooded. Except for a few rhodoliths caught in rock pools close to shore, the extensive rock platforms are swept free of all debris by the surf.

Rapid, Higher-Energy Events

Compared to the relatively immature rhodoliths from the north shore of Fuerteventura (Fig. 7), rhodoliths from supratidal deposits on the islands of Sal and Maio in the Cape Verde archipelago are larger (9–10 cm in diameter) and therefore older in age. Rhodoliths cemented into thin slabs of limestone occur as narrow bands of beach rock on the windward north shores of Maio (Fig. 9) and also on Sal. It is thought these larger rhodoliths were transferred shoreward from an original living depth of around 50 m or more by major storms of near hurricane intensity (Johnson et al. 2016, 2017). The most extensive over-wash deposit on Maio exhibits a density of 450 rhodoliths/m² spread out over an area of 27,000 m² inland

Rhodoliths, Fig. 6 Coastal berm, Extensive berm (1.4 m high) formed by whole rhodoliths above a wave-cut platform on basalt near Majanicho on Fuerteventura, Canary Islands (locality 3, Fig. 5c)



Rhodoliths, Fig. 7 Bleached rhodoliths, Popcorn size, lumpy rhodoliths from the berm near Majanicho on Fuerteventura

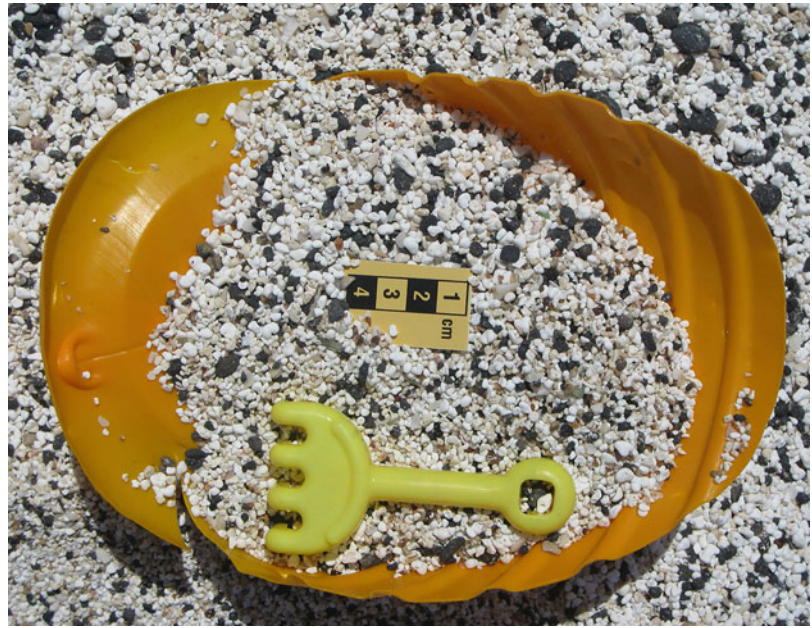


at an elevation about one meter above sea level (Johnson et al. 2016). A rare specimen of a single rhodolith encrusted around a fragment of ceramic tile indicates that the Maio deposit is not a fossil deposit, but dates from historic times.

More so than hurricanes, tsunami events release the highest concentration of energy that infrequently impacts shorelines and, hence, also have the potential to disrupt rhodolith banks (Fig. 4). Characterized as a giant tsunami deposit on the northeast flank of Santiago in the Cape Verde archipelago (Ramalho et al. 2015), basalt boulders up to 8 m in diameter and weighting as much as 700 mg are clustered near a cliff edge 650 m inland at elevations up to 190 m above sea level. Below the cliff line, chaotic conglomerates and sand

sheets are plastered against slopes as steep as 20° from near sea level up to about 100 m in elevation. The sand sheets are shingled in a landward direction and dominated by broken and whole rhodoliths in a matrix of bioclastic sandstone. Geomorphologic studies of the stratovolcano on Fogo, the youngest volcano in the Cape Verde archipelago, show that the northeast flank of that volcano suffered a major gravitational collapse sometime between 65,000 and 124,000 years ago. Santiago Island is located only 55 km away from Fogo and the chaotic deposits on the northeast side of that island are surmised to have resulted from the flank collapse on Fogo (Ramalho et al., 2015). The maximum depth of the sea floor separating Fogo and Santiago islands is 4,000 m, far below

Rhodoliths, Fig. 8 Beach sand, Mixed carbonate and volcanic sand derived from eroded rhodoliths and basalt at Paya Hierro on Fuerteventura (Fig. 5c, locality 4)



the insular shelf surrounding parts of Santiago where rhodoliths were most likely to thrive. However, given that the avalanche debris from Fogo has an estimated volume from 130 to 160 km³, the resulting catastrophic displacement of water in tsunami waves was sure to have swept along the entire west coast of nearby Santiago including its insular shelf.

All the oceanic islands of Macaronesia are volcanic in origin, but they differ widely in age, generally with older islands situated more to the east and younger islands toward the west. In the Canary Islands, for example, the basal volcanic complex on Fuerteventura dates from the late Oligocene to early Miocene, whereas the island of El Hierro was volcanically active as recently as 2012. The last major eruption of the volcano Cumbre Vieja on La Palma (Fig. 5a) occurred in 1949, but there is widespread speculation that the volcano's instability could result in a major flank collapse that would trigger a massive tsunami potentially affecting the entire North Atlantic basin. Essentially, this danger is a risk associated with all oceanic volcanoes with the inherent consequences of widespread coastal impact.

Variations in Rhodolith Limestone

Based on contemporary rhodolith deposits from the Canary Islands as a guide, it is possible to identify examples of different kinds of rhodolith limestone preserved elsewhere throughout the Macaronesian island groups (Johnson et al. 2017). Upper Miocene rhodolith limestone is banked against former rocky shores as over-wash deposits in a continuous outcrop belt around the northeast and southeast coast of São Nicolas



Rhodoliths, Fig. 9 Beach rock, Limestone slabs with large rhodoliths preserved as beach rock, Praia Real, north coast of Maio in the Cape Verde Islands

in the Cape Verde islands, western Fuerteventura in the Canaries, as well as southeast Santa Maria in the Azores. A middle Miocene hurricane deposit composed of large rhodoliths is preserved trapped among former sea stacks on

Ilhéu de Cima off Porto Santo in the Madeira archipelago. Extensive Pleistocene dune rocks dominated by rhodolith sand occur on Maio and São Nicolas in the Cape Verde islands, as well as Porto Santo in Madeira. The rhodolith limestone formations of Macaronesia prove to be valuable in the classification of related limestone deposits in places as remote from one another as Baja California (Mexico), New Zealand, Hawaii, and Western Australia.

Economic Resources

Both in fossil form and as present-day maërl beds, rhodoliths are valuable as economic resources in different ways. During the Middle Miocene (from about 16 to 14 million years ago), rhodoliths and the bioclastic sediments derived from rhodoliths were especially widespread on the European margins of the Paratethys seaway (Colletti et al. 2017) that later devolved to the Mediterranean Sea. Carbonate banks as thick as 40 m were deposited around the Leitha Mountains and Ruster Hills of present-day southern Styria, Austria. The Leitha Limestone (*Leithakalk*) has a long history of use as a building stone dating back to Roman times, when the cities of Carnuntum and Flavia Solva were built (Moshhammer et al. 2015). Evidence of quarrying comes from first century CE underground mines at Alfenz, where pillars up to 8 m in height remain in place to support roof rocks below which the rhodolith-rich stone was excavated. During the Middle Ages, the Leitha Limestone was prized by stone masons for specific usages, including stairways, columns, and sculptures. During the late nineteenth century, wide use of the Leitha Limestone was made for buildings on Vienna's famous Ringstrasse. More recent use of this product as a facing stone can be found today at entrances to the Vienna subway system on the Ringstrasse (Fig. 10). A few Austrian quarries

still exploit the Leitha Limestone for use in restoration projects on historic buildings.

As an ancillary product associated with building stone, rhodoliths and rhodolith limestone have been widely used to make slaked lime in limekilns. Beginning in the early 1400s during the age of discovery, local markets for mortar and whitewash were developed as a result of Spanish and Portuguese settlement throughout the Macaronesian archipelagos. Whereas island basalt was readily available for use as building stone, the mortar needed for cement and plaster was a commodity in short supply. One of the largest such sites exploited on a commercial scale for slaked lime anywhere in Macaronesia was established on Isla de Lobos off the north-east coast of Fuerteventura in the Canary Islands (Fig. 5c, locality 1). Now protected as a provincial park, much of the island was stripped during the late 1700s and early 1800s both for its rhodolith limestone and for loose rhodoliths that washed into the island's bays in prodigious numbers. Loose rhodoliths were taken off the beaches in cartloads for processing in the limekilns. On Porto Santo in the Madeira Islands, open-pit mines fed rhodolith sand to the limekilns and extensive subsurface galleries were excavated for valuable rhodolith limestone on Ilhéu de Baixo. Abandoned limekilns are known from sites on Maio and São Nicolas in the Cape Verde Islands, as well as Santa Maria Island in the Azores (Johnson et al. 2016).

Due to the porosity of limestone formed by coralline red algae, it provides a superior reservoir rock capable of holding large volumes of oil and gas. As an example, the Asmari Formation of southwest Iran is a rhodolith-rich limestone originally deposited in a tropical setting during the late Oligocene and early Miocene. This extensive subsurface formation constitutes a prolific reservoir, said to be responsible for production of 90% of Iranian oil (Colett et al. 2017). It is no coincidence that the historic oil fields of Hungary were

Rhodoliths, Fig. 10 Limestone building stone, Example of the rhodolith-rich Leitha Limestone (Middle Miocene) widely used as a facing stone in Vienna, Austria



sourced in the Leitha Limestone as reservoir rock. With an estimated reserve of more than a billion barrels of oil waiting to be tapped, submarine strata in the South China Sea attributed to the Lower Miocene Zujang Formation consist of coralline algal limestone with abundant fossil rhodoliths and corals (Coletti et al. 2017).

Depletion of the European maërl beds for use as fertilizer in agriculture has a dim future, due to regulations by the European Union that list the dominant rhodolith species of Atlantic maërl as a nonrenewable resource. Furthermore, licensing for maërl extraction has now ceased in the United Kingdom (Coletti et al. 2017). The situation in Brazil is far from settled, as proponents lobby for concessions in government licensing of the vast maërl beds on that country's Atlantic shelf.

Summary

Rhodoliths are long-lived, free rolling accretions of coralline red algae that first appeared in the fossil record in the Early Cretaceous 140 million years ago and likely reached peak production of carbonate limestone surpassing that of coral reefs during the Miocene Climatic Optimum 14 million years ago. Adjacent to continental shores and around certain oceanic islands, the distribution of rhodolith banks crosses a wide range of latitudes from tropical to temperate settings. Roughly 110 living species of rhodoliths formed by coralline red algae arrayed in 15 genera are known, worldwide. The largest rhodolith banks occur in tropical to subtropical settings on the South Atlantic shelf of Brazil and in temperate to subtropical settings on the Indian shelf of Australia. Rhodoliths constitute a vital ecosystem that shelters an associated biota numbering in the 100s. In terms of coastal processes, rhodoliths are subject to transport by daily surface waves and bottom currents, as well as storms, hurricanes, and even tsunami events that result in substantial supratidal deposits ashore. In particular, rhodoliths may be recycled in great volumes to form beaches, berms, and sand dunes. Rhodoliths and rhodolith limestone have played important roles in the provision of building stone and slaked lime, but the typically porous limestone has its greatest impact as a reservoir rock in active oil fields.

Cross-References

- Beach Processes
- Beachrock
- Coastal Changes, Gradual
- Coastal Changes, Rapid
- Coral Reef Islands
- Eolian Processes

- Storm Surge
- Volcanic Coasts
- Wave-Dominated Coasts

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Rhythmic Patterns

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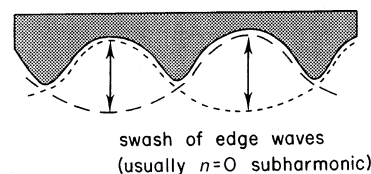
Beaches are seldom straight or smoothly curved in the longshore direction. Instead, they commonly include seaward projections of sediment, termed cusps, or embayments locally cut into the shore. Such features may be isolated, but more often occur in groups of alternating cusps and embayments that have a fairly regular spacing; they are then referred to as rhythmic patterns. There are several recognized types of rhythmic patterns, including beach cusps, sand waves, and giant cusps (Komar 1998).

A wide range of spacings of rhythmic patterns can be found on beaches. Along the shores of ponds and small lakes the spacings between adjacent cusps may vary from less than 10 cm to 1 m. On ocean beaches with small waves, the spacing may be on the order of 2 m, while those built by large storm waves may be 50 m or more. Other rhythmic patterns, sand waves and giant cusps, have still larger spacings, typically ranging from 150 to 1500 m, but with most being between 500 and 750 m, with the cusps projecting on average some 15–25 m seaward from the embayments.

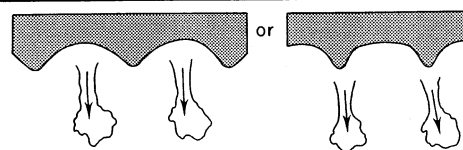
The Classification of Rhythmic Patterns

In the past, the classification of different types of rhythmic patterns has been based on the lengths of their spacings. Beach cusps were considered to have the smallest spacings, less than 25 m, while sand waves and giant cusps have larger spacings. These latter terms can be considered to be nearly synonymous, different names for the same or very similar features. Research in recent years has led to a better understanding of the formation of the various types of rhythmic patterns, and this now makes it possible to develop a genetic classification that depends on the processes of waves and currents that are responsible for their formation, rather than depending simply on their spacings (Komar 1998). Furthermore, it is clear that rhythmic patterns having a wide range of spacings can be generated by a single mechanism, and more than one mechanism may be capable of producing rhythmic patterns having the same spacing. It is clear therefore that a genetic classification is needed, one that reflects the processes of formation. Such a genetic classification is depicted in Fig. 1 (Komar 1983), one that distinguishes between beach cusps, systems of rip-current embayments and cusps, series of

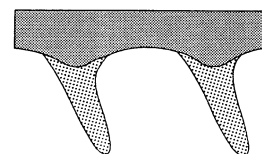
reflective beach cusps



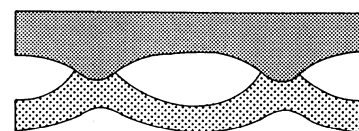
rip current embayment–cusp system



transverse and oblique bars



crescentic bar–cusp system



Rhythmic Patterns, Fig. 1 A genetic classification of rhythmic patterns where the origin of the series of cusps and embayments can be attributed to different processes of waves and currents. (After Komar 1983)

transverse bars that produce cusps along the shore, and crescentic bars that are chiefly an underwater feature but can produce a rhythmic pattern that extends onto the dry part of the beach. In general, this order represents an increase in cusp spacings, with the latter three mechanisms yielding what had formerly been referred to as sand waves or giant cusps.

Beach Cusps as a Type of Rhythmic Pattern

The most easily recognized rhythmic pattern seen on beaches are the cusped deposits of sand and gravel built by waves and known as beach cusps. Because of their marked regularity with nearly uniform spacings, beach cusps have attracted many observers and much speculation as to their origin. Arguments still persist with regard to the processes of wave motions and sediment transport that control their formation and determine the lengths of their spacings.

Beach-cusp formation is most favorable when the waves approach normal to the beach, that is, with their crests parallel to the longshore trend of the beach. This may explain why pocket beaches are particularly favorable sites for beach cusp formation. Furthermore, regular waves having long crest lengths are particularly conducive to cusp formation; beach cusps generally are not formed by irregular, confused seas.

Various investigators have described contrasting patterns of water circulation induced by the swash of waves around the cusps and within their embayments. In some cases there is an alternating surge inward and out of the embayments. Water flows from one embayment where the wave-swash runup has been a maximum, around the nearest cusp, and into a neighboring bay where it rushes up the beach face with the next wave. Thus, the maximum runup alternates in its timing between adjacent embayments. In other cases, the arriving waves break evenly along the beach, but the wave surge then piles up against the steep cusps and is divided into divergent streams that flow into the adjoining embayments. These streams head off the wave surge that had flowed directly up into the embayment. The two side streams from the cusps on either side meet at the center of the bay, and together form a seaward flow of considerable strength. This return flow can resemble a rip current, but unlike true rip currents, the flow is discontinuous and the mechanisms of formation are quite different. These contrasting patterns of water circulation observed around beach cusps suggests that more than one mechanism may give rise to their formation.

A number of hypotheses have been proposed to account for the formation of beach cusps. For a theory to be acceptable, it must account for the uniformity of spacing within an observed series of cusps, and the way in which this spacing is related to the wave parameters. The hypothesis that has been most successful in explaining the formation of beach cusps and accounting for their spacings is one based on the presence of edge waves in the surf zone (Guza and Inman 1975). An

edge wave is a type of wave that is trapped in the surf by refraction across the slope of the beach, moving in the longshore direction as it alternately refracts while moving offshore, then bends entirely around to return to the shore, reflects from the beach, and repeats the pattern of movement. The important result is that the presence of the edge wave affects the intensity and distance of swash runup on the beach, producing a regular runup spacing along the length of shore. This hypothesis explains the formation of beach cusps as the rearrangement of the sediment into a regular pattern of alternating cusps and embayments, corresponding to the longshore wave length of the edge waves and spacing of their maximum runup on the beach. The validity of this hypothesis has been demonstrated in the controlled conditions of laboratory wave basins, and by a few studies on ocean beaches that happened to be measuring edge waves at the same time beach cusps formed (see review in Komar 1998). In that the longshore length of edge waves is determined by the wave period and slope of the beach, this hypothesis yields a mathematical equation that predicts the beach-cusp spacing (Guza and Inman 1975). Measured beach-cusp spacings ranging from 0.1 to nearly 100 m have been shown to agree with this mathematical relationship (Komar 1998). Thus, there is strong supporting evidence for the edge wave hypothesis of beach cusp formation.

There is, however, an alternative hypothesis that has been proposed to account for the formation of beach cusps, the so-called “self organization” hypothesis of Werner and Fink (1993). It envisions an initially smooth, straight beach, lacking cusps, but with waves arriving and swashing up the beach face with some degree of irregularity. According to the hypothesis, this irregularity in the wave swash produces variable amounts of beach sand movement along the shore, with a tendency for the sand to preferentially accumulate in a few isolated areas. The critical aspect of this hypothesis is that the zones of accumulated sand then affect the subsequent patterns of wave runup, thereby increasing the sizes of the accumulated sand and causing them to evolve into a pattern of beach cusps having regular spacings. It is this trend toward increasing regularity to which the name “self organization” refers. The main supporting evidence for this hypothesis comes in the form of computer simulation models that demonstrate the possibility of such an evolution. While not actually having mathematical relationships that predict the eventual beach cusp spacing, the computer models demonstrate that the spacing depends on the wave period and height, and on the beach slope, and in fact the model yields cusp spacings that are similar to those predicted by the edge-wave hypothesis. This similarity in prediction has made it difficult to distinguish which mechanism is responsible for the formation of beach cusps on ocean shores. It is possible that both hypotheses can, in different situations, account for the formation of beach cusps, and in some cases may actually work together.

As noted above, the early definition of “beach cusps” restricted them to spacings of 25 m or less, but provided no explanation for their formation. Although we are still uncertain as to the specific generation mechanism, the proposed hypotheses appear to offer satisfactory explanations, so the term “beach cusps” is now used in genetic classifications like that in Fig. 1. For the most part the old and new uses of the term refer to the same rhythmic pattern, but we now recognize that edge waves and perhaps self-organization can generate beach cusps that have spacings up to 100 m.

Rip Current Embayments and Cuspate Shores

Within the series of rhythmic patterns diagrammed in Fig. 1, generally the next larger form beyond beach cusps is the system of erosional embayments and intervening cusps formed by nearshore current systems that include seaward-flowing rip currents. In most instances the rip currents erode sand from the beach and transport it offshore, forming embayments at the rip-current positions, with cusps midway between. In rarer instances, particularly on steep beaches, coarse sand and gravel may accumulate at the shoreward ends of the rips, developing cusps at those positions. Rip currents, and hence the embayments and cusps, typically have spacings that range from tens to hundreds of meters. As such, the resulting rhythmic pattern corresponds to what has been referred to variously as sand waves or giant cusps.

The effect of the nearshore currents on the shore, forming series of embayments and cusps, is the surface expression of the underwater topography that is molded by the currents acting together with the waves. The seaward-flowing rip currents tend to erode channels across the full width of the beach within the embayments, segmenting the off-shore bar. This leaves a system of cusps midway between rip currents, but each cusp seen on the shore is part of a shoal that extends out to the remaining segment of offshore bar.

This form of rhythmic pattern with embayments cut by rip currents is often important to property erosion in that the embayments narrow the beach width and remove most of the buffer protection offered to properties backing the beach. Although the rip embayments themselves do not usually produce much erosion of dunes and sea cliffs, they provide an area of deeper water where storm waves can approach close to shore before breaking against the coastal properties.

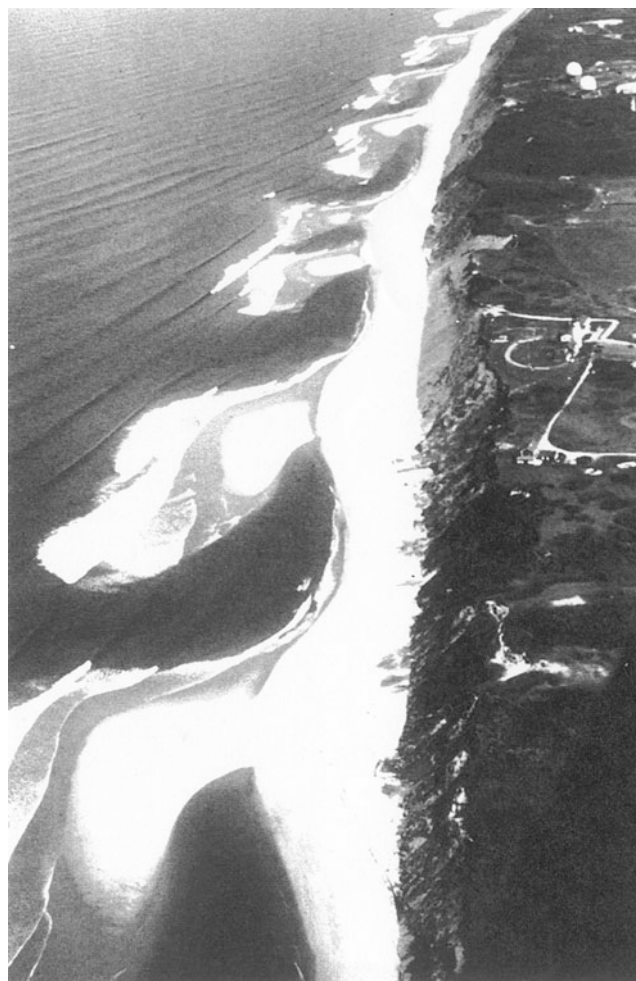
Rhythmic Patterns Produced by Welded and Transverse Bars

A variety of sand bars have been observed in the nearshore that run obliquely to the longshore trend of the beach. These have been termed “welded” or “transverse” bars (Fig. 1). An example is shown in Fig. 2 on the ocean shore of Cape Cod, Massachusetts, consisting of a series of bars and a cusped shore that has a distinctive longshore rhythmicity of several

hundred meters. Therefore, the presence of welded or transverse bars can also give rise to a rhythmic pattern.

A number of suggestions have been made for the origin of bars that trend obliquely to the shore, and for the corresponding rhythmic pattern. It has been observed that when waves break at pronounced angles to the beach, the offshore bars that were originally parallel to the shore and segmented by evenly spaced rip currents, rotate to align themselves with the incoming wave crests. This may be the origin of the welded bars and rhythmic pattern seen in Fig. 2.

Another type of oblique bar is found on coasts of low wave energy, for example in lakes or along the shore of the Gulf of Mexico. Referred to as transverse bars, they tend to occur in families that run parallel to one another, directed toward the offshore. At each point where a transverse bar joins the shore, a large cusp develops on the dry beach. Transverse bars can be fairly permanent features – examples on the shores of the Gulf of Mexico have been observed to persist in aerial photographs



Rhythmic Patterns, Fig. 2 A series of welded bars and associated rhythmic pattern on Cape Cod, Massachusetts. (Photo courtesy of David S. Aubrey, Woods Hole Oceanographic Institution)

that span 25 years, showing little or no tendency to migrate alongshore during that time. Investigations have demonstrated that this type of transverse bar affects the paths of nearshore currents, with the current being concentrated over the bar and flowing offshore along its length, thereby perpetuating the bar's existence and extending its length.

Crescentic Bars and Large-Scale Rhythmic Patterns

Crescentic bars are one form of submerged offshore bars where rather than being linear, they have a regular lunate or crescentic shape (Fig. 1), together with a uniform repetition along the length of beach. This regularity generally cannot be appreciated by observers on the dry beach, since most of the feature is underwater. However, there may be an associated series of cusps on the beach if the landward ends of the crescentic bars attach to the shore. In this instance, the presence of offshore crescentic bars leads to the development of another form of rhythmic pattern.

Crescentic bars are much larger features than beach cusps, and generally are somewhat larger than the rhythmic patterns due to rip currents or welded bars. In some instances, large crescentic bars form the outer-bar system of a beach, while the inner bar is linear and segmented by the more closely spaced rip currents. The range in lengths of crescentic bars is difficult to establish, since for many reported occurrences it is not possible to determine conclusively whether crescentic bars or some other form of rhythmic pattern is being described. Crescentic bars appear to range up to 2000 m in length (Komar 1998). At times there can be multiple crescentic bars on a beach, the further offshore the bars the larger their spacings. The corresponding rhythmic pattern on the beach would similarly have very large spacings between successive cusps.

Like beach cusps, the regularity in shapes and the even spacings of crescentic bars have inspired a number of suggestions as to their formation. The mechanism proposed by Bowen and Inman (1971), again by the movement of edge waves, provides the most reasonable explanation. In this case, however, important is the velocity of water movements associated with the edge waves, not their swash runup on the sloping beach which may be responsible for beach cusps. According to this mode of formation, beach sediment in the outer surf zone drifts about under the currents of the edge waves, until the sand reaches zones where the water velocity is low and the sand can accumulate. According to computer models of edge wave motions, this rearrangement of the sand would yield lunate-shaped bars that are remarkably similar to those observed on ocean beaches, a result that argues in favor of this hypothesis. Bowen and Inman (1971) conducted a series of laboratory wave-basin experiments that further confirmed this predicted sand accretion pattern, leading to the formation of crescentic bars. At this time, there is no reasonable alternative hypothesis for crescent-bar formation that

satisfactorily accounts for their regularity in shapes and spacings, and for the formation of the associated rhythmic pattern.

Rhythmic Patterns, Irregular Shores, and Coastal Erosion

Depending on the mechanism, rhythmic patterns may consist of alternating cusps and embayments whose spacings range from a few meters (beach cusps), to on the order of 100 m (rip-current embayments or welded bars), and on up to 500–1500 m (crescentic bars). If one includes series of cusps or forelands or capes like those that exist along the southeast coast of the United States, the series can be extended up to tens of kilometers. When only one type of rhythmic pattern is present on the beach, it gives rise to a fairly regular spacing of alternating cusps and embayments. However, it is common for more than one type of pattern to occur simultaneously on a beach, and the summation of what are otherwise regularly spaced patterns can lead to an irregular beach and shore. An example is shown in Fig. 3, the Cape Hatteras coast of North Carolina, photographed in 1970 before the lighthouse was moved (Dolan 1971). Apparent are the series of large cusps and embayments, with a fair degree of regularity along the



Rhythmic Patterns, Fig. 3 A rhythmic pattern of alternating cusps and embayments on the Cape Hatteras coast of North Carolina, photographed in 1970, with the largest embayment producing beach erosion and threatening the lighthouse. (From Dolan 1971)

length of coast covered by the photograph. However, there is a level of irregularity, with some embayments being larger than others. Although the cause in this example is uncertain, it is probable that the irregularity of the rhythmicity is produced by the summation of two rhythmic patterns, that due to a rip-current cusps/embayment system, together with a larger-scale variation due to offshore crescentic bars. Of interest in this example, the respective embayments produced by rip currents and the crescentic bars appear to have combined in the area of the Cape Hatteras Lighthouse, resulting in the total loss of the beach and erosion of the dunes to the extent that the lighthouse was in danger. Therefore, an understanding of the origin and types of rhythmic patterns can be important to interpretations of the causes of coastal erosion problems.

Cross-References

- [Accretion and Erosion Waves on Beaches](#)
- [Bars](#)
- [Beach Features](#)
- [Beach Processes](#)
- [Cusate Forelands](#)
- [Rip Currents](#)
- [Surf Zone Processes](#)

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Ria

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Definition

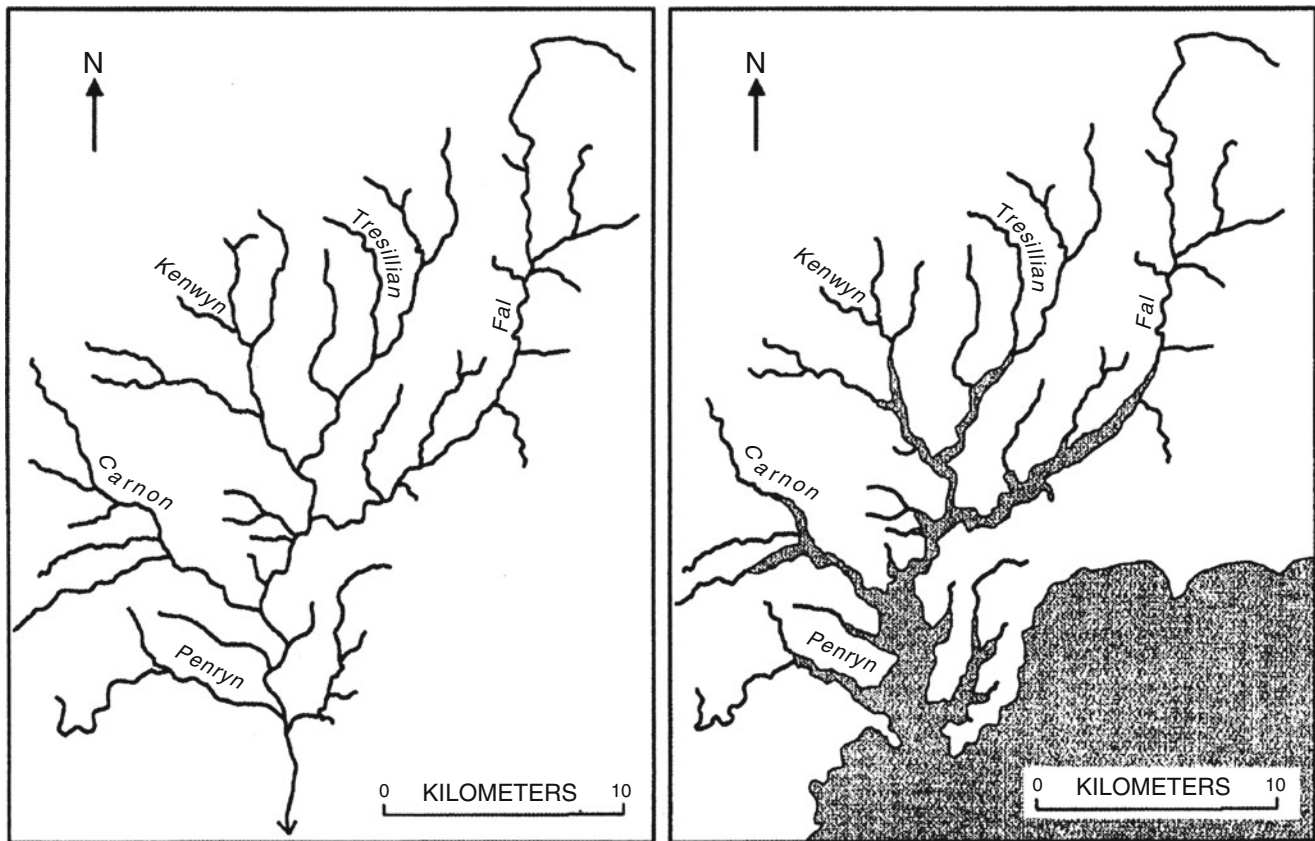
A ria is a long, narrow, often branching inlet formed by marine submergence of the lower parts of a river valley that had previously been incised below present sea level.



Ria, Fig. 1 The ria at Aber Benoît, Brittany. (Copyright, Geostudies)

Rias are the drowned mouths of unglaciated valleys, usually bordered by steep slopes rising to mountains, hills, or plateaux (Fig. 1). The term is of Spanish origin, derived from large inlets on the coasts of Galicia such as the Ria de Arosa and the Ria de Muros y Noya, fingering far inland. They are known as abers in Brittany and Wales.

Von Richthofen (1886) defined a ria as a drowned valley cut transverse to the geological strike, but the Rias of Galicia do not meet this strict definition (Cotton 1956). Perusal of coastal textbooks indicates that the term has come to be used as a synonym for a drowned valley mouth without any structural constraint (Goudie 2004; Bird 2008). Rias generally have a dendritic (tree-like) outline, remaining open to the sea, as in Carrick Roads (Fig. 2) in southwest England, Chesapeake Bay in the United States, and Port Jackson (Sydney Harbor) in Australia. The long, straight valley-mouth gulfs on the southwest coast of Ireland, such as Bantry Bay, are examples of rias, even though they follow a geological strike that runs transverse to the general coastline. Where rivers have cut valleys across geological structures, there may be tributaries that follow the geological strike, submerged to form a trellised pattern, as in Cork Harbor in southern Ireland. On the Dalmatian coast of the Adriatic Sea, rias are elongated straits along valleys that follow the geological strike, linked to the sea by transverse channels.



Ria, Fig. 2 The Carrick Roads ria on the south coast of Cornwall, England. (Left) The river systems as they were 20,000 years ago, during the Late Pleistocene low sea level phase. (Right) The present outlines,

after partial submergence by the Late Quaternary marine transgression. (Copyright-Geostudies)

Evolution of Rias

Existing rias were formed by marine submergence during the Late Quaternary (Flandrian) marine transgression, but in southwest England, there is evidence of several phases of valley incision during Pleistocene low sea-level phases, alternating with earlier ria formation during interglacial marine transgressions and at least one Late Pleistocene higher sea-level phase indicated by an emerged beach (Kidson 1977).

The Galician rias are generally wide and deep (up to 30 m) marine inlets in valleys that may have been shaped partly by tectonic subsidence and the recession of bordering scarps. Subsidence has probably contributed to the persistence of the ria at the mouth of Johore River in southeastern Malaysia, which remains a wide and deep inlet, whereas other Malaysian valley mouths have been infilled as alluvial plains, some with protruding deltas. Other rias persist because they were initially deep and sedimentary filling has been slow, as on the New South Wales coast (Roy 1984).

Rias show varying degrees of sedimentary filling, partly from inflowing rivers and partly inwashed from the sea. There are marshy deltas at the heads of the several branches of Carrick Roads in Cornwall, while other rias in southwest

England have banks of inwashed marine sand, particularly on the Atlantic coast, as in the Padstow ria, where at low tide the Camel River is narrow, flowing between broad exposed sandbanks that are submerged when the tide rises. Muddy sediment has been derived from periglacial deposits on bordering slopes.

Slopes bordering rias generally show only limited cliffing on sectors exposed to strong wave action, yielding sand and gravel to local beaches. Typically there are sandy, gravelly, and rocky embayed shores, some spits, muddy sediment and salt marshes at their heads, and sandy shoals at their entrances (Castaing and Guilcher 1995). The Rade de Brest is noteworthy for its several bordering sand and gravel spits (Guilcher et al. 1957). Finer sediment is deposited in fringing tidal marshes and mudflats and on the adjacent sea floor.

There is no clear distinction between a ria and an estuary, most rias being estuarine in the sense that inflowing rivers provide freshwater that meets and mixes with seawater moved in and out by the tide. Exceptions are found on Baltic coasts where salinity is low or absent.

Rias on high limestone coasts in the Mediterranean are known as calas or calanques (Fig. 3), while those on arid coasts are termed sharms or sherms. In Chile, there is a



Ria, Fig. 3 The calanque at Wied i Ghasri on the north coast of Gozo, Malta. (Copyright-Geostudies)

transition southward from rias to fiords with increasing influence of glaciation on valleys.

Summary

Rias are marine inlets formed by partial submergence of river valleys. Some definitions include relationships to geological structure, but these are not essential. Most are estuarine, but some occur on coasts bordering waters with low or no salinity. Many have been modified by sedimentation.

Cross-References

- [Dalmatian Coasts](#)
- [Estuaries](#)
- [Karst Coasts](#)
- [Salt Marsh](#)
- [Sharm Coasts](#)

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Rip Currents

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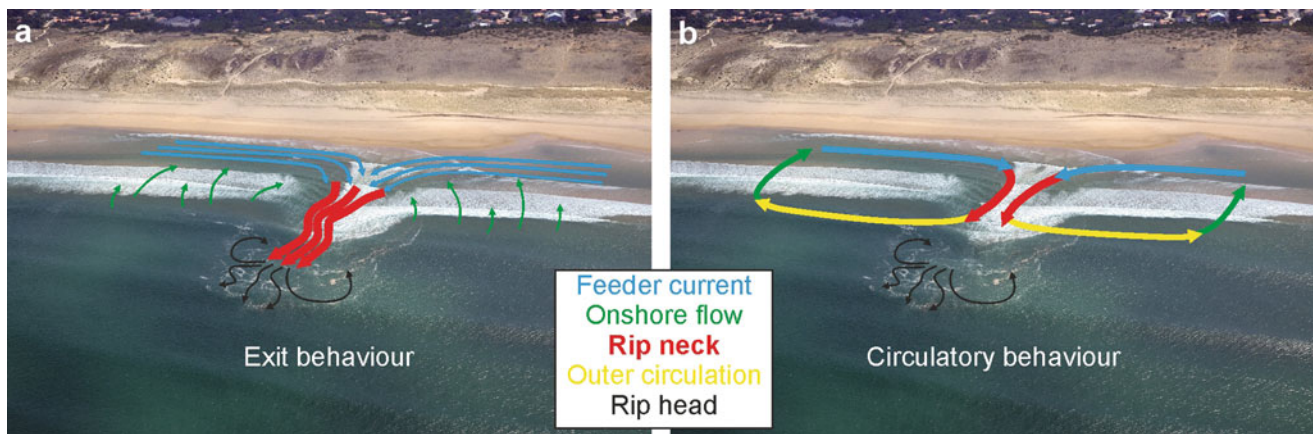
Definition

Rip currents are strong, concentrated, and often channelized seaward flows found on beaches that extend from close to the shoreline through the surf zone and varying distances beyond. They provide a major mechanism for the seaward transport of water and sediments; have a pronounced effect on nearshore morphology; can exacerbate shoreline erosion; aid in the dispersal and mixing of pollutants, nutrients, and biological species; and represent a major drowning hazard to recreational beach users.

Introduction

Many of the world's beaches, both oceanic and lacustrine, are characterized by the presence of rip currents driven by breaking wave activity across surf zones. The term rip current was first used by Shepard (1936) in order to distinguish rip currents from the conceptually misleading terms “rip tide” and “undertow,” which are unfortunately still commonly used to describe rip currents today. Early qualitative studies (Shepard et al. 1941; McKenzie 1958) suggested that rip currents (i) existed as a response to an excess of water built up on shore by breaking waves; (ii) often displayed a periodic longshore spacing; (iii) increased in flow intensity, but decreased in number as wave height increases; (iv) varied in location and flow intensity over time; and (v) flowed fastest at low tide. A significant amount of theoretical, field, numerical, and laboratory modeling and remote video monitoring studies have subsequently attempted to explain these characteristics with varying degrees of success. In particular, there has been a rapid increase in field and numerical modeling studies since 2000 that have greatly improved our knowledge of rip currents (Castelle et al. 2016).

The traditional view of nearshore cellular circulation was based on early Lagrangian measurements and observations made in rip currents at La Jolla, California (Shepard et al. 1941; Shepard and Inman 1950). Onshore mass transport of water due to breaking waves results in the formation of two converging longshore feeder currents close to the shoreline, which meet and turn seawards as a narrow, fast-flowing rip neck (Fig. 1a). The rip neck flows through the surf zone and variable distances beyond before decelerating and dissipating as an expanding rip head. The nearshore circulation cell is



Rip Currents, Fig. 1 (a) The traditional view of rip current system circulation patterns characterized by dominant exiting flow behavior beyond the surf zone. (b) A new view of rip current system circulation characterized by dominant recirculating flow behavior within the surf

zone (Reprinted from Earth-Science Reviews, 163, B. Castelle, T. Scott, R.W., Brander, R.J. McCarroll, Rip current types, circulation and hazard, 1–21, 2016, with permission from Elsevier)

completed when this water is transported shoreward again by breaking waves. More recently, an alternative view of rip current circulation based on new field measurements techniques has been proposed (MacMahan et al. 2010) where rip current flow is confined primarily within the surf zone in semi-enclosed vortices with only episodic exits of flow offshore (Fig. 1b).

In reality, both types of circulation patterns (Fig. 1) that can occur as rip currents are complex phenomena that can (i) exist both on planar beaches and those with three-dimensional surf-zone morphology; (ii) occupy distinct channels, flow against hard structures, or completely lack any morphologic expression; and (iii) be both persistent or transient in occurrence and location. This complexity is largely linked to a variety of different physical forcing mechanisms, controls and feedback relationships which combine to generate concentrated offshore flows in the nearshore (Castelle et al. 2016).

Formation

The primary mechanism behind the formation of rip currents is the presence of alongshore variations in breaking wave height and energy dissipation, which act to produce wave setup gradients that force water from regions of higher mean water level (higher waves, more intense breaking) to regions of lower mean water level (lower waves, less intense breaking). Bowen (1969) showed that these gradients are intrinsically related to spatial variations in wave momentum flux, or radiation stress (Longuet-Higgins and Stewart 1964) resulting from wave breaking. Alongshore variability of breaking wave height and energy dissipation has also been linked to the existence, strength, and rotational nature of surf-zone vortices, with length scales varying from several meters to

approximately 100 m, that are also a key element of rip current generation (Bruneau et al. 2011). The development of these vortices has been linked to vorticity generated at instantaneous time scales by individual short-crested breaking waves with along-crest variation in wave energy dissipation (Castelle et al. 2016).

Types

Several attempts have been made to classify the different types of rip currents (e.g., Short 1985; Dalrymple et al. 2011), often using different terminology to describe what are essentially the same rip current type. Castelle et al. (2016) describe three broad categories of rip current types (Fig. 2) based on the dominant physical forcing mechanism causing alongshore variability in time-averaged breaking wave energy dissipation. Each category was further divided into two types for a total of six different types of rip currents.

Hydrodynamically controlled rip currents are restricted to morphologically featureless, alongshore-uniform beaches, or planar sections of beaches, such as low tide terraces or seaward slopes of sandbars. They are unpredictable in space and time making them difficult to measure in the field and occur as two types. *Flash rip currents* are episodic bursts of water flowing offshore associated with transient surf-zone eddies driven by wave-breaking vorticity (Fig. 2a). They are short lived (several minutes), tend to migrate down-drift, and most commonly occur with shore-normal wave incidence and plunging breakers. *Shear instability rip currents* are much less common and only occur on alongshore-uniform beaches when exposed to highly obliquely incident ocean swells. Instabilities in the longshore current caused by shearing in the cross-shore direction can result in the generation of vortices leading to locally strong and narrow offshore flowing rip



Rip Currents, Fig. 2 Examples of different types of rip currents based on the classification of Castelle et al. (2016): (a) hydrodynamically controlled flash rip current at Rehoboth Beach, Delaware; (b) channel

rip current at Marina Beach, California; (c) a series of semi-regularly spaced channel rip currents along Tallow Beach, NSW, Australia; (d) a boundary-controlled rip current at Bondi Beach, NSW, Australia

currents. Visually, both types appear as turbid offshore-directed plumes of sediment and water.

Bathymetrically controlled rip currents are relatively persistent in space and time with the location and nature of morphologic control leading to two types. *Channel rip currents* are forced by primarily shore-normal waves and wave-breaking patterns across three-dimensional surf-zone morphology. As such, they are associated with intermediate beach states, occupying deep channels between adjacent surf-zone sandbars where wave breaking is less intense, or absent (Fig. 2b). Channel rips are the most well-documented and understood rip current type given their predictable location, relative logistical ease of field measurements, and common occurrence worldwide. They can be relatively stationary in position over temporal periods of days, weeks, and months. Their morphologic structure can be variable with channel depth, width, and alongshore spacing varying on scales of meters, tens of meters, and hundreds of meters, respectively, depending on incident wave conditions (Fig. 2c). In general,

the size and alongshore spacing of channel rip currents are related to regional wave climate, increasing with increasing wave height and energy. While numerous theoretical attempts have been made to account for the regular alongshore spacing of rip current channels, more recent studies involving long-term datasets of nearshore video imaging have shown that rip spacing is more often irregular (Turner et al. 2007).

Focused rip currents are also bathymetrically controlled and occur in fixed locations, but unlike channel rip currents, the alongshore variability in breaking wave height is related to the presence of offshore bathymetric features in the outer surf zone or inner shelf such as seaward bars (on multi-barred beaches), submarine canyons, and transverse ridges. Wave shoaling processes, particularly wave refraction, associated with these features result in opposing alongshore currents that deflect offshore as a rip current. These focused rip currents can shift location alongshore with different angles of wave approach and wave period and are not necessarily confined to channels.

Boundary-controlled rip currents occur adjacent to both natural and anthropogenic rigid features such as headlands, rock outcrops, groins, jetties, and piers and are persistent, often permanent, features depending on the local wave climate. They exist due to lateral bathymetric controls created by the boundaries themselves in relation to the angle of approach of oblique waves. *Shadow rip currents* occur on the down-wave (lee) side of a rigid boundary due to alongshore variations in wave height and energy dissipation due to the wave-shadowing effect of the obstacle. In general, shadow rip current activity increases with increasing wave angle, period, and wave height. *Deflection rip currents* occur on the up-wave side of rigid obstacles when strong alongshore currents are physically deflected seaward when they interact with the obstacle (Fig. 2d). Due to their more predictable nature and ease of replication, more studies exist on deflection, rather than shadow rip currents.

Despite these six primary rip current types, it is important to recognize that in reality, rip currents may form due to a combination of the forcing mechanisms described above, creating a potentially wide variety of mixed rip current types (Castelle et al. 2016). For example, rip currents associated with small surf-zone bathymetric anomalies may incorporate elements of both channel and flash rip currents. Inner bar rip currents on multi-barred beaches can exhibit characteristics of both channel and focused rip currents. There are also different types of rip currents specific to certain beach types. Large and powerful “megarips” that extend significant distances offshore can occur on highly embayed, or pocket, beaches during storm wave conditions. Rip currents have also been related to swash processes. These “swash” or “minirips” primarily occur on steep reflective beaches with pronounced beach cusp morphology where concentrated backwash in cusp embayments can force water various distances offshore of the wave step (Dalrymple et al. 2011).

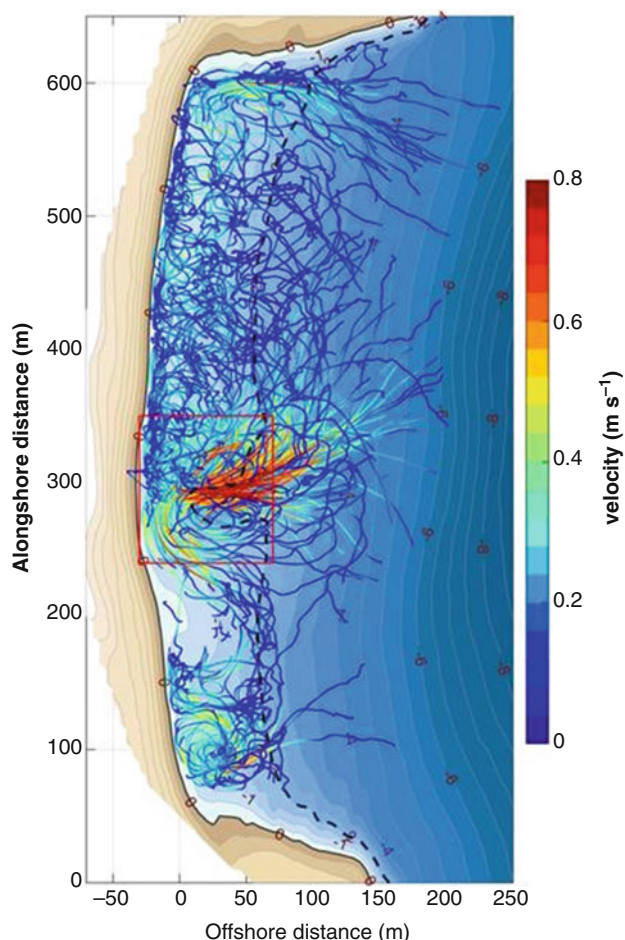
Flow Behavior and Characteristics

Our existing understanding of rip current flow behavior is largely based on field measurements obtained in channel rip currents (Fig. 2) due to the relative logistical ease of instrument deployment. Numerous studies have deployed one or more flowmeters at various locations within a rip current system providing Eulerian measurements of flow past a fixed point. In general, rip current flow velocities have been shown to increase with increasing wave height and also increase steadily from the alongshore feeders (where present) to the rip neck (Fig. 1), which is associated with the strongest flows (e.g., Brander and Short 2001). Under low-to-moderate wave energy conditions, hourly mean rip-neck flow velocities are typically on the order of $0.3\text{--}0.5\text{ ms}^{-1}$ with a maximum 1-min time-averaged velocities reaching 2 ms^{-1} (MacMahan

et al. 2006). Due to a lack of field measurements, less is known about flow velocity in other rip current types although boundary-controlled rip currents have been shown to exhibit similar velocities to channel rips (Scott et al. 2016). Flow velocities in flash rip currents in low-energy wave conditions have been measured at $0.2\text{--}0.3\text{ ms}^{-1}$ (MacMahan et al. 2010), while megarips and channel rips under higher-energy wave conditions may experience sustained mean flows in excess of 2 ms^{-1} (Short 1985). Measurements of vertical flow structure within rip currents are few (e.g., Sonu 1972), but show that velocity increases away from the bed, reaching a maximum just below the surface.

Rip flow is inherently unsteady in time and has been observed to pulsate at infragravity frequencies ($0.004\text{--}0.04\text{ Hz}$) and very low frequencies ($<0.004\text{ Hz}$) due to wave breaking associated with large wave groups and shear instabilities, respectively. These pulsations can accelerate instantaneous rip current flow by over 1 ms^{-1} for periods of $25\text{--}250\text{ s}$ (MacMahan et al. 2006). Channelized rip current flow has also been shown to be tidally modulated and related to tide range. On microtidal beaches, rip current flows are stronger around low tide when wave energy dissipation is maximized at shallow water depths, and sometimes cease at high tide (MacMahan et al. 2006). Morphologic constriction and lateral flow of water into rip channels from adjacent sandbars may also contribute to higher flow velocities in channel rip currents at lower water levels (Brander 1999). However, nearshore bathymetry on meso- and macrotidal beaches may enhance rip current flow at higher stages of the tide. On tideless beaches such as inland seas and lakes, rip current flow is solely dictated by episodic wind and storm-generated waves.

Lagrangian field observations and measurements involve tracking fluid trajectory and provide a greater two-dimensional spatial representation of rip current circulation and surface velocity patterns. While early studies of rip currents used simple Lagrangian techniques (Shepard and Inman 1950), the recent development of low-cost global positioning system (GPS) devices attached to Lagrangian surf-zone drifters (e.g., MacMahan et al. 2009) has allowed large numbers of drifters to be deployed in surf zones allowing for a much greater spatial and temporal coverage of rip current circulation patterns (Fig. 3). Field experiments using this approach (e.g., MacMahan et al. 2010) have contributed to the new view of dominant recirculation within the surf zone (Fig. 1b) characterized by counter- and clockwise rotating eddies that vary in velocity magnitude, shape, and orientation. However, despite evidence of dominant recirculation, exits of water and drifters beyond the surf zone do occur, often in episodic bursts. In channel rip currents, the primary exit mechanism has been shown to be very low-frequency motions (VLFs) and resulting eddies that detach from the main rip current. Under normally incident waves, drifter exit rates per hour have ranged from 0 to 20% increasing to 55%



Rip Currents, Fig. 3 Tracks and time-averaged velocities of approximately 30 GPS-equipped surf-zone Lagrangian drifters deployed over a period of 3 h at Whale Beach, NSW, Australia. Red box denotes a channel rip current in the middle of the beach. A deflection and shadow rip current are present against the top and bottom headlands, respectively (Modified from McCarroll et al. 2014)

with higher wave angles and up to 73% in macrotidal environments (Castelle et al. 2016).

Although field measurements in boundary rip currents are few, results are similar to modeling and laboratory studies, which show that deflection rip currents are typically associated with very high exit rates, on the order of 80% per hour. In contrast, shadow rip currents are characterized by dominantly recirculating behavior with much lower exit rates on the order of 20%. For both types, exiting flow behavior is maximized when the offshore length of the boundary structure (e.g., headland or groin) is greater than the surf-zone width (Castelle et al. 2016; Scott et al. 2016).

Summary

From a coastal science perspective, rip currents are an integral element of many surf zones and beaches around the world.

Significant advances in our understanding of rip current formation, flow behavior, and characteristics have been achieved in recent years due to improvements in field and modeling techniques. There has also been a large and recent increase in scientific attention into the hazard that rip currents represent to recreational beach swimmers (see “► [Beach Safety Research](#)”). Future efforts should be made in obtaining field measurements of flow circulation in a range of wave-tide-morphology conditions for rip current types that are less well understood and to better explore the role of rip currents in nearshore sediment transport and beach erosion.

Cross-References

- [Bars](#)
- [Beach and Nearshore Instrumentation](#)
- [Beach Processes](#)
- [Beach Safety Research](#)
- [Coastal Dynamics](#)
- [Lifesaving and Beach Safety](#)
- [Littoral Cells](#)
- [Natural Hazards](#)
- [Surf Zone Processes](#)
- [Vorticity](#)
- [Wave-Tide-Dominated Coasts](#)

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Ripple Marks

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Definition

Ripple marks have been defined in several ways: initially based upon morphology and metrics, and more recently on genesis. Merriam-Webster's dictionary defines ripple marks as "a series of small ridges produced especially on sand by the action of wind, a current of water, or waves." A more useful definition for marine scientists is that of Allen (1978) "...regular ridge-like structures, transverse to current, which arise and are maintained at the interface between a moving viscous fluid and a moveable, non-cohesive sediment (usually sand) by the interaction between fluid and transported sediment." Definitions based solely on morphology do not discriminate between what we "understand" to be ripple marks (small-scale rhythmic bedforms that almost anyone can identify on a beach) and genetically similar bedforms such as dunes, megaripples and sand waves. Indeed, there seems to be a continuum of morphology and size in both eolian and subaqueous bedforms from the smallest ripple (mm scale) to giant forms kilometers in length (Amos and King 1984; Ashley 1990; Ellwood et al. 1975; Wilson 1972). A common schematic illustration of a ripple form and the terminology used to describe it is shown in Fig. 1.

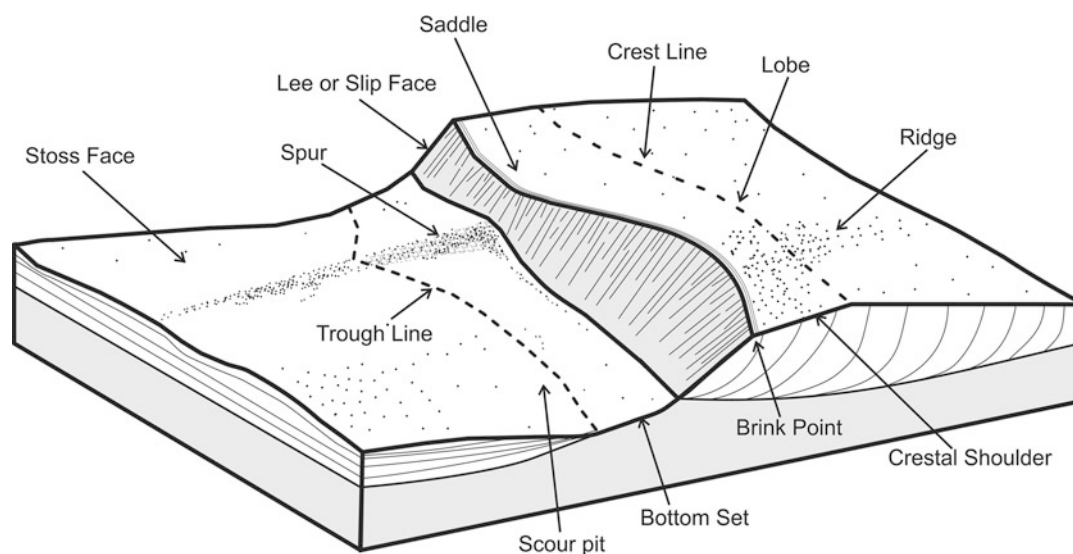
General Introduction to Ripple Marks

Ripple marks are perhaps one of the most widespread and well-recognized features of the sedimentary record

(http://www.seddepseq.co.uk/SEDIMENTOLOGY/Sedimentology_Features/Ripples/Ripples.htm). In the modern era, ripples are recognized easily (by almost anyone) by their scale, morphology, and setting. In the geological record, it is the characteristic internal structure of cross-lamination that defines them, and which is used extensively in paleoenvironmental reconstructions. They are documented to be present from the deep sea (Keijpers and Nielsen 2016), through to the continental shelf (Li and Amos 1998), into estuaries and beaches of the nearshore (Evans 1941; Allen 1962), and onto wind-swept, and arid sandy deserts (Bagnold 1941). Recently, they have been documented on snow fields (Filhol and Sturm 2015) and are also within tsunami deposits (Scheffers 2006). They have been shown to be a part of the landscape for as long as sedimentary deposits have existed; the oldest recorded being siliciclastic tidal deposits formed 3.2 Ga ago (Noffke et al. 2006). Recently, unique sets of wind-formed ripples, believed to have been created by a similar process that causes water-formed ripples on Earth, have been observed on Mars (Lapotre et al. 2016). The literature on ripple marks is immense and has increased steadily since the first publications of Darwin (1883) and Gilbert (1899, quoted in Johnson 1916). A search of Web of Science lists in excess of 3000 papers/annum referring to ripple marks since circa 2000. The majority of papers describe ripples from the geological record (largely based upon characteristic internal sedimentary structures) from which the depositional setting (water depth, wave climate, steady current magnitude, and direction) is inferred.

Ripple Types

Ripples are an expression of the organization of moving grains into morphological elements collectively known as bedforms. Ripple marks fall principally into two broad classes: Eolian ripples and water-formed ripples (Allen 1982). Eolian ripples are influenced by saltation bombardment at wind speeds above the threshold of motion of the sand in which they form to create ballistic ripples, which lack internal structure and have wave lengths related to saltation length (Bagnold 1941). Bagnold considered ripples to be constant in size with time, whereas other bedforms may grow in size with time almost without limit. Water-formed ripples fall into two broad categories signifying genesis: current-formed ripples and wave-formed ripples. Water-formed ripples are generally maintained (but not generated) by the interaction between the near-bed flow (the lower part of the benthic boundary layer) and the evolving bed roughness resulting from ripple genesis, in turn resulting in lee eddies and enhanced turbulence. They are maintained by lee eddy avalanching of sediment moving as bedload, which creates well-defined internal structure, and have wavelengths controlled by grain size.



Ripple Marks, Fig. 1 A schematic 3-D illustration of current ripple morphology and resulting internal structure

Early classifications were based upon the shape (with ripple type further subdivided into: straight, sinuous, linguoid, catenary, or lunate, and height, wavelength, asymmetry, and cord coherence) and planform (2-D, 3-D, bifurcation pattern, crest sinuosity) of the ripples. As a result, Inman (1957), Inman and Bagnold (1963), and then Allen (1968) recognized: short-crested ripples, intermediate-crested ripples, and long-crested ripples. Most published work on ripples is on forms considered in equilibrium with the forces that create them. Less is known about the mechanisms and bed interactions that result in the “order” of ripple formation that is produced from the “chaos” of the sediment transport phenomenon. Soulsby and Whitehouse (2005) undertook a comprehensive study of the metrics of ripple marks (height, wavelength, spacing, crest length, stoss (up-current) slope, lee (down-current) slope, and grain size) based upon over 3000 data sets, in order to examine trends and distinctions in ripple types. They also reviewed data on process-related studies of ripple type where the hydrodynamics and morphodynamics have been measured.

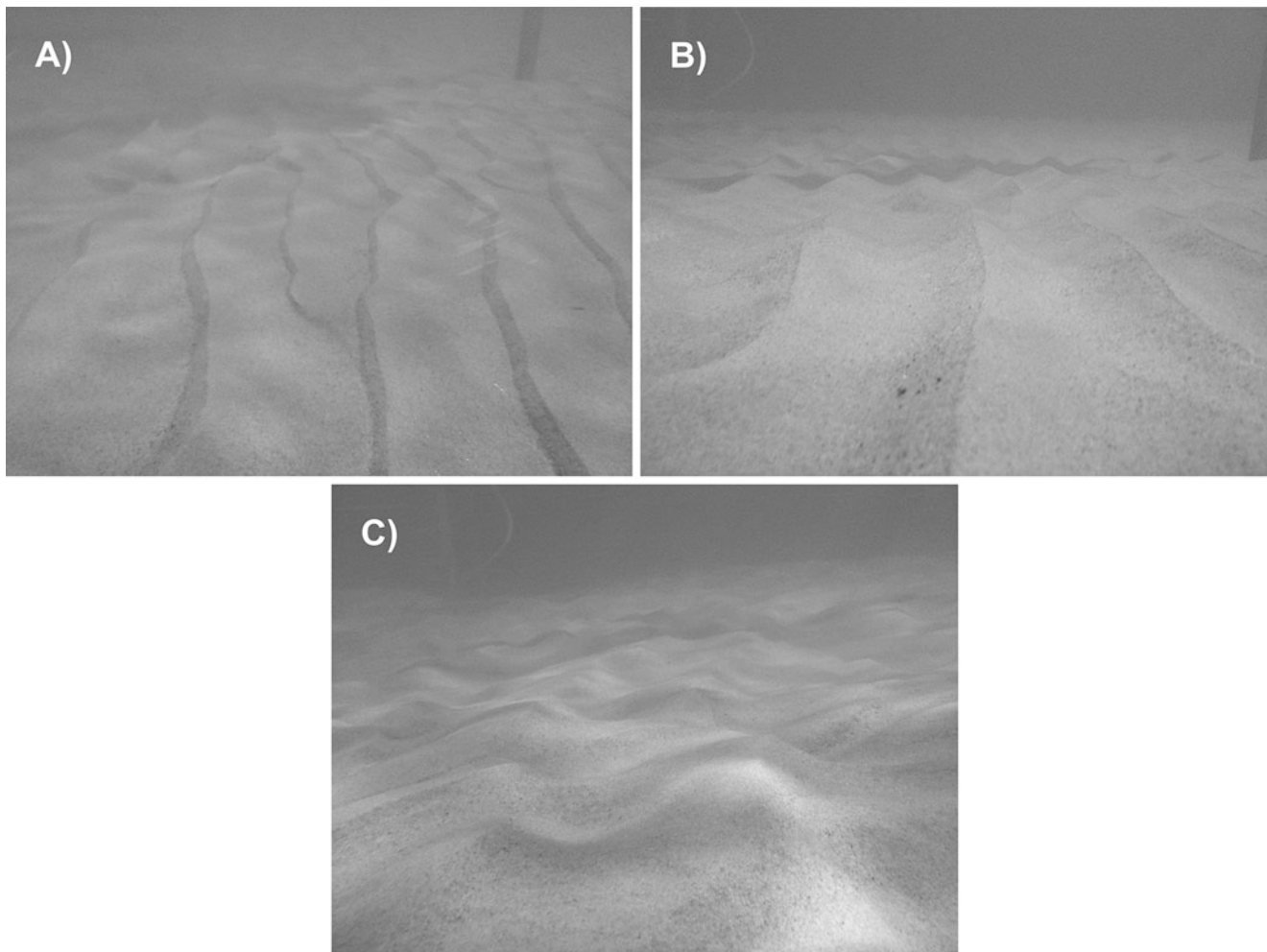
Essential Concepts and Historical Evolution in Understanding

Darwin (1883) described the evolution and migration of subaqueous ripples to be linked to vortices in the flow and subsequent sand transport. Classification of ripples based purely on metrics do not consider genesis and hence are deficient in two ways: (1) they cannot be modeled or predicted in advance from a knowledge of flow conditions, and (2) they cannot be used to hindcast conditions that formed them. Fundamental observations by Sorby and later Gilbert (1914,

quoted in Leopold 1980) showed that subaqueous ripples varied dependent on flow type and flow intensity, discriminating between steady and oscillatory near-bed currents or a combination of the two. Most of the ripples they examined were flow transverse; that is, the crest lines were oriented normal to the direction of flow. Exner (1925, quoted in Allen 1982) claimed that ripples begin as random irregularities of the bed that become self-maintaining and self-organizing due to perturbations of flow and resulting divergences/convergences in sand transport. Bagnold (1941) and others suggested that small-scale ripples were genetically different from large-scale ones that he termed “dunes.” Ripples were classified on the basis of genesis (rather than morphology) as (steady) current ripples, wave (oscillation) ripples, and combined form ripples (Allen 1982). The subsequent classification of large-scale bedforms and their distinction from ripples was proposed by Allen (1982). Ripples have become synonymous with sand transport as bedload and grain-to-grain interactions in the mobile bed layer (Bagnold 1946; Harnes et al. 1982; Middleton and Southard 1984). The rhythmic nature of ripples, according to Langbein and Leopold (1968), is due to the presence of kinetic waves in sediment flux not unlike stop-and-go congestion on a busy highway.

Current-Formed Ripples

An exhaustive literature review by Soulsby and Whitehouse (2005) has concluded that current ripples vary in height from 8 to 33 mm, vary in length from 0.08 to 0.8 m, have crest lengths that vary from 1 to 3 chord lengths, have lee slopes of 10–14°, stoss slopes of 7–9°, and migrate downstream at 0.3–0.4 mm/s. A photograph of current ripples (taken by the authors) formed in medium sand is shown in Fig. 2(a).



Ripple Marks, Fig. 2 (a) 2-D, bifurcating current ripples in medium sand. The ripples are migrating from right to left and have well developed crests and lee faces. (b) 2-D wave formed (vortex) ripples in

medium sand. (c) 3-D combined flow ripples in medium sand where the steady current is aligned with the direction of wave propagation

A seminal study on current ripples was published by Allen (1968), who brought together the knowledge of the day to clearly link morphology, dynamics of formation, and internal structure through vivid descriptions and illustrative diagrams. He proposed that current ripples occur in turbulent flows between the thresholds for traction and saltation/suspension on sand less than 600 microns in diameter. He showed that wave-formed ripples were largely symmetrical, sharp-crested, and either two-dimensional (2-D) in planform or brick-patterned. Tanner (1967) suggested that ripples were stable at Froude numbers between 0.2 and 0.6 and Reynolds numbers between 5×10^{-3} and 10^{-5} . The more complex combined-flow ripples show a superposition of forms that are highly three-dimensional (3-D) in planform (Amos et al. 1988; Arnott and Southard 1990; Southard et al. 1990). Allen (1982) reviewed and summarized the hydrodynamic research in the published literature on the near-bed processes forming unidirectional bedforms. He consequently constructed a variety of phase diagrams of bedform stability in terms of flow strength (power, pressure, or speed)

and grain size (diameter, dimensionless grain diameter, or grain Reynolds number). He showed that, while bedforms in sand were well understood, those in silt or gravel are less well known. Most of the relationships proposed ignored the ballistic contribution, important in the evolution of eolian ripples (Bagnold 1941), which has a counterpart in subaqueous flows. As well, the contribution of the enhanced bed roughness created by ripples (form drag) contributes to the total bed shear stress which feeds back into bedform evolution, migration rate (bedload transport rate), and to the suspension of sand.

Wave-Formed Ripples

Wave-formed ripples are usually symmetrical in planform, sharp crested, and form concordant (the internal structure mimics the ripple morphology; see Fig. 2b). They vary in height between 8 and 50 mm, in wavelength from 0.07 to 0.55 m, are trochoidal in cross section, and are largely 2-D (constant shape in the transverse direction). These ripples are usually stable in time. Bagnold (1946) and Manohar (1955)

undertook fundamental laboratory studies on the hydrodynamics of wave ripple formation, migration, and evolution. This was supplemented by field measurements off California by Inman (1957) and Inman and Bowen (1963) who showed a complex interaction between the motion of sand grains at the bed, the structure of the benthic boundary layer, the formation and dynamics of bed generated vortices, and ripple morphology. Bagnold recognized three distinct equilibrium forms: rolling grain ripples, vortex ripples (Fig. 2B), and post-vortex ripples; each form stable within a range of grain sizes and conditions of near-bed oscillatory flow. Inman and coworkers showed that these vortices were important in the suspension and transport of sand which had a direct link to cross-shore sand transport and hence shoreface evolution. Komar (1974) and Clifton and Dingler (1984) were able to show that the cord-length (crest-to-crest distance) of wave-formed ripples either scaled to near-bed wave orbital diameter (orbital ripples), or grain diameter (anorbital ripples) and that ripple type could be predicted from a knowledge of grain size, water depth, wave orbital diameter, and wave orbital peak velocity (Sleath 1984). Suborbital wave ripples have been defined as a class in transition between the other two cases. Bagnold's and Komar's works have significant implications for paleo-environmental reconstructions of sedimentary rocks where wave-formed ripples have been found (Allen 1981). A weakness of such paleo-environmental reconstructions is the link between wave-formed ripple type and internal sedimentary structure.

The link between ripple marks and resulting internal sedimentary structures came from the work of van Straaten (1954) and later Reineck and Singh (1966) on intertidal sand flats in the North Sea. They were able to show that the internal structure of ripples is a product of the migration of the ripple through the production of cross-laminae due to the pulse-like nature of accumulation on the lee face of a migrating ripple (see Fig. 1). The orientation of the cross laminae was thus a record of the direction of bedload transport and perhaps the magnitude of this transport. The interpretation of ripple internal structure has been examined by Harmes et al. (1982) and Rubin and Hunter (1987) who describe a complex suite of internal structures from (time) variable or invariable ripples. They describe forms transverse to the direction of sand transport, parallel to transport direction or oblique, resulting in either 2-D or 3-D cross-lamination or cross-bedding. The superimposition of differing types of internal structure punctuated by erosion surfaces separating foreset bundles appears to be the norm rather than the exception. The cause of superimposition of structures has been assigned to either a time varying flow (Allen 1978) or a complex benthic boundary layer (Rubin and Hunter 1987). Complex superimposed ripple patterns have been observed in nature due to the combined influences of waves and steady flows under storms (Amos et al. 1988; Arnott and Southard 1990).

These patterns are further complicated by the rotation of the flows due to rotational storm wind stresses leading to complex, polygonal transitional patterns in the ripple field (Allen 1982). As ripples in the modern setting are often the result of storm actions involving both waves and steady flows, a discussion of this class of ripple is warranted.

Combined Flow Ripples

The response of the seabed to combined wave-current flows has been observed in records of highly dynamic bedforms and active sediment transport across the inner continental shelf (Komar et al. 1972; Li and Amos 1998, 1999a, b)). Remarkably, there are more diagrams for assessing the likely bed state (of given composition) in combined wave-current flows that there are for waves only (Kleinhans 2005); but these tend to be limited in range of grain sizes and flow conditions (Perillo et al. 2014a). The geometry and ripple profile in combined flows are dictated by the relative strength and angle of the oscillatory and steady motions, and ripples are thus classified as wave-dominated, current-dominated, or transitional, with the boundaries between these regimes often chosen arbitrarily (Sleath 1984). For co-linear (near parallel, aligned or opposing) wave-current flows, ripple marks are asymmetric and are skewed on the current lee side with a flatter upstream slope and steeper downstream (Fig. 2c). Conversely, for currents at nearly right or large angles relative to the direction of wave propagation, wave (symmetric) and current (asymmetric) ripples coexist in a honeycomb (or cross-ripple) fashion with zig-zag bifurcations; often with regular, straight, and round-crest wave-ripples and irregular, curved, and sharp-crested ripples (Li and Amos 1999a). Straight-crested current ripples tend to occur at lower bed stresses and in weaker wave activity, while linguoid ripples may form immediately from degraded beds at higher wave-induced stresses (Amos et al. 1988). Short/small and large/long combined-flow ripples have been distinguished; both characterized by a 3-D planform with round crest, convex-up S-shaped profile becoming flatter with increasing wavelength, with scour at the toe of the stoss face; and hummocky cross-stratification evident particularly in finer sediment and for longer wave periods (Dumas et al. 2005). Cataño-Lopera et al. (2009) observed that combined, small wave-current ripples formed on the stoss-side of long ripples are much larger than those observed on the lee side, thus displaying a saw-tooth profile with a subdued form between the trough and crest. Highly dynamic, short ripples often imposed on longer ripples have been widely reported in combined, wave-dominated environments though the origins of the latter remain unclear (Hanes et al. 2001; Kassem 2016; van Rijn 2007). Moreover, the dependence of planform geometry on oscillatory period and sediment size on combined flows remains poorly understood (Perillo et al. 2014a, b).

A complex ripple pattern will be generated when combined influences of waves and unidirectional currents exceeds the threshold for sand transport. This threshold has been documented by Hammond and Collins (1979). The type of ripples and the way near-bed flow interact to produce these ripples is unclear. Amos et al. (1988) published a phase diagram for ripple generation in terms of the partitioned wave and current Shields parameter (θ) for field data collected on medium sand ($d_{50} = 0.23$ mm). They recognized, at times, current ripples (straight crested and linguoid) with superimposed subordinate wave ripples, wave ripples with superimposed subordinate current ripples, and well-developed wave and current ripples. Below the traction threshold, bedforms degraded to lower plane bed, whereas at the saltation/suspension, threshold ripples degraded in the “break-off region” of Grant and Madsen (1982). Large wave ripples were observed by Li and Amos (1999b) above the break-off region in association with sheet flow. These ripples were generally symmetrical, round crested, and 3-D in planform and are typically 50% of the near-bed orbital diameter that forms them. Wave-formed ripples and current ripples co-exist at dimensions that scaled faithfully to that expected were the flow purely steady or oscillatory. The current ripples migrated in the direction of the current, whereas wave ripples were stationary and at right angles to the waves. The wavelength of wave ripples (λ) scaled to the orbital diameter (d_o) relationship of Miller and Komar (1980): $\lambda = 0.65d_o$, as if there were no steady current. The onset of ripple formation took place at a combined Shields parameter: $(\theta_w + \theta_c)_{crit} = 0.04$. The waves and currents in this study were largely orthogonal and for a single grain size. It is expected that as waves and currents converge to be co-linear so too the bedforms would converge into a combined pattern of forms (such as that of Fig. 2c). Traykovski et al. (1999) undertook laboratory studies on combined flow ripples and showed that the angle between flows was important and that when flows were parallel, bedforms were 2-D in form. The height and subsequent roughness of ripples under combined flows has been examined by Li and Amos (1998) based upon data collected on the Canadian continental shelf. Their data constrained the upper limit for ripple development and the onset of larger hummocky megaripples above this threshold. The threshold Shields parameter at the break-off region (θ_{bf}) is defined by Grant and Madsen (1982) as: $\theta_{bf} = 1.8 \theta_{cr} S_*^{0.6}$, where θ_{cr} is the critical Shields parameter for bedload transport and S_* is the dimensionless sediment parameter. The data on combined flow ripples has been compiled by Soulsby and Whitehouse (2005) in order to derive predictive formulae for ripple development based upon the ratio of the wave to current Shields parameter to produce a time-stepping seabed predictor of ripple height, wavelength, and orientation in sandy sediments and is available as an Excel visualizer from these authors.

New Research on Ripple Dynamics

Recent advances in nondestructive technology, remote sensing, and more powerful computing have aided the proliferation of research into ripple marks and bedform dynamics. Large eddy simulations are increasingly being used to identify coherent structures formed by waves and combined wave-current flows over different ripple geometries (Grigoriadis et al. 2012, 2013), and to examine bed shear stress distribution along a ripple profile while multiphase modes increasingly capture flow dynamics over rippled beds, and the ensuing sediment entrainment and thus bedform evolution (Penko et al. 2011, 2013). High-resolution measurements of ripple morphology acquired through X-Ray computed tomography (CT) allows nonintrusive characterization of density profiles throughout the water-sediment interface, and visualization of the evolving ripple architecture during active sediment transport (Long and Montreuil 2011; Montreuil 2014). Long and Montreuil (2011) highlighted, through CT scans, the presence of a level of maximum bed density near the ripple surface, representing the boundary between bedload and suspension, that is dictated by turbulent vortices, suspended sediment, and the separation and re-attachment points of vortices. Coupled with particle image velocimetry (PIV) and more robust acoustic velocimetry, new insights into ripple formation and evolution and their role in sediment transport are being revealed. Such techniques could provide new insight into processes that have been elusive to date, such as the flux of water across the water-sediment interface, the sediment content entrained within a bedform-generated vortex/coherent turbulence structure, and the effect of these vortices on the structural integrity of the bedforms. Furthermore, the transition regime between laminar and turbulent flows, under which most ripples are generated, remains poorly understood (Charru et al. 2013), as are the effects of near-bed turbulent coherent structures on bed shear stress, sediment mobility, and bedforms.

Biostabilization of Ripples and Ripple Mark Preservation in the Geological Record

It has been recognized for some time that microalgae are able to stabilize the sediment-fluid interface by bed armoring (Neumann et al. 1970), or by the production of extracellular polymeric substances (EPS) (Frankel and Mead 1973). Stabilization is believed to occur due to a decrease in the bed roughness, hence fluid drag, or by the adhesion of sediment grains through the production of EPS (de Boer 1981; Paterson 1989). A reduction in the migration rates of diatom-bound wave ripples has been observed in the laboratory (Grant 1988), and microbial activity in modern, silty supratidal sediments has been related to sediment stabilization and the production of multidirectional ripple marks (Noffke 1998). In situ

evidence for the temporal and spatial scales at which microalgae are capable of stabilizing noncohesive sediments was presented in 2008, when the complete cessation of wave ripple migration for a period of at least 4 weeks was observed on a sandy intertidal flat in Chichester Harbour, southern England (Friend et al. 2008). Results of the study suggest that microalgae can effect a reduction in bed roughness on a spatial scale of 10^{-2} m, with a subsequent reduction in bed shear stress and bed erodibility. It appears that a unique combination of environmental forcing conditions, in conjunction with a microalgal bloom, caused the initial cessation of ripple movement, which was prolonged by the production of diatom- and dinophyte-derived EPS. It is reasonable to suppose that ripple stabilization by similar processes may have contributed to ripple mark preservation in the geological record.

Summary

The onset of ripple formation is as vague today as when we first wrote this article in 2005. Although there is a solid body of work defining the flow conditions under which the wide variety of ripple types form, there is still controversy over the role of ripple form drag on sediment transport rate and direction. Many believe that it is only the skin friction (caused by the grain roughness) that is important, while others believe that both forms of bed shear stress are important. There have been some advances in the action of ripple-formed lee eddies in the advection of sand into the water column (Kassem et al. 2015), and some credence to the notion that ripples *decrease* bedload transport and *enhance* suspended load, though the sum effect on the total transport is as yet unknown. Work has advanced on a quantitative definition of the bed roughness caused by ripples (Chiról et al. 2015) and how such descriptions of bed roughness are related to the near-bed hydrodynamic roughness. Finally, recent experimentation on the formation of current-formed ripples using a CT scanner has resulted in intriguing new insights into the bed density distribution through the migrating ripples, and the ventilation of fluids between the ripple form and the overlying flow (Montreuil 2014).

Cross-References

- Beach Features
- Beach Processes
- Coastal Sedimentary Facies
- Eolian Processes
- Longshore Sediment transport
- Seabed Roughness of Coastal Waters
- Sediment Suspension by Waves
- Shelf Processes
- Surf Zone Processes

- Tidal Flats
- Wave–Current Interaction
- Waves

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Rock Coast Processes

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Our ability to identify and measure the effect of rock coast processes has improved with the application of modern analytical techniques, geochronometric dating, and physical and mathematical modeling, but we are still largely ignorant of their precise nature (Trenhaile 1987; Sunamura 1992). The relative importance of rock erosional processes is often determined on the basis of ambiguous morphological evidence. Although the processes responsible for the slow lowering of rock surfaces have been inferred from micro-erosion meter data, the technique is unable to measure the dislodgement of large rock fragments by waves and frost.

It is difficult to obtain quantitative process data because of the imperceptible changes that generally occur on rock coasts within human lifetimes, the importance of storms and other high intensity – low frequency events, the lack of access to high and frequently precipitous cliffs, and the occurrence of exposed and often dangerous environments for wave measurement and subaqueous exploration. Changes in relative sea level and climate have also caused the nature and intensity of marine and subaerial processes to fluctuate through time, and because of slow rates of erosion, rock coasts often retain vestiges of environmental conditions that were quite different from today.

Mechanical Wave Erosion

Wave quarrying appears to be the dominant erosional mechanism in the vigorous storm wave environments of the middle

latitudes, based on the frequent occurrence of fresh rock scars and coarse, angular debris consisting of joint blocks and other rock fragments on shore platforms and at the foot of cliffs. Weaker waves in polar and tropical regions also play important roles, however, in eroding weathered rocks and removing loose debris.

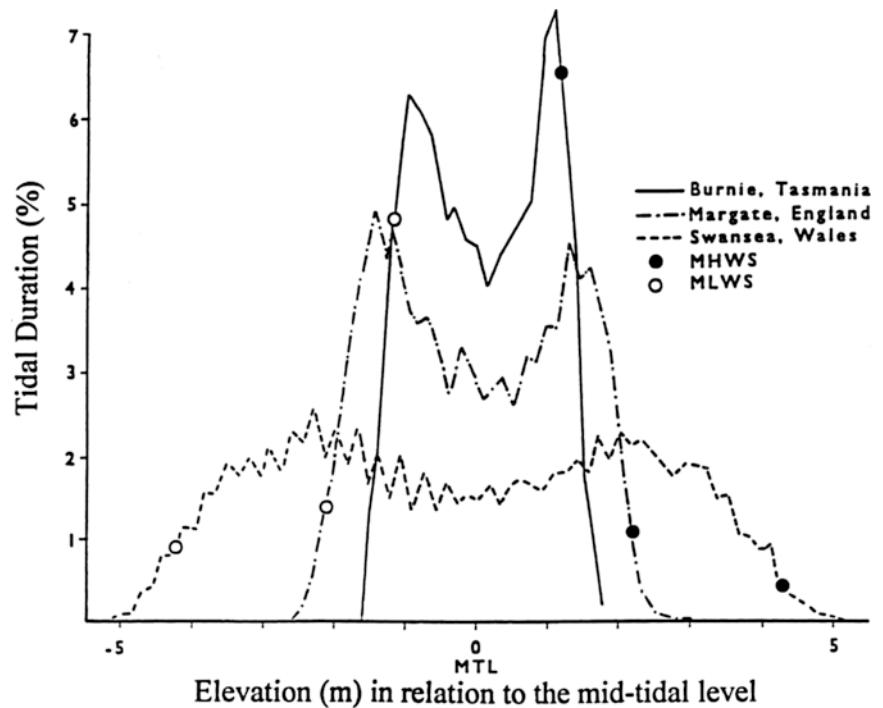
The forces exerted by waves on coastal structures depend upon their deep water characteristics, tidal elevation, and submarine topography. A broken wave may be a less effective erosional agent than a wave that breaks directly against a cliff or other steep, natural structure, but broken waves occur much more frequently. Therefore, the compression of air in joints and other structural discontinuities by broken waves is probably of much greater importance than the direct impact of waves on rocks (water hammer) and the generation of high shock pressures against near-vertical structures by breaking waves. Rock fragments and sand can be effective abrasional agents in the intertidal and shallow subtidal zones, although even large waves may be unable to agitate material sufficiently at the base of thick accumulations (Robinson 1977). Gently sloping abrasional surfaces are generally much smoother than wave quarried surfaces, but deep grooves can develop where abrasion is concentrated along joint planes and other structural weaknesses. Potholes are approximately cylindrical depressions that form where sand or large clasts are rotated by swirling water in the surf or breaker zones. They are particularly common in the upper intertidal zone where abrasives are trapped at the foot of scarps, and in structural or erosional depressions, and they also inherit, and subsequently modify, corrosional hollows in calcareous rocks.

Air compression, water hammer, and other processes responsible for wave quarrying require the alternate presence of air and water, and they therefore operate most effectively in a narrow zone extending from the wave crest to just below the still-water level. Most mathematical models also suggest that standing, broken, and breaking waves exert the greatest pressures on vertical structures at, or slightly above, the water surface. Wave erosion on a rock coast must therefore be greatest at the elevation that is most frequently occupied by the water surface. Over long periods of time, the tidally controlled water surface is most frequently at, or close to, the neap high and low tidal levels, and wave action is increasingly concentrated within the neap tidal levels as the tidal range decreases (Trenhaile 1987, 1997) (Fig. 1). Therefore, in controlling the elevation of the water surface, tides allocate the expenditure of wave energy and direct the work of mechanical wave erosional processes within the intertidal zone.

Although mechanical wave erosion is vertically distributed according to the tidal distribution of the water surface, it may be skewed toward the upper portions of the tidal range because of the occurrence of deeper water, and therefore

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Fig. 1 Tidal duration distributions – the amount of time that the water surface is at each intertidal elevation. MHWS and MLWS are the means of the high and low water spring tidal levels, and MTL is the mid-tidal level. (After Carr and Graff 1982)



higher waves, at high tide, and because large storm waves operate when sea level is meteorologically raised above the tidal level. In areas with low tidal ranges, this latter effect could elevate the zone of maximum erosion above the level of the high spring tides. Furthermore, as only vigorous storm waves, which operate at higher elevations than weak waves, are able to erode resistant rocks, whereas weaker waves, operating at lower elevations, are able to erode less resistant rocks, the difference between the level of greatest erosion and the most frequent tidal level may increase with the resistance of the rock (Trenhaile 1987).

Weathering

Coasts are particularly suitable environments for physical and chemical weathering owing to alternate wetting and drying in the spray and intertidal zones, and the presence of salts. It is normally assumed that physical weathering is most important in high latitudes and chemical and biological weathering in low latitudes, in part because of the occurrence of suitable climates, but also because of fairly weak waves in these areas. Weathering also weakens rocks in the vigorous storm-wave environments of the middle latitudes, however, and it may be an essential precursor for the dislodgement and removal of rocks by weak waves in sheltered areas.

Physical weathering mechanisms include the alternate expansion and contraction of clay minerals under cycles of wetting and drying, temperature-dependent adsorption of water, thermally induced changes in volume, and frost action.

The alternate expansion and contraction of clay minerals experiencing tidal- and weather-induced cycles of wetting and drying causes discontinuities to develop in shales and other argillaceous rocks, and also in the rocks that are adjacent to them. Clay minerals within small rock capillaries have negative charges that attract the positively charged ends of water molecules. This can breakup finegrained, clay-rich rocks, which adsorb water and expand as temperatures rise, and desorb and contract as temperatures fall (Hudec 1973). Temperature-dependent wetting and drying may be responsible for much of the field evidence that has traditionally been attributed to frost action, although several theories suggest that the two mechanisms could act together to generate more deleterious pressures within rocks than are generated by either mechanism acting alone. Thermally induced changes in volume also reduce the strength of some rocks. In southern Wales, for example, the expansion of limestones and the contraction of mudrocks under dry, hot conditions are responsible for diurnal variations in joint widths of up to 0.5 mm (Williams and Davies 1987).

Although much remains to be determined about the mechanisms involved in frost action, we do have a general sense of the conditions that are most suitable for their operation. Coastal regions may be almost optimum environments. High levels of saturation can be attained in the supratidal and intertidal zones, and because intertidal rocks can freeze in air during low tide, and thaw in water during high tide, they experience many more frost cycles than in areas further inland. Intertidal frost action may also be particularly effective because of rapid changes in temperature caused by the

sudden emergence and submergence of the rocks. Whereas rock temperatures rapidly increase when they are inundated in seawater, however, at least 5–6 h are needed for rocks to dissipate released latent heat and to cool to the freezing point of air. It is questionable whether critical levels of saturation can be maintained in the rocks over this period in the upper portion of the intertidal zone, and effective frost action in the lower portion of the intertidal zone may be inhibited by limited exposure to low air temperatures (Robinson and Jerwood 1987). Although the presence of salts in solution can inhibit frost action, several studies have suggested that the greatest rock deterioration occurs in solutions that contain between 2 and 6% of their weight in salt; this suggests that frost action may be particularly effective in rocks that are saturated with seawater.

Frost- and temperature-dependent wetting and drying can only be effective erosional agents where there are suitable rocks, and waves that are strong enough to remove the coarse debris and prevent progressive burial of the cliff. Tidally induced frost action is inhibited at high latitudes by low water temperatures, but water and air temperatures suggest that it may occur at various times of the year in cool temperate regions (Trenhaile 1987). Normal frost action, resulting from changes in air temperature, may also be more effective in the midlatitudes than in higher latitudes, where there are less frequent fluctuations about the freezing point. Atmospheric and tidally induced frost cycles are therefore probably most effective in cool, storm wave environments, and waves and frost also tend to be most effective on the same types of rock; strong wave action may therefore obscure or inhibit the effects of frost action in exposed areas.

Chemical and salt weathering are most important in warm temperate and tropical regions. Chemical weathering requires a good supply of water to promote chemical reactions, and more crucially, to remove the soluble products. Chemical reactions are accelerated by high temperatures in the tropics, but the lack of liquid water rather than low temperatures is probably primarily responsible for the fairly low rates of chemical weathering in high latitudes. Mechanical salt weathering occurs through the growth of crystals from solutions in rock capillaries, and crystal hydration and temperature-induced expansion. Chemical and salt weathering contribute to the formation of tafoni and honeycombs, the smoothing and lowering of shore platforms by the suite of processes collectively referred to as water layer leveling, case hardening, and the impregnation of joint planes by dissolved ions to form frameor box-like structures, and the formation of various types of weathering pits (Trenhaile 1987). The existence of a permanent level of saturation in the intertidal zone, separating a weak, weathered oxidation zone from a strong and largely unweathered saturated zone below, has been a basic tenet of Australasian workers for almost a century. Present evidence suggests that rocks can

only be permanently saturated below the low tidal level, however, where they are constantly submerged.

There is continuing debate over the processes responsible for the sharp pinnacles, ridges, grooves, and circular basins that are characteristic of coastal limestones in the spray and splash zones. Although these features are similar to karren formed by freshwater on land, surface seawater is usually saturated or supersaturated with calcium carbonate. It has been suggested that solution could occur in rock pools at night, when the carbon dioxide produced by faunal respiration is not removed by algae. Lower pH then causes calcium carbonate to be transformed into more soluble bicarbonate. Solution can be inhibited or prevented by other biochemical processes, however, including dissolved organic substances coating rock surfaces and building complexes with calcium ions. Although chemical solution does appear to be possible in seawater, recent studies have provided support for the contention that marine karren and other characteristic features of limestone coasts are primarily bioerosional in origin.

Bioerosion

Bioerosion is the removal of the substrate by direct organic activity. It is probably most important in tropical regions, where there are fairly weak waves and an enormously varied marine biota living on coral, aeolianite, and other calcareous substrates. A wide range of techniques are used to breakdown rocks. Microflora and fauna that lack hard parts may use only chemical mechanisms, but other fauna secrete fluids that chemically weaken the rock, before mechanically abrading them with teeth, valvular edges, and other hard parts.

Microflora bore into rock and they change the chemistry of the water that is in contact with it – indeed cyanophyta (blue-green) and other algae may be the most important biological agents on rock coasts. Algae, lichen, and fungi are pioneer colonizers in the intertidal and supratidal zones and they allow subsequent occupation by gastropods, echinoids, chitons, and other grazing organisms that effectively abrade rock surfaces as they feed on epilithic and the ends of endolithic microflora. Grazing organisms are of enormous importance in some environments. For example, it has been estimated that they are responsible for about one-third of the surface erosion in the mid-tidal zone on Aldabra Atoll where sand is available for abrasion, and as much as two-thirds where sand is absent (Trudgill 1976). At least 12 faunal phyla contain members that bore into rocks, especially in the lower parts of the intertidal zone. They include *Lithotrya* and other boring barnacles, sipunculoid and polychaete worms, gastropods, echinoids, *Lithophaga* and other bivalve molluscs, and Clionid sponges. Borers directly remove rock

material, and they also weaken the remaining rock, making it more vulnerable to mechanical wave erosion and weathering. In the tropics, carbonate rocks favor chemical borers, but mechanical borers are active on a variety of substrates in temperate and cool seawater environments. There is a great deal of published data on bioerosional rates of erosion (Trenhaile 1987), but they are of variable reliability and relevance. Nevertheless, most reported rates of erosion on vertical and horizontal limestone surfaces are between about 0.5 and 1 mm yr.⁻¹, which may reflect the maximum boring rate of endolithic microflora.

Ice

Until recently, it was generally believed that coastal ice is an ineffective erosional agent in the coastal zone, but its potential contribution to the development of rock coasts in cold environments is now being reassessed. The formation of subhorizontal shore platforms in the South Shetland Islands has been attributed to fast ice freezing to the underlying bedrock, quarrying by grounded ice and stranded ice rocked by the tides, and abrasion by rocks frozen into the ice base (Hansom and Kirk 1989). Ice-push, by wind-driven floating ice loaded with rock fragments, also assists gelifraction and frost wedging in quarrying gneissic joint blocks in macrotidal Ungava Bay. Much of the icefoot melts in place and does not contribute to debris removal, but it may facilitate deep frost penetration by providing a thermal barrier to sporadic warming by tidal water. Frost weathering and the effects of ice abrasion, dislodgement, and quarrying are also considered to be the main processes responsible for the formation of wide, subhorizontal platforms in the upper St. Lawrence Estuary (Dionne and Brodeur 1988). Although there is clear evidence of the erosive efficacy of shore ice in this area, weak slates and shales, high tidal range, strong currents, and large erratic blocks provide particularly suitable conditions for frost and ice action, and it remains to be determined whether these cold region mechanisms assume similar roles in less favorable places.

Mass Movement

Active marine cliffs possess short rather than long-term stability because of undercutting, oversteepening, and the removal of basal debris by wave action. Mass movements therefore play an important role in the development of cliffed coasts, and there is a close relationship between the morphology of cliffs and the type of mass movement that takes place on them. Mass movements range, according to local circumstances, from the quasi-continuous fall of small debris to infrequent but extensive landsliding. Although rock falls are more frequent than deep-seated slides, they are generally

much smaller. Falls occur in well-fractured rocks, especially where notches are cut into the cliff foot by waves, or, as in the tropics, by solution or bioerosion. Rock columns defined by joints or bedding planes also topple or overturn by forward tilting. Rock and slab falls, sags, and topples are essentially surficial failures induced by frost and other types of weathering, basal erosion, hydrostatic pressures exerted by water in rock clefts, and the reduction in confining pressures resulting from cliff erosion and retreat.

Deep-seated mass movements are triggered by groundwater build-up and basal undercutting. Translational slides usually occur where there are seaward dipping rocks, alternations of permeable and impermeable strata, massive rocks overlying incompetent materials, or argillaceous and other easily sheared rocks with low bearing strength. Slumps or rotational slides are common in thick, fairly homogeneous deposits of clay, shale, or marl. Sliding takes place in rocks that have been weakened by alternate wetting and drying, clay mineral swelling, or deep chemical weathering. Slides tend to occur during or shortly after snowmelt, or prolonged and/or intense precipitation. Water from septic systems, irrigation, runoff disruption, beach depletion through the building of coastal structures, and other human activities are playing increasing roles in some areas (Griggs and Trenhaile 1994). The damming of rivers has also reduced the bed load reaching the coast, depleting beaches and exposing cliffs to more vigorous wave action.

Cross-References

- ▶ [Cliffed Coasts](#)
- ▶ [Cliffs, Erosion Rates](#)
- ▶ [Cliffs, Lithology Versus Erosion Rates](#)
- ▶ [Karst Coasts](#)
- ▶ [Mass Wasting](#)
- ▶ [Notches](#)
- ▶ [Shore Platforms](#)
- ▶ [Weathering in the Coastal Zone](#)

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Salt Marsh

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Salt marshes or saline wetlands are vegetated intertidal flats dominated by low-growing halophytic (salt-tolerant) shrubs and herbaceous plants, particularly grasses. Typically, salt marsh borders freshwater or brackish environments. Largely confined to temperate coastlines, they occupy a similar niche to tropical mangrove forests; that is, the upper intertidal zone of inlets, estuaries, lagoons, and embayments, or fronting the open sea where low-energy conditions persist (Frey and Basan 1985). In warm temperate, subtropical, and some tropical regions, salt marsh and mangrove communities sometimes intermingle, but can be separated by definition on the basis of floristics or intertidal position.

Salt marsh originates with the spread of vegetation onto an accreting intertidal mudflat. Fine suspended sediments (silts and clays) and organic material washed in by tides, and subsequently trapped by roots of salt marsh vegetation, generate a gently sloping depositional terrace or platform between the high spring tide level and the mid-tide line (Bird 2000). At a smaller scale, characteristic features of salt marsh include tidal creeks, levees, cliffs, and salt pans. The patterns and morphology of these features are collectively dictated by a number of factors: the extent of vegetative cover; the climatic, hydrographic, and edaphic influences on this vegetation; the availability, composition, deposition, and compaction of sediments; organism–substrate relationships; topography; tidal range; wave and current energy and stability of the coastal area (Frey and Basan 1985).

The salt marsh platform is typically dissected by a system of meandering creeks and levees that channel the ebb and

flow of tidal waters. Wide shallow creeks are common where the salt marsh is in an early developmental phase or where sediments have high sand content, but as the marsh platform rises or expands with continued accretion, well-defined steep-edged channels become more commonplace. The seaward edge of a marsh may grade smoothly into adjacent tidal flats, or it may end abruptly in a small vertical cliff up to several meters high, or the transition may be marked by an irregular series of channels separated by ridges (Haslett 2000; Ke and Collins 2002).

The wide variety of sediments underlying coastal marsh systems reflect the range of adjacent terrestrial and marine habitats from whence these originate. The shallow waters and low-velocity currents characteristic of marsh surfaces promote deposition of fine-grained inorganic sediments such as mud, clay, silt, and fine sands, and rarely gravels. Most are reworked and modified by roots of plants and the action of animals, leading to the deposition of organic components, including peats which are formed from the degradation of roots, stems, and leaves of marsh plants and animal skeletons including shells and carbonate tests (Chapman 1977; Dawes 1998).

Salt marshes support an abundance of organisms of both terrestrial and marine origin. They are highly productive ecosystems, but the extreme physiological and ecological stresses of the intertidal environment maintain characteristically low species diversities. Clear zonation patterns of vegetation and associated fauna are generally evident, influenced by factors such as salinity (which ranges from near ocean strength to near fresh in most systems), frequency and duration of tidal exposure, and climate (Haslett 2000). Low-marsh communities toward the seaward edge normally comprise pioneering halophytes such as sea grasses (e.g., *Zostera*) and glassworts (e.g., *Salicornia*) through to those tolerant of brackish conditions (e.g., *Spartina*). High-marsh communities, being relatively more stable and terrestrially influenced, include more diverse assemblages of less salt-tolerant species such as rushes (e.g., *Juncus* spp.), grasses (e.g., *Puccinellia*, *Sporobolus*), and herbs (e.g., *Aster*, *Plantago*).

Terry R. Healy: deceased.

Vegetation cover of a salt marsh surface does not usually follow a continuous sequence from sea-edge to land. In addition to creeks and channels, highly saline, dry, or water-filled shallow depressions often feature in the surface of the marsh. They form where vegetation has either failed to establish, or where it has died back, or where drainage channels have slumped and blocked. Strong evaporation from the resulting pools of water or bare substrate surfaces advances the accumulation of salts, inhibiting seed germination and further establishment of plants, and giving rise to the conspicuous vegetation-free areas known as salt pans (Bird 2000).

Cross-References

- Coastal Soils
- Dikes
- Estuaries
- Hydrology of the Coastal Zone
- Mangroves, Ecology
- Mangroves, Geomorphology
- Peat
- Reclamation
- Tidal Creeks
- Tidal Flats
- Vegetated Coasts
- Wetland Restoration
- Wetlands

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Sand Mining

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Definition

Sand mining is the general term for the extraction of sand or aggregate from the continental shelf. Sand, from all sources,

is one of the two most utilized natural resources. Construction aggregate, beach nourishment, and fill material are its major uses. Land use competition, transportation costs, and environmental regulations have motivated a shift from subaerial to submarine production.

Introduction

The mining of sand and aggregate from submarine deposits is a substantial enterprise across the world. The material is excavated for three general purposes: construction aggregate including concrete and asphalt, beach nourishment including dune building and similar forms of shore protection, and fill material. Lesser quantities of sand are mined for their mineral content, e.g., titanium, silica, garnet, for various industrial purposes, and for numerous other niche uses.

According to Owen (2017), sand is second only to water in its exploitation as a natural resource. Peduzzi (2014) reported “The absence of global data on aggregates mining makes environmental assessment very difficult and has contributed to the lack of awareness about this issue.” According to Morelock et al. (2005) Japan produces more aggregate from marine sources than any other nation taking as much as a fourth of its sand and gravel from underwater. Johns (2010) confirms that statement and, citing others (Tsurusaki et al. 1988; Morimoto 1985), added that as much of 40% of the fine aggregate used for concrete is extracted from marine sources. Continuing to cite Tsurusaki et al. (1988), Johns (2010) reported that over 70×10^6 tons of marine aggregate were produced in 1988 in Japan. Bokuniewicz (2000), citing others, reported that annually Japan mines 80 million tons (44 million m^3) making that nation the world’s major producer of aggregate from offshore resources. As is the situation elsewhere, competition for land use and enhanced environmental regulations make onshore sources more difficult to exploit (Morelock et al. 2005).

The Working Group on the Effects of Extraction of Marine Sediments on the Marine Ecosystem (WGEXT) of the International Council for the Exploration of the Sea (ICES) provides annual data on the production of marine sand and gravel for much of Europe, the USA, Canada, Greenland, and Iceland. The most recent report (ICES 2016) indicates that $79 \times 10^6 m^3$ of aggregate were extracted in those nations during 2015. The bulk of that work was done in the Netherlands ($26.1 \times 10^6 m^3$), the UK ($19.5 \times 10^6 m^3$), the USA ($9.5 \times 10^6 m^3$), and Denmark ($7 \times 10^6 m^3$). According to the ICES (2016) report, 44.7% of the total was used as construction or industrial aggregate (assumedly concrete), 43.7% went to beach replenishment, 7.0% to construction fill and land reclamation, and 7.9% was exported. Bokuniewicz (2000) stated that marine deposits satisfy 15% of Great Britain’s need for sand and aggregate.

Historically, sand and aggregate have been mined from upland deposits (and this continues to be the case in areas distant from submarine sources). For several reasons, this has become problematical. Especially in areas with dense human population, there are competing uses for the land. Development, whether residential, commercial, or industrial, likely has a higher value for the land owner. The mining excavation leaves an unnatural depression. The term *sand pit* is apt and often is applied to the surface sand mines. *Borrow pit* is another commonly used term. In most circumstances, the removal of the material also removes much of the value of the real estate. The land no longer is suitable for development. If the walls are not properly graded, the pit is a safety hazard; and if the pit even partially fills with water, the hazard is even greater. In some instances, it might be possible to turn the site into an artificial pond, possibly enhancing the value of the adjacent property for residential development. But, in general, once the sand resource is exhausted, only problems remain. Also, to maintain economic viability, it usually is necessary for the borrow pit to be continually active. When the site is not being mined, the local fixed equipment is unused and there is no income to cover the ongoing cost of real estate taxes. Finally, environmental and other regulations have come to require more from the mine operator including, in many places, some sort of restoration/reclamation of the mine site. Indeed, the restoration might itself require fill material excavated from another site.

Transporting the sand, especially large quantities, from sand pits presents another set of problems. In rare occasions, when the borrow site is close to processing plant or the target beach, in the case of beach nourishment, it might be possible to use a system of conveyor belts. In some locales, piping a sand-water slurry directly to the beach might be feasible. This would require a large quantity of water and, if the water were fresh, would render it unsuitable for more beneficial uses. Also, it would be necessary to handle the water discharge in an environmentally appropriate manner. In a few circumstances where the sand pit is adjacent to a navigable water body, it might be possible to use the benefits of scale and barge the sand to the processing plant or target beach. The same is true for deposits adjacent to rail service.

More usually, however, it will be necessary to truck the sand or aggregate. A large dump truck might carry 15 m^3 (20 yd^3) (Earth Haulers, Inc. 2017), so a beach nourishment or fill project requiring $76,000 \text{ m}^3$ ($100,000 \text{ yd}^3$) would require about 5000 truckloads. In addition to the increased congestion, this level of traffic likely would displease residents and businesses and along the route and would cause increased wear on the roadways. Finally, as sand weighs roughly 1600 kg/m^3 (2700 lb/yd^3), the weight of a fully loaded 15 m^3 truck might exceed the allowable load on the roadway, it might be necessary to use smaller, 10-wheel

trucks carrying 7.6 m^3 (10 yd^3) (Engineering Forum 2017) which would double that number of trips. Nourishment projects in excess of $760,000 \text{ m}^3$ ($1,000,000 \text{ yd}^3$) are relatively common. A concrete plant would need a steady stream of aggregate. Owen (2017) suggests that 100 km (60 mi) is a practical maximum distance for hauling by truck.

The increasing costs and difficulties of using and the depletion of subaerial deposits contribute to the increased exploitation of the offshore resources. Also, because the seabed is “owned” by some level of government, there are no continuing costs of real estate taxes. In the USA, each marine mining episode usually is permitted individually. Beach nourishment occurs (a)periodically at any one locale; the dredging equipment is highly mobile and can shift from site to site. The sand or aggregate can be transported from the source area to the plant or beach using any of a very few methods, primarily depending on the distance between the source area and the target site.

Submarine Sand Deposits

The decision to use submarine resources is predicated on the existence of an accessible sand body. Hobbs (1997, 2007) described three general types of marine sand deposits: shoals, filled channels, and stratigraphic. Finkl and Andrews (2007a, b, 2008) and Finkl and Hobbs (2009) provided a more detailed categorization of shelf sand deposits (Table 1).

Prospecting for marine sands and aggregates is little different than prospecting for other geological resources. As described by Finkl and Hobbs (2009), prospecting usually happens once a need has been identified and begins with a review of the regional literature. Unfortunately, much of the information often is available only in unpublished technical reports such as those prepared in district offices of the US Army Corps of Engineers and local agencies. The literature, especially when coupled with study of pertinent bathymetric data and consideration of the types of deposit (Table 1), often can rule out some areas and highlight others for further study. High resolution sub-bottom seismic surveys, sometimes run simultaneously with sidescan sonar surveys, can further indicate likely deposits. Finally, the potential deposits need to be proved with cores.

Should there be multiple viable sources in a specific region, the selection of the one to be exploited involves weighing several factors including distance from the processing plant or beach and water depth. Transport distance has a direct connection to cost. The bottom sediments at sites of past excavations in shallow water, above a wave base with a relatively frequent return interval, will be agitated which could speed the site’s recovery.

Sand Mining, Table 1 Characteristics of shallow, marine sand bodies (Modified from Finkl and Hobbs 2009)

Terminology Finkl and Andrews 2007a, b, 2008	Terminology Hobbs (1997, 2007)	Conspicuousness	Lateral limits	Top	Bottom	Seismic definition	How found	How proved	Mining characteristics
Deltas	Shoal	Good, on surface	Sometimes indistinct	Sea floor, some buried	Usually sharp	Good	Geological knowledge, morphology, setting	Coring	Moderate
Bar systems	Shoal	Good, on surface	Sharp	Sea floor	Usually sharp	Good	Bathymetry	Coring through bottom reflector	Relatively easy, dredge the shoal
Paleoshores	Shoal	Fair, on surface, deeper water	Sometimes indistinct, muted	Sea floor	Fair	Fair	Bathymetry, correlation	Seismic mapping, coring	Moderate, prior mapping to set spatial limits
Ridges etc.	Shoal	Good, on surface	Sharp	Sea floor	Usually sharp	Good	Bathymetry	Coring through bottom	Relatively easy, dredge the topography
Sheets and veneers	Stratigraphic- gradational	Poor	Poorly defined	Sea floor, varied	Varied, vague	Fair to poor	Chance	Coring, sidescan sonar for extent	Moderate to poor. Thin. Prior mapping to set spatial limits
Storm deposits	Stratigraphic- gradational	Poor	Poorly defined	Sea floor	Varied	Poor	Morphology, chance	Coring, sidescan sonar for extent	Moderate to poor. Thin. Prior mapping to set spatial limits
Inter-reef flats	Stratigraphic- gradational	Fair	Usually clear	Sea floor	Varied	Fair	Location	Coring, sidescan sonar	Moderate, care needed to protect reef
Reef gaps	Stratigraphic- gradational	Good	Usually clear	Sea floor	Varied	Fair	Location	Coring, sidescan sonar	Moderate, care need to protect reef
Paleochannel fills	Filled channel	Poor, buried	Sharp	Varied	Usually sharp	Good	Geologic knowledge, chance	Coring, seismic mapping	Poor. Need to stay within tight lateral limits above a specified depth
Karst fills	Filled channel	Poor, buried	Sharp	Varied	Usually sharp	Good	Geologic knowledge, chance	Coring, seismic mapping	Poor. Need to stay within tight lateral limits above a specified depth
—	Stratigraphic facies	Poor	Undefined	Varied	Vague	Poor	Chance	Coring	Easy. Stay within broad boundaries

Sand Mining, Table 2 Main effect of mining marine aggregates (Modified from Tillin et al. 2011)

Effect	Direct or indirect	Spatial scale	Temporal scale	Confidence in evidence
Removal of aggregate	Direct	Impacts on benthic marine organisms and morphology. Limited to dredged area	Recovery may begin after active dredging stops	Very good evidence of the impacts on habitats and biological assemblages available in numerous sources
Removal of aggregates	Direct	Impacts on cultural and archaeological materials	May be permanent and irreversible	Good
Formation of sediment plumes	Direct	Sand deposition with 500 m (0.3 mi) of dredge and up to 3 km (2 mi) for particles remobilized by local conditions such as tidal currents	Plumes of fine sediments can last 4–5 tidal cycles	Very good evidence from numerous studies
Visual disturbance	Indirect	May affect seabirds and marine mammals. Level of the impact depends on the visual acuity of the animal	Extends slightly beyond the duration of the dredging	Not well studied
Noise disturbance	Indirect	Noise from the dredging may project several (sometimes many) km from dredge site. Behavioral response varies among animal species	Confined to the period of dredging	An active area of research
Risk of collision	Indirect	Limited to site of activity	Confined to the period of dredging	Little evidence
Sediment deposition		Sand deposition with 500 m (0.3 mi) of dredge and up to 3 km (2 mi) for particles remobilized by local conditions such as tidal currents. See sediment plumes, above	Most sands settle almost immediately, most smaller particles settle within a full tidal cycle	Well studied with both direct observations and models

Environmental Concerns

Before an identified resource can be exploited, the environmental consequences of mining must be considered and weighed against the potential benefits of the proposed mining effort. Before being permitted, the proposed work must comport with the national and local environmental regulations. Johns (2010) provided a good overview of the environmental review processes in the United Kingdom, the United States, and Japan. Tillin et al. (2011) presented a description of the scales of the environmental effects of mining marine aggregate deposits (Table 2).

Pilkey and Cooper (2014) present a discussion of several of the problems tied to offshore sand mining. Hobbs (2002), Hobbs et al. (2003), and Diaz et al. (2004), among others, review the environmental consequences of offshore sand mining. Those consequences almost always are detrimental to the local marine ecosystem and begin with the death of all organisms in the mined sediments. The removal of the benthic infauna, in turn, is local destruction of food for many fish with the potential to harm the commercial and sport fisheries. The mechanical actions of the dredge alter the physical characteristics of the natural habitat, again with potential harm to bottom dwelling organisms. In tropical regions, the dredge can harm corals while the sediment resuspended by the process of dredging can smother the living coral. It has been suggested (Finkl and Hobbs 2009) that when dredging a relatively large area, leaving some undisturbed areas might shorten the time it would take for organisms to move into the disturbed area as the “refuge patches” might serve as loci from

which the bottom dwelling fauna and infauna might spread. Although not a problem of marine sand mining, as such, it should be noted that there is increasing discussion of negative environmental effects of beach nourishment.

Depending on the magnitude of the sand mining, the alterations of the submarine topography can modify the local pattern of wave refraction, changing the distribution of wave energy along the shore and across the bottom (Maa and Hobbs 1998; Byrnes et al. 1999; Maa et al. 2001, 2004). The character of the change depends on whether the dredging either removes a topographic high, a shoal, eliminating the concentration of wave energy caused by refraction around the shoal, or creates a depression which would allow locally faster and, perhaps, higher waves. If the excavation is held to a relatively small surface area, the resultant depression can trap fine sediments resulting in a habitat different from that which existed prior to the disturbance. Similarly, if the dredging covers a larger area, bottom likely would be furrowed, again, altering the habitat; though the furrows would be reduced with the passage of storms.

The process of marine sand mining can have environmental consequences that occur at a distance from the actual mining site. The noise of the machinery can disrupt the travels of fish and marine mammals and any plume of fine-grained sediment could envelope sessile and planktonic organisms.

Because there will be adverse environmental consequences to sand mining, it has become the general practice to require pre- and postmining surveys to assess those consequences. In some locales, it is appropriate that the premining assessment continue for over a year to capture the full suite of

organisms that utilize the area and might be harmed by the operation. The postmining assessments might be either short term, merely to document the changes resulting from the mining, or long term, designed to track the “recovery” of the disturbed area. Additionally, it is necessary to access many years of data about the local waves in order to effectively model potential changes to wave refraction. Research Planning, Inc. et al. (2001), Hobbs et al. (2002), and Hobbs (2006) describe a proposed standard protocol for monitoring and evaluating the impacts of marine sand mining.

Methods of Production

While clam-shell and other mechanical dredges are used to mine marine sands, hydraulic dredges are more common. The



Sand Mining, Fig. 1 Photograph of a hydraulic suction cutter head aboard ship. The flat surface drags on the bottom while jets of sea water mobilize the surface sediment and the resultant slurry is pumped to the surface

mechanical dredges provide only a small volume of material with each cycle. The dredge platform usually needs to be secured with anchors or temporary pilings when active. When the sediment within usually short reach of the dredge has been extracted, the platform must be moved. In contrast, many hydraulic methods, such as a trailing arm hopper dredge, produce a continuous stream of sand-laden slurry, at least until the hopper is full. With a hydraulic dredge, the cutting head (Fig. 1) is dragged along the bottom while jets of sea water are directed at the bottom to agitate it. The sand-water mixture is pumped to the surface where it is discharged into the ship's hold (Fig. 2a, b) or into barges. After a barge is full, it is towed to the processing plant or other discharge site. When the hold of the hopper dredge is full, dredging is discontinued while the vessel transits to a discharge site. Often this is a moored pumping station. Again, the hopper dredge uses ambient sea water to pump the cargo of sand to the pumping station, which, in turn, pumps the slurry to shore. In some cases, the hopper dredge might pump the aggregate directly to shore.

Although the excavation methods are identical, there is an important difference between *sand mining* and the beneficial use of (sandy) dredge spoil. In the former case, there is a need to prospect, to define the resource, and to go through the process of obtaining permits for mining. In the latter, the permits have been granted for the purpose of deepening a channel or the like, and the dredge spoil might be put to use. In dredging, the excavation is the product, not the sediment. Indeed, using the dredge spoil, if possible, yields two benefits: (1) whatever constructive purpose the material serves and (2) the abrogation of need to find a site for disposal of the spoil. Often a problem of (channel-)dredging is finding a suitable disposal area for the dredged material.



Sand Mining, Fig. 2 (a) and (b). Two views of a hopper dredge while the hold is being filled with the sand-water slurry pumped from the bottom

Summary and Conclusions

Marine sand mining is a major enterprise around the world. The sand and aggregate are used for construction, including concrete and asphalt, beach nourishment, fill material, and several other purposes. Especially when compared to the resources on shore, the marine resources benefit from the economies of scale and relative ease of transportation. Although marine sand mining once functioned under an “out of sight, out of mind” mind set, concerns for the environmental consequences of the activity now are substantial. Those consequences occur during the dredging both at the immediate site in the general locale and may continue, especially at the dredge site, for several years.

Cross-References

- Beach Nourishment
- Dredging of Coastal Environments
- Sand Rights

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Sand Rights

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Introduction

Sand Rights is a concept that merges the physical laws of sediment transport with societal laws of public trust. The basis doctrine is that human actions will not interfere, diminish, modify, or impede sand and other sediments from being transported to and along rivers, beaches, shores, or any flowing or windblown paths or bodies without proper restitution. Under this doctrine, projects should be designed or reevaluated to mitigate any interference that the project may have with sand transport.

As early as Justinian, cultures have understood that certain natural resources are incapable of private ownership. “By the law of nature these things are common to mankind – the air, running water, the sea and consequently the shores of the sea.” (Institutes of Justinian 2.1.1, quoted in *National Audubon Society v. Superior Court* (1983) 33 Cal.3d 419, 433–34.) Traditionally water, navigation, fisheries, and tidelands had been covered by the public trust doctrine. In certain cases, states have extended the public trust doctrine to marine resources. The basis doctrine of Sand Rights would establish the application of public trust to inland and coastal sand resources.

Physical Principles

The coast is a dynamic area – the junction of the land, the air, and the sea. The coastline can be divided into a series of

littoral cells or segments to contain one or more sources of sand, mechanisms for moving sand along the coast, and a sink from which the sand cannot be carried back onshore by wave action. On the western US coast, headlands, rivers, and submarine canyons can often delineate the cell boundaries. On the eastern US coast, the littoral cells are defined differently, often as reaches between inlets, but the same processes apply. Similarly, littoral cells can be defined on the Gulf of Mexico and the Great Lakes. Sand is supplied to the coast from coastal streams, erosion of coastal cliffs, erosion of offshore reefs, or transport of offshore sediments onshore. Transport of sand up and down the coast is usually by waves and currents. Sediment sinks include submarine canyons, harbors, lagoons, sand dunes, and deep-water bars and shoals.

Coastal erosion is a major concern for many areas and sea-level rise will compound this problem. While coastlines have receded and advanced over the past centuries, human activities can interrupt or modify the supplies and delivery of this material to and along the coast. Dams and other flood control structures can block sediment and trap it in reservoirs or prevent channel erosion and reduce sediment supply. It has been estimated that hundreds of millions of cubic meters of sand are trapped behind dams in the Los Angeles, CA region and similar amounts are trapped by all the flood control structures on the Mississippi. The worldwide estimates for the amount of material trapped in reservoirs would be in trillions of cubic meters. In addition to trapping sediment, flood control structures reduce the carrying capacity of streams and thus the amount of sediment delivered to the coast.

Finally, in-stream sand and gravel mining removes material that was being transported to the coast and reduces the overall amount of beach sediment that reaches the coast.

Coastal cliffs are a second source of beach material. Many coastal cliffs are composed of sandstone or contain a large percentage of sand-sized sediment. As these cliffs erode, they can supply significant amounts of beach quality material to the coast. When roads or rail-roads are built at the base of a coastal cliff, these structures can interfere with cliff erosion and reduce sediment delivery to the coast through road-clearing efforts. Supplies of sediment from coastal cliffs also can be reduced by the construction of seawalls or other armoring that prevents cliff erosion.

Waves can carry sand from offshore sources to the near-shore and onto the dry beach. These offshore sources, such as deep-water bars or ebb-tidal shoals, are often used for beach nourishment. Human activities of dredging from these offshore and depositing it on beaches has augmented the natural transport of offshore sediment to the coast, or made more sources of offshore sediments available as littoral material. However, if areas within the zone of on-and offshore transport are deepened, these areas will refill with littoral material that will be lost temporarily as beach material. Also, offshore structures can block the onshore transport of sediment.

Orville Magoon: deceased.

Once sand is on the beach and in the nearshore area, waves and currents move it up and down the coast regularly. Structures, such as groins and jetties interrupt longshore transport of sediment and relocate areas of erosion and deposition. If there is no effort to place sediment down-coast of these structures or to by-pass sediment around the structures there will be accretion on the upcoast side and erosion on the down-coast side. The effect may be compounded by the construction of an entire series of structures. Offshore structures like breakwaters and reefs block or reduce wave energy shoreward of the structure. This modification of wave energy will alter the sediment carrying capacity in the nearshore and also the erosion and accretion areas.

Submarine canyons, deep-water bars, and sand dunes are natural sinks for coastal sediment. In general, once sediment is carried into one of these areas, it is no longer readily available as beach material. Harbors, inlets, and deepened navigation channels can also be temporary sinks. These features can function just like natural sediment sinks unless they are regularly maintained and the trapped material is reintroduced to the littoral system.

Finally, sand mining accounts for significant losses of beach material. Large-scale commercial sand mining has taken place on various beaches (e.g., Monterey, CA) and in various streams (e.g., San Juan Creek, Orange County, CA). This sand mostly goes for construction purposes, but it is also used in less familiar ways such as for making glass or pottery. Noncommercial sand mining also occurs. Every summer truck-loads of sand are removed daily when seaweed, flotsam, and trash are raked from the beach. All these activities result in a net deficit of sand moving to and along the coast.

Societal Significance of Coastal Sand

Coastlines have had importance throughout history. Fishermen built their homes and shops by the coast and gained access to the great food supply of the ocean. Oceans provided some of the earliest and most lasting modes of transportation and commerce routes. The coast has always had an important recreational value. The coast has also played an important role as a place of solace and spiritual renewal. It has become clear that everyone benefits from the coast.

Sand serves valuable purposes. It is a commercial resource, as is attested by the extensive sand-mining operations worldwide. Beaches are major recreational resources and tourist attractions for most coastal states and nations. Sandy beaches are also a major buffer and zone of protection for inland areas from coastal storms and hurricanes.

Beach erosion is a major problem in many areas. The coastline has never been static, yet, the problem of erosion has been exacerbated by those human activities that either have reduced supplies of sediment to the coast, or have

changed its transport along the coast. As civilization demanded more from the coast, activities such as construction of cities and harbors interrupted natural supplies of coastal sediments. However, until recently, the connections between development far from the coast, changes to coastal sediment supplies and coastal erosion were not well-recognized.

Doctrine of Public Trust and Sand Rights

The doctrine of public trust traces its lineage to ancient Roman law that established certain types of property to be common to all people and incapable of private ownership. These common properties, *rse communes*, include running water in the sea and the land beneath them. The sovereign or government is under obligation to manage these *rse communes* lands for the public interest, in perpetuity. A private owner can have a bare legal title to the lands, the *jus publicum*, but this interest is subject to and restricted by the superior public interest.

The public trust doctrine was lost during the Middle Ages and was reinstated in England in the middle of the 16th century. Under this interpretation, the restraint was placed only upon the sovereign and only for tidelands, not for navigable waters. However, Parliament retained the power to enlarge or diminish the rights of the public over the tidelands, provided some legitimate public purpose was asserted.

The Roman and English versions of public trust have been carried over into American law where it is now firmly established. Since most land use law has been assigned to state and local governments, the development of public trust varies from state to state. Traditionally, public trust was reserved for navigation, commerce, and fisheries. Many states have expanded the use of public trust to water resources, protection of the environment, and recreational values.

In Florida, and states such as Oregon and Hawaii, the customary rights doctrine has been invoked to protect the public use of beaches. The Florida Supreme Court has observed:

The beaches of Florida are of such a character as to use and potential development as to require separate consideration from other lands with respect to the elements and consequences of title. The sandy portion of the beaches are of no use for farming, grazing, timber production, or residency – the traditional uses of land – but has served as a thoroughfare and haven for fishermen and bathers, as well as a place of recreation for the public. The interest and rights of the public to the full use of the beaches should be protected. (*City of Daytona Beach v. Tona-Rama, Inc.* (1974) 294 So. 2d 73,77)

In California, the California Coastal Commission has used existing statutory authority to require mitigation of quantifiable sand losses. Both the losses from the construction of seawalls along coastal bluffs and the reductions to inland sediment supplies have been addressed through financial compensation into regional nourishment programs. Public agencies can use existing authorities to fund many nourishment projects.

Conclusion

The coast is one of the nations' most valuable areas and sand beaches are critical parts of the coast. The Sand Rights doctrine extends public trust to coastal beaches and those sand resources that are essential to perpetuation of coastal beaches. This doctrine requires that all decision-makers give careful consideration to proposed or existing projects that interfere with the delivery of sand to, or transport of sand along the beach. The basis doctrine is that human actions will not interfere, diminish, modify, or impede sand and other sediments from being transported to and along beaches, shores, or any flowing or windblown paths or bodies without proper restitution. Under this doctrine, projects should be designed or evaluated to mitigate any interference that the project may have with sand transport.

For further related reading see the following bibliography.

Cross-References

- [Beach Erosion](#)
- [Beach Nourishment](#)
- [Coastal Boundaries](#)
- [Dams, Effect on Coasts](#)
- [Erosion: Historical Analysis and Forecasting](#)
- [Erosion Processes](#)
- [Navigation Structures](#)
- [Sediment Budget](#)
- [Shore Protection Structures](#)

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Sandy Coasts

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Definition

Sandy coasts are those coasts dominated by an abundance of sand-size sediments (0.063–2 mm). The location of these coasts is a function of both sand sources and coastal processes. As a consequence, they are more abundant in certain climate and geological or plate settings. They are most prevalent in humid climates supplying abundant terrigenous sand to passive margin coasts and where exposed to more energetic wave and tidal environments that can both winnow out the finer mud and concentrate and deposit the abundant sands in a range of wave and tide deposited features. In addition, sandy coasts are also influenced by secondary regional features such as geology of the hinterland and shore, geological inheritance, sources of shelf siliclastic and carbonate sands, and littoral drift.

Terrigenous Sources

Sand is the most abundant sediment of the world's open coasts. It occurs from the poles to the tropics, but with latitudinal and regional maxima. Sand is globally abundant for two reasons. First, the ultimate source rock of most terrigenous sand is granite, which makes up the cores of all continents. As granite erodes the less resilient weaker, unstable minerals (feldspars) chemically weather to fines, leaving behind the harder sand-size quartz (or silica) grains together with minor percentages of sand-size heavy minerals, all of which are resilient to abrasion. Secondary sources are sedimentary and metasedimentary rocks, all of which contain variable percentages of sand-size material. The erosion and weathering of all these rocks potentially supplies sediment ranging from boulders to mud. However, erosion and transport processes selectively erode and transport the fines most readily in suspension, then the sands as traction and bedload, while the coarser gravel, cobbles, and boulders become increasingly intransigent. Table 1 highlights the transportability of different size fractions.

Transport

Fine sediments are readily transported in suspension by rivers and streams to the coast. At the coast, their continued

Sandy Coasts, Table 1 Sediment size and settling rates. (From Short (1999), © John Wiley & Sons Limited, reprinted with permission)

Size	Grain diameter	Time to settle 1 m	Distance traveled per hour in 1 m/s current
Clay	0.001–0.008 mm	Hours to days	3.6 km
Silt	0.008–0.063 mm	5 min–2 h	3.6 km
Sand	0.063–2.000 mm	5 s–5 min	10's m
Cobble	2 mm–6.4 cm	1–5 s	<<1 m
Boulder	>6.4 cm	<1 s	0

suspension and transport by waves and tidal currents result in their deposition being restricted to quieter estuarine, deltaic, or shelf locations. Sand on the other hand once it reaches the coast is deposited rapidly as bedload, building river mouth bars and deltas. It can then be reworked on- and longshore by waves into bars and beaches, and by tides into tidal sand waves, ridges and deltas, all shallow water and shore features. The size of gravel and boulders restricts their erosion, transport, and availability. They will only reach the coast at the mouth of short, steep rivers, along eroding rocky coasts and when delivered by glaciers.

Accumulation and Recycling

For the above reasons, even though fines dominate many of the world's river sediment discharges, their deposition is precluded from energetic shores. Sands, however, are not only deposited right at the coast but once deposited are ideally suited to reside there. Gravel and boulders will also remain at the shore once deposited; however, they are more restricted in extent to usually high relief and high latitude coasts. Quartz sand and particularly heavy minerals are extremely resistant to physical abrasions and can be reworked time and time again by rivers, waves, tides, and winds and are termed polycyclic. Consequently, many sedimentary shores are composed of quartz sand, because it is globally abundant in source rocks (20–40% of crust); it is the major bedload product of denudation; it is relatively easily transported during high river discharge events; it can reside in the most energetic coastal environments; it is resilient and over time and can accumulate along coasts and can be reworked during rising sea levels, leading in places to massive polycyclic accumulations of Quaternary (shelf) sand deposits.

Carbonate Sands

The second major source of coastal sand is marine carbonates that live from shallow intertidal waters to the inner shelf. While carbonate detritus is more easily broken and abraded by physical processes, their in situ and shore linear sources

can act as a continual supply, with waves and tidal currents eroding, abrading, and transporting the sand size and coarser material shoreward. Carbonate sand coasts in fact dominate large areas of the world's tropical and arid temperate coasts (Short 2013) (Fig. 1). The source of these sands are both the shallow coralline algae and reefs in the tropics, while in more temperate latitudes, shelf carbonates (molluscs, red algae, encrusting bryozoans, and echinoids) are delivered both during and subsequent to sea-level transgressions from depth as great as 100 m (James and Bone 2010).

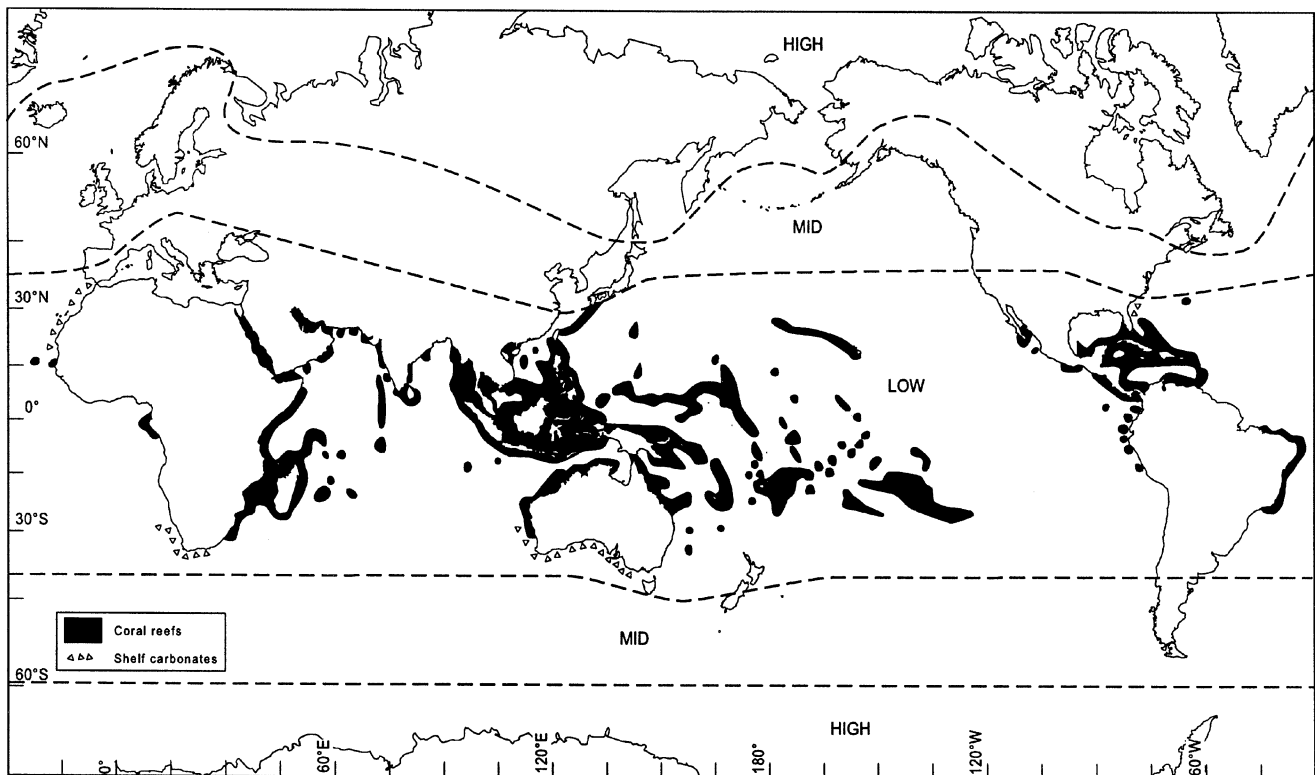
Climate

Climate plays a major role in the occurrence of sand coasts. Most terrigenous sand is derived from denudation of the hinterland and transported by rivers to the coast. Regions of greatest chemical and physical weathering will potentially supply the greatest volumes of sediment including sand to the coast. Figure 2 shows the global distribution of mechanical erosion, the world's major rivers and their solid discharge. Using this as an indicator of bedload sand discharge, the greatest supply of sediment to the coast is associated with low to lower mid-latitude river systems (40°N–40°S). All the rivers and their headwaters are associated with humid tropical and mid-latitude climates that supply the higher rainfall and warmer temperatures to physically and chemically erode and transport the sediments. While there are some major rivers in the higher latitudes (>40°N), the smaller discharges and dominance of physical weathering supplies generally low quantities of predominantly coarser sediments (sand through boulders). The impact of both climate and latitude on shelf sediment supply is highlighted by Fig. 3, which shows the latitudinal distribution of shell, coral, rock and gravel, sand and mud. Sand dominates the subtropics and lower mid-latitudes (20–40°), and while sand volumes are still large in the tropics, mud still dominates owing to the intense chemical weathering and lower-energy coasts.

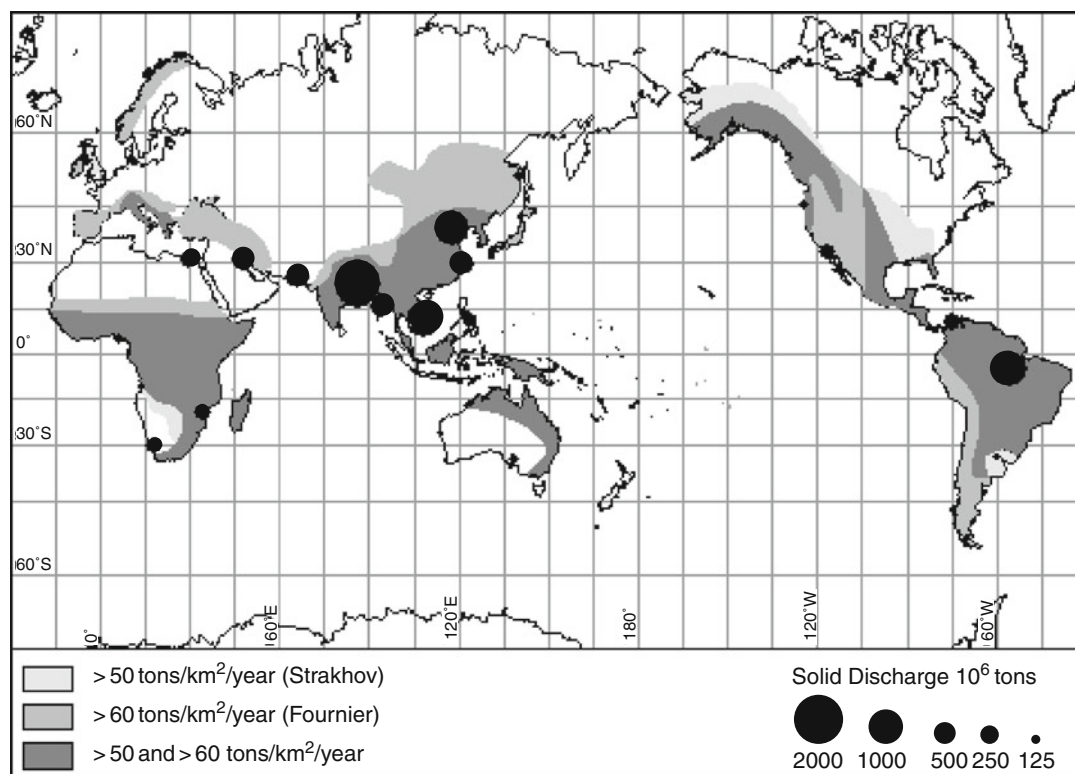
Hinterland geology also plays an important secondary role, as sand-rich source rocks (granites, sandstone, meta-sedimentary) will supply more sand and less suspended sediment than finer grained rocks (e.g., basalt, shale, limestone).

Plate Setting

Location of a coast relative to its tectonic plate setting is also a major contributing factor in the location of sand coasts. Inman and Nordstrom (1971) classified the world's coast according to their tectonic setting and noted the dominance of coastal plains and deltaic coasts, usually associated with sand coast, on passive margin coasts (Table 2). The reasons for this are highlighted in Table 3. The passive margin coasts are supplied



Sandy Coasts, Fig. 1 Extent of the low-, mid-, and high-latitude coasts, extensive coastal reefs and known areas of major supply of shelf carbonate sands to the coast. (Modified from Davies (1980) and reprinted by permission of Pearson Education Limited)



Sandy Coasts, Fig. 2 Global distribution of mechanical erosion. Circles are proportionate to the amount of solid discharge per annum by the world's major rivers. (Modified from Davies (1980) and reprinted by permission of Pearson Education Limited)

with abundant sand and finer sediments by larger river systems, draining distant mountains. At the coast, they supply abundant, mature, stable sand, and fines, which build extensive sand-rich coastal strand plains composed of deltas and barriers, as along the east coast of the Americas and India. In contrast, convergent coasts have high relief and short rivers which tend to supply, along with mass movement, a limited amount of coarse, poorly sorted unstable sediments, as along the west coast of the Americas.

Landforms

Deltas

Sand supplied by river to the coast is initially deposited in deltaic systems. The proportion of sand in a delta and the overall morphology of the delta is a function of both the composition and volume of the fluvial sediment supply, the geological inheritance that influences the shape and size of

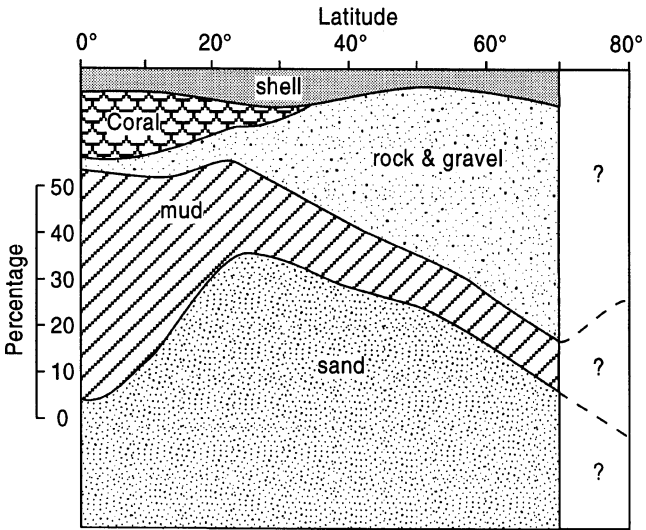
depositional basin or accommodation space, and the contribution of the river flow, waves, tides, littoral currents, and winds to the reworking, redistribution, and construction of the deltaic and adjacent depositional systems (see e.g., Davis 1983). River-dominated deltas tend to deposit the sand as digitate river channel, levee, and river mouth bar deposits, as in the Mississippi, with limited shoreline sand deposits. Tide-dominated deltas accumulate the sand in subaqueous shore perpendicular to linear tidal sand waves and ridges, again with little sand at the shore. Only along wave-dominated deltas do extensive shore parallel beaches, barriers, and strand plains occur, which if exposed to strong winds may be capped by coastal sand dunes.

Beaches

Beach systems are the most common and visual expression of sand coasts. They form the shore of all sand coasts whether backed by deltas, barriers, and sand dunes or even fronted by sand flats. Beaches can, however, be formed from fine sand through boulders, and like sand supply itself, the sediment type is a function of tectonic setting (Table 3) and latitude/climate (Table 4). As Table 4 indicates sand beaches are more likely to occur in the low- to mid-latitudes, with beach type also a function of sediment size, and through climate, wave environment. Some of the world’s longest beach systems occur along the east and Gulf coast of the United States and the east coast of South America, with beaches up to several hundred kilometers in length in south Brazil. While the visible beach usually forms a narrow strip along the shore, most beaches also extend offshore to the limit of modal wave base. On high energy coasts, this may be as deep as 30 m and as far as 2–3 km offshore.

Dunes

All beaches composed of fine to medium sand and exposed to onshore winds, particularly strong winds, will be backed by coastal sand dunes, composed of fine to medium well-sorted sand. The location and occurrence of dunes is also a function of latitude/climate (Table 4) with the world’s largest coastal sand dunes occurring on passive margin coasts exposed to



Sandy Coasts, Fig. 3 Relative frequency of occurrence of inner continental shelf sediment types by latitude. (From Hayes (1967) and with permission from Elsevier Science)

Sandy Coasts, Table 2 Tectonic coastal types and shoreline types. (From Short (1999), ©John Wiley & Sons Limited, reprinted with permission; modified from Inman and Nordstrom (1971))

Tectonic setting and shoreline type	Mountainous	Hilly, narrow shelf	Plains, narrow shelf	Plains, wide shelf	Hilly, wide shelf	Deltaic	Reef	Glaciated	World's coast
Convergent	96.2							23.9	39.1
Passive-neo	1.0	69.5	15.5			3.8			4.3
Passive-Afro		21.3	73.3	1.0		20.7	16.7	47.8	6.8
Passive-Amero			11.2	98.0	11.3	37.2	49.8	28.2	35.4
Marginal	2.8	19.2		1.0	88.8	38.3	33.6		8.8
Total (%)	100	100	100	100	100	100	100		
World's coast (%)	39.4	5.0	4.7	31.9	7.1	1.2	1.3	4.6	100

moderate to strong onshore winds, as along the Brazilian trade wind coast and the south to west-facing coasts of South Africa, southern Australia, South America, west coast United States, and parts of western Europe. Most coastal sand dunes extend up to a few tens to hundreds of meters inland

reaching elevations of a few tens of meters. They can in very exposed locations migrate up to tens of kilometers inland and to elevations of over 100 m and up to 200–300 m in extreme locations. In these locations, the coastal sand dunes become a very dominant and prominent feature of the coastal landscape.

Sandy Coasts, Table 3 Coastal characteristics of convergent and passive margin coasts

	Convergent margin	Passive margin
Age	Young (1–10s of millions of years)	Old (100s of million of years)
Relief	Steep, mountainous	Low-gradient plains
Landforms	High mountains and volcanoes	Coastal aggradation plains
	Narrow continental shelf	Wide, low continental shelf
	Deep-sea trough	Continental slope
Tectonics	Active, earthquakes	Quiescent, stable
Weathering	Physical, mass movement	Chemical, fluvial, rivers
Drainage	Short steep streams	Long, meandering rivers
Sediments		
Quantity	Low	High
Size	Fine-coarse (mud-boulder)	Fine (mud-sand)
Sorting	Poor	Well
Color	Dark	Light
Composition	Unstable minerals	Stable minerals (quartz)
Coastal landforms	Rocky, few beaches	Extensive barriers and deltas
Examples	West coast Americas	East coast Americas
	New Zealand	Southern Africa
	Iceland	Southern Australia
	Japan	North Alaska
		India

Tidal Sand Deposits

Coastal sand deposits can also include tide-deposited sand, which are by definition will be located in the inter- to subtidal. Ebb and flood deltas composed of marine sand are a component of all barrier islands and all barrier inlets. Intertidal sand can also accumulate in overwash flats, in the tidal deltas and tidal flats. Along wave-dominated barrier systems, tidal sand deposits occur at occasional inlets. They become more dominant with decreasing wave height and increasing tide range as inlets increase in number and size, while along tide-dominated shores, intertidal sand flats and subtidal sand ridges dominate the coast and nearshore zones.

Barriers

All of the above, beaches, dunes, and tidal inlets are usually components of a larger sand barrier system. The barriers represent longer-term accumulations of wave-, wind-, and tide-deposited sand and other sediments. The barrier may be narrow transgressive barrier islands, wide regressive strand plains capped by beach and/or foredune ridges, or backed by larger coastal dune systems. Some of the world's most extensive sand barrier coasts include the barrier islands along the east and Gulf coast of the United States, the prograding attached barriers or embayed barriers as in much of southern Australia, and large barriers capped by dune systems as in parts of Brazil, South Africa, and Australia. Barriers can also contain tidal sand deposits and where not capped by dunes, they are commonly backed by washover fans and aprons.

Sandy Coasts, Table 4 Sediment and beach-dune characteristics on high-, mid-, and low-latitude coast. (Modified from Short (1999), © John Wiley & Sons Limited, reprinted with permission)

Latitude	High	Mid	Low
Climate type	Polar	Temperate	Tropical
Latitudinal range	50°/60°–90°	30°/40°–50°–60°	0°–30°/40°
Sediments	Coarse (cobbles, shingle) dark, unstable minerals	Terrigenous and shelf quartz sand, temperate shelf carbonates	Fine terrigenous quartz sand, coral and algae reefs
Beach type	Reflective beach face possible multibar	Reflective to dissipative	Reflective
Other climatic impacts	Permafrost in barriers	Dune calcarenite in arid regions lithifies beaches and dunes	Beachrock lithifies intertidal beach
Dunes	Low and poorly developed owing to: coarse sand, light winds, short season, frozen winter surface, poor vegetation cover, prone to overwash	Largest on west-facing coasts in mid-latitudes. Well-developed: full range of dune types (foredunes through massive transgressive dunes)	Moderate to minimal owing to low waves and sediment supply and low winds: predominantly foredunes, some dune transgression on exposed trade wind coasts

Large-Scale Sand Coasts and Landforms

As indicated in Tables 2 and 3, the largest accumulations of sand at the coast are associated with humid climates on passive margin coasts, in the lower-mid to low-latitudes. The most extensive sand coasts on each continent are:

North America: The southeast and Gulf coasts, both low-gradient passive margin coasts supplied by numerous rivers including the Mississippi, with the sand reworked onshore by a high-energy wave environment. Coastal dunes are, however, poorly developed. Also parts of the west coast exposed to high shelf sediment supply and winds to build dunes.

South America: Massive long-term sand supply to entire east coast, leading to an essentially sand barrier-dune coast from the Amazon south to Argentina. Long beach-barrier-dune systems for much of the coast.

Africa: Beach, barrier, and dune systems ring most of the continent, with sand supplies by local rivers, including the Nile, Niger, and Orange, and on the most exposed coasts by shelf supply of quartz, and in the south also carbonate sands.

Asia: Sand dominates most exposed western (Europe-Mediterranean), southern (India), and eastern shores (Southeast Asia-China), with some substantial river systems and deltas and extensive sand barriers in south India and Sri Lanka.

Australia: 50% of passive margin coast consists of sand deposits, with low-energy beaches in north through high-energy beaches and dune systems across south. Supply from rivers and shelf in north, and from shelf quartz in southeast and carbonate in the south and west.

The largest single accumulation of coastal sand in the world is *Fraser Island* on Australia's east coast. The massive sand island is 125 km long and up to 25 km wide. It averages 100 m in elevation reaching 244 m and has an area of 1840 km² and a conservative volume (above sea level) of 185 km³. It is the largest and last of five near continuous sand islands which extend for 320 km along on a passive margin coast backed by a humid hinterland. The long-term (Pleistocene-Pleistocene) accumulation of sand on the islands has been favored by numerous rivers supplying quartz sand to the updrift coast, predominately northern longshore transport toward the islands, coastal orientation to receive the southerly waves, and coastal inflection north of the island which places the island at the terminus of the longshore transport system, bedrock headlands to tie the island, moderate to occasionally high-wave energy to move sand long and onshore, and moderate south-east trades to the build successive layers of massive dunes, thereby removing the sand from the longshore transport system and storing it on the island.

Summary

Sand is the most ubiquitous sediment type found on the world's coastlines, usually in the form of beaches at the shoreline, as well as extending seaward as nearshore deposits and landward as dunes, overwash, and tidal deltas. Sand size particles originate from two major sources at the coast. Terrigenous quartz grains are eroded from the hinterland bedrock (granite, sandstones, etc.), transported to and reworked and deposited at the shore. Carbonate sand is produced by a range of organisms living from the shoreline out onto the inner shelf. They are eroded, broken up, and ultimately deposited at the shore as sand size particles. Fine to coarse sand size particles dominate the energetic shoreline as their relatively fast settling velocity ($\sim \text{cm sec}^{-1}$) enables them to be both transported in suspended and bedload in large quantities to the coast and be able to remain in this energetic environment, that winnows out finer material.

Cross-References

- [Barrier](#)
- [Barrier Island Landforms](#)
- [Beach Sediment Characteristics](#)
- [Carbonate Beaches](#)
- [Deltas](#)
- [Dune Ridges](#)
- [Muddy Coasts](#)
- [Tidal Environments](#)
- [Tidal Inlets](#)
- [Wave-Dominated Coasts](#)

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Scour and Burial of Objects in Shallow Water

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Scour around objects on or near a sediment bed are caused by flow modification due to the object. The presence of the object generates vortices that locally change the bottom stress inducing changes in the sediment transport rate. The presence of an obstacle on a sediment bed where the stress is below the threshold (onset) of motion may induce local intensification and sediment transport, while an obstacle on a bed surface where the stress exceeds the threshold will alter the stress field, producing local depressions and ridges in the bed. The scour phenomenon occurs in unidirectional and oscillatory flow and in types of fluid ranging from air to water to sediment-laden turbidity currents and pyroclastic flows. Obstacles producing scour range from millimeter size grains to topographic features many meters high and kilometers in length. The resulting bedforms range from sand streaks and ripples to large desert dunes and scour moats and sediment drifts around seamounts in the deep ocean.

Here we are primarily concerned with scour in nearshore waters and on beaches where the flow is both unidirectional and oscillatory. Scour naturally occurs wherever a larger object occurs on an otherwise smaller grained bed. For example, a seashell or a kelp-rafted rock on the seabed will form scour features ranging in size from rhomboid marks around small objects to crater-like crescentic depressions twice the size of the shell or rock. On larger scales, rip currents scour large channels over and through longshore bars. Above the waterline, wind-blown scour features form around kelp clumps and rocks on the beach, while accretionary dunes and sand shadows form around hardrock outcrops on the coastal desert floor. In recent decades, the importance of scour in the burial of mines has led to increased interest in scour phenomena.

Introduction

Scour and scour marks are the erosional and accretionary bedform patterns that occur in the vicinity of obstacles that are on or near a sediment bed. Scour always involves some degree of perturbation in the flow system that changes the

local pattern of erosion and deposition relative to that of the general flow. The primary scour pattern may be erosion as in the formation of crescentic scour around a rock on the beach (Fig. 1) or depositional as in the dune deposits in the stagnation area of flow around an outcrop (Fig. 2).

Any form that locally concentrates vorticity near the bed can elevate bed shear stress and initiate onset of grain motion, leading to local bed scour including bumps and depressions on the bed itself (Fig. 3). Once initiated, a pattern of scour may spread down current in the form of a growing field of current ripples (Fig. 3b–d), while vortex ripples under wave action may spread both against and with wave propagation from a single initiating irregularity in the bed (Inman and Bowen 1962; Tunstall and Inman 1975).

The most commonly studied scour patterns are those associated with single bluff (blunt) bodies placed on or protruding from the bed. There is an extensive engineering literature of the scour around the piles of bridges and piers (e.g., Collins 1980; Chiew and Melville 1987). Engineers refer to scour as *local* when it results from the effects of a structure on the flow pattern and *general* when it would occur irrespective of the presence of a structure; it is termed *clear water scour* when the bed upstream of a structure is at rest (e.g., Raudkivi 1990).

In sedimentology, the interest has usually been in the scour pattern around individual objects, referred to as *scour marks* (e.g., Pettijohn and Potter 1964; Reineck and Singh 1975) or as *obstacle marks* (Allen 1984, 1985). Allen further subdivides obstacle marks into *current crescents*, *current shadows*, and *scour-remnant ridges*. It appears, from the extensive literature on the subject, that current crescent and *crescentic scour mark* are the most general terms for the crescentic feature formed around an object on the bed, and that the feature may be either erosional as in Fig. 1 or accretional as in Fig. 2. The appearance of other associated features such as current shadow, scour-remnant ridges, and ripples are wake phenomena that depend upon the height to width aspects of the object, the nature of the flow system, and the type of sediment.

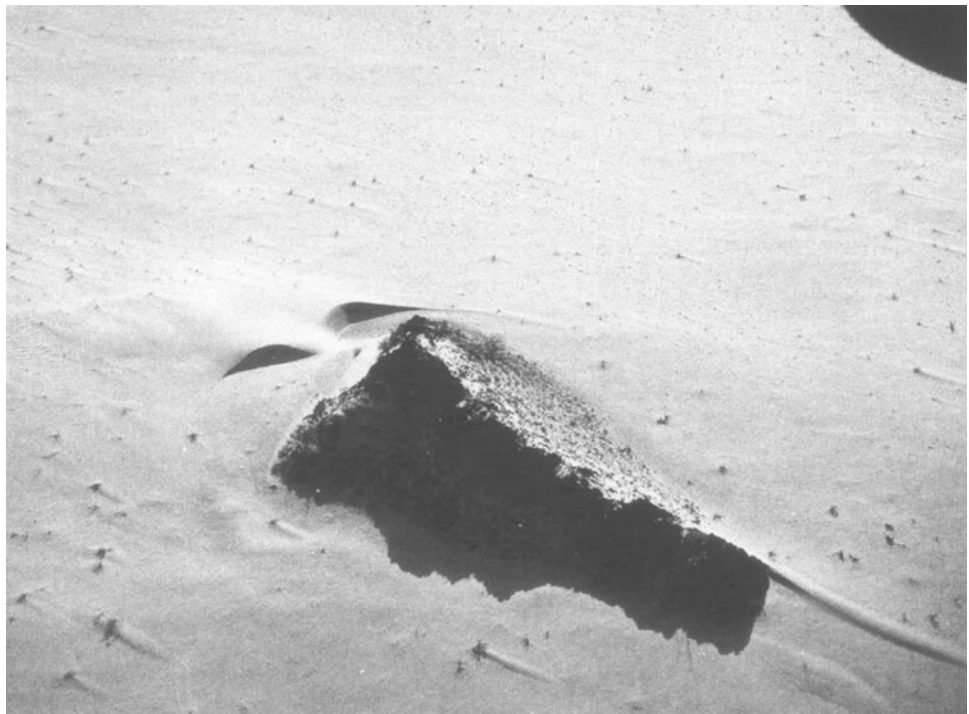
Bagnold (1941) introduced *sand shadow* for the various shapes of sand accumulation in the shelter of an obstacle. Allen's (1984) current shadow is based on Bagnold's sand shadow. Allen defines scour-remnant ridges as the small ridges of sand, snow, or mud preserved in place on the lee side of resistant objects after the surrounding material has been eroded away. Thus scour-remnant ridges are residual leeward phenomena while current (or sand) shadows are depositional and/or erosional leeward bedforms.

Characteristics of the flow around a vertical cylinder, such as a bridge pile in steady currents, have been investigated extensively (e.g., Shen et al. 1969; Breusers et al. 1977). It has been found that a horseshoe vortex above the scoured bed is a dominant factor in the scour process, and that the vortex has a

Scour and Burial of Objects in Shallow Water, Fig. 1 Wind-formed crescentic scour pattern around a rock (~30 cm diameter) on the beach berm, Coronado, CA. Wind blows from left to right. (D.L. Inman photograph)

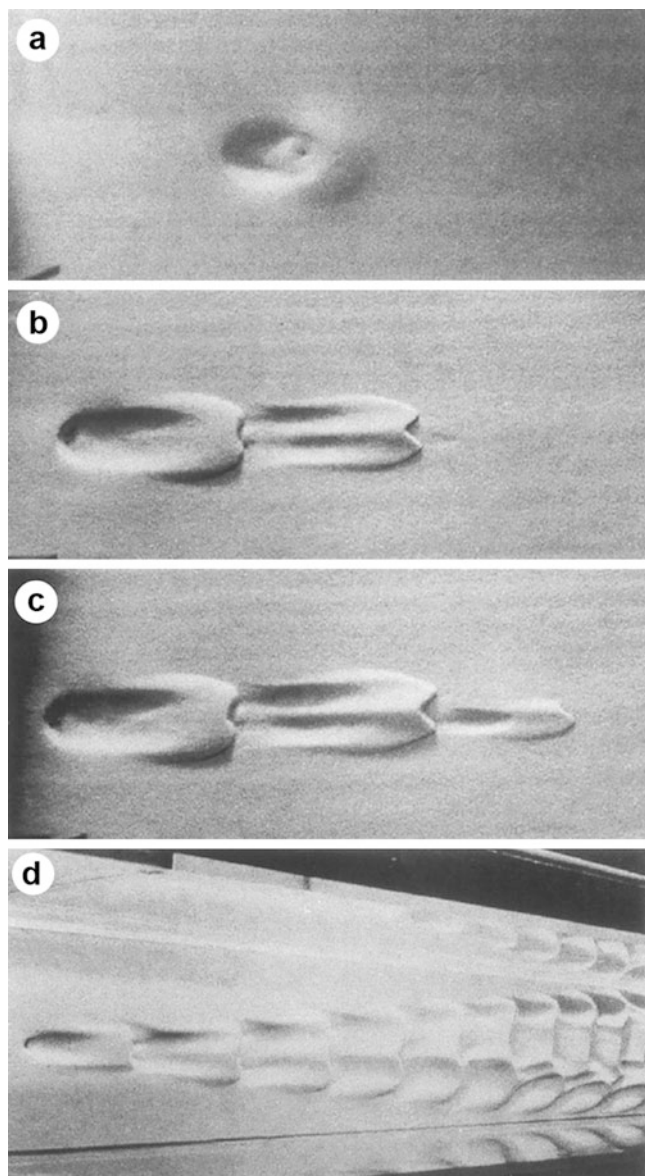


Scour and Burial of Objects in Shallow Water, Fig. 2 Wind-deposited crescentic pattern consisting of two barchan dunes formed around an outcropping ridge (~8 m high, 40 m long) on the desert floor, northwestern Gulf of California. Wind blows from upper left to lower right as indicated by sand shadows in the lee of bushes and by the double shadow from the sand drift passing along the sides of the ridge. (D.L. Inman photograph)



close relationship with the bed profile near the cylinder. The vortex behavior caused by the object is thus a very important factor to consider in the estimation of bed scour as described under Scour Mechanics.

Relatively few studies have been conducted on scour induced by waves and currents. Nishizawa and Sawamoto (1988) and Sumer et al. (1992) have studied the flow around a slender vertical cylinder under waves using flow visualization



Scour and Burial of Objects in Shallow Water, Fig. 3 Progressive downstream propagation and lateral widening of a single perturbation (a) on a fine sand bed. Flume is 17 cm wide (sidewalls visible in panel as in (d) and bottom stress is near the threshold of grain motion. (Photographs by permission of John Dingler)

techniques. They have reported relationships between the flow characteristics and nondimensional parameters of flow similitude such as the Strouhal number (acceleration forces/inertial forces) based on experimental results for a flat bottom. There is less literature for a scoured seabed, probably because these flows often show a highly complex three-dimensional rippled pattern owing to the complicated shape of natural obstacles and the unsteadiness of the main flow. Even for simple shapes such as a cylinder, unsteady three-dimensional flow over the bed generates vortices that govern the local scour process.

Continued scour around objects on a sand bed usually leads to complete burial of the object. Shells and rafted objects dropped to the seafloor eventually bury and disappear if the sand bed has sufficient thickness (e.g., Inman 1957, Figures 20, 21). A study of mine scour and burial was conducted for the Office of Naval Research during 1953–56 on the sandy shelf off the Scripps Institution of Oceanography in La Jolla, CA, in water depths extending from the surf zone to 23 m (Inman and Jenkins 1996). The mines were 1.6 m long cylinders, 60 cm in diameter. It was found that the burial process began with scour depressions around the mine and continued with rollover into the depressions (Fig. 4). This process repeated itself until the mine was completely buried. Burial time ranged from hours and days in the surf zone to almost one year at 17 m depth.

Scour Mechanics

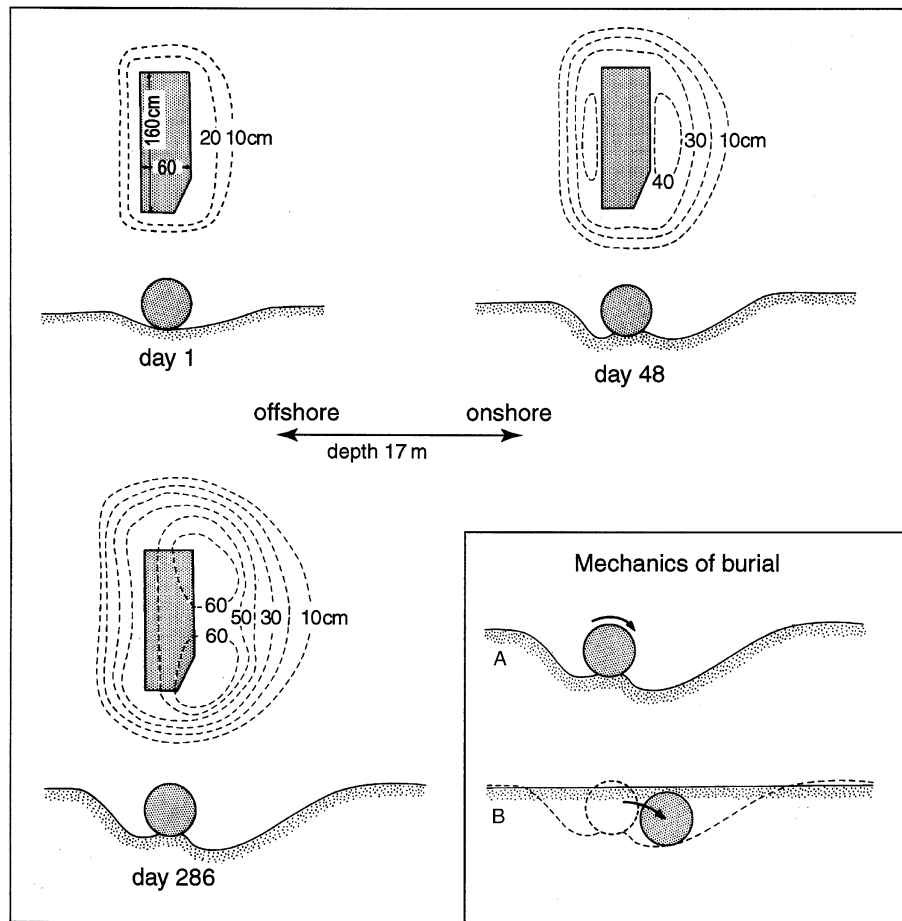
The scour phenomenon around objects differs from other types of sediment flux in that the presence of an object on or near the bed induces local changes in an otherwise uniform pattern of bed stress, thereby causing local patterns of erosion and/or accretion that may differ from the general bedform pattern. The object may be either blunt or streamlined and the resulting scour pattern may be erosional, depositional, or both. Scour develops from a variety of mechanisms whose relative importance depends upon the scale and intensity of the flow and the relative size and shape of the obstacle. The most common and largest bedforms result from scour around bluff bodies where the formation of a horseshoe vortex generates a scour hole that begins on the upstream side and wraps around the object (Fig. 1) as described below.

The mechanics of the scour around a body are inherent in the vorticity field generated when a fluid moves over a bed or solid surface. For example, consider the velocity profile above the bed and up current from an object on the bed (Fig. 5). The shear near the bed in the bottom boundary layer generates vorticity between the layers of differing flow velocity creating a vorticity sheet. *Vorticity* is the angular momentum of a fluid element, while a *vortex* is the arrangement of many of these fluid elements into a pattern of angular motion.

The presence of an obstacle on the bed rearranges the sheet of vorticity in the boundary layer, and creates new vorticity by the shear flow around the obstacle. The flow disturbance of the obstacle creates a stagnation point (s) at the bed interface upstream of the obstacle. The bed vorticity in the approaching flow collects at the stagnation point forming a local excess of vorticity that organizes into a *forward bound vortex* (Fig. 5) that moves the stagnation point (s) upstream of the vortex. The forward bound vortex initiates the scour process by causing intense velocity shear stress at the base of the

Scour and Burial of Objects in Shallow Water, Fig. 4

Scour and burial of cylindrical mine by wave action over a fine sand bottom off Scripps Beach, La Jolla, CA. (After Inman and Jenkins 1996; SIO Reference Series No. 96-13). Note vertical exaggeration in profile view



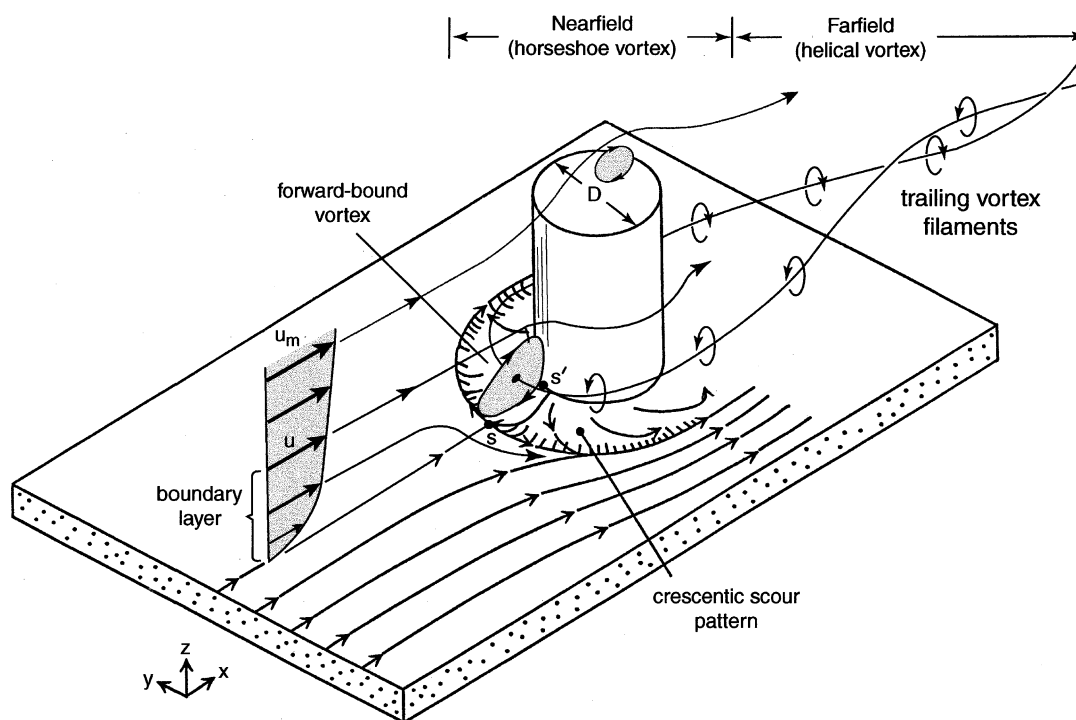
obstacle. The incoming vorticity from the flow builds up in the bound vortex and the excess leaks around the base of the cylinder forming a pair of *trailing vortex filaments* on either side of the obstacle. The bound vortex with its pair of trailing filaments form a vortex system known as a *horseshoe vortex*. The trailing vortex filaments extend the region of scour from the upstream base of the obstacle, around the sides, and downstream. As the trailing filaments extend downstream, the vorticity of the filaments diffuse into the interior of the fluid, thereby slowing the filament rotation and weakening the shear stress on the bed. Consequently, the scour diminishes downstream of the obstacle forming a scour pattern around the obstacle known as *current crescent* or *crescentic scour mark*.

The horseshoe vortex and its associated crescent scour are nearfield bedform responses that occur over distances of about two obstacle diameters. Further downstream in the farfield the trailing filaments of the horseshoe vortex begin to entwine into a helical vortex system. At each crossover of the helical pairs, the induced velocities of the vortex system approach a null on the bed, allowing for complimentary depositional features such as ripple marks in the current shadow downstream of the crescentic scour. Initially, both

the near and farfield scour and its shadow system are referred to as *forced* forms because these erosional and depositional responses are controlled by the length scales of the flow field around the obstacle. However, once the scour process erodes deeply into the bed, the crescentic scour features will modify the flow field around the obstacle. In turn, the modified flow field further modifies the form of the scour and the associated bedforms in the current shadow. The fully developed horseshoe vortex is a consequence of the scour depression. Therefore, once this feedback takes place, the scour depression becomes an *interactive* part of a fluid-bedform system where the bedform interacts with and extensively modifies the flow field above it.

Other Scour Mechanisms

Other scour mechanisms become important in very shallow water typical of the swash and backwash motion of wave runup on the beach face. These mechanisms are associated with thin flows where water velocity often exceeds the critical limit for wave propagation and where capillary waves become important. Also, in thin flows, common "V"-shaped ship



Scour and Burial of Objects in Shallow Water, Fig. 5 Definition sketch of fluid motion and scour features around an upright cylinder extending through the surface of a sediment bed under unidirectional flow. (After Schlichting 1979; Allen 1984); compare with Fig. 1

waves are formed by small objects and induce stress perturbations on the sediment bed. As a consequence, the beach face often shows rhomboid marks caused by one or more of the mechanisms associated with thin flow (Fig. 6). Large rhomboid marks 3–10 m across are known to occur on beaches following the backwash from tsunami waves (i.e., Shepard 1963, Figure 100(B)).

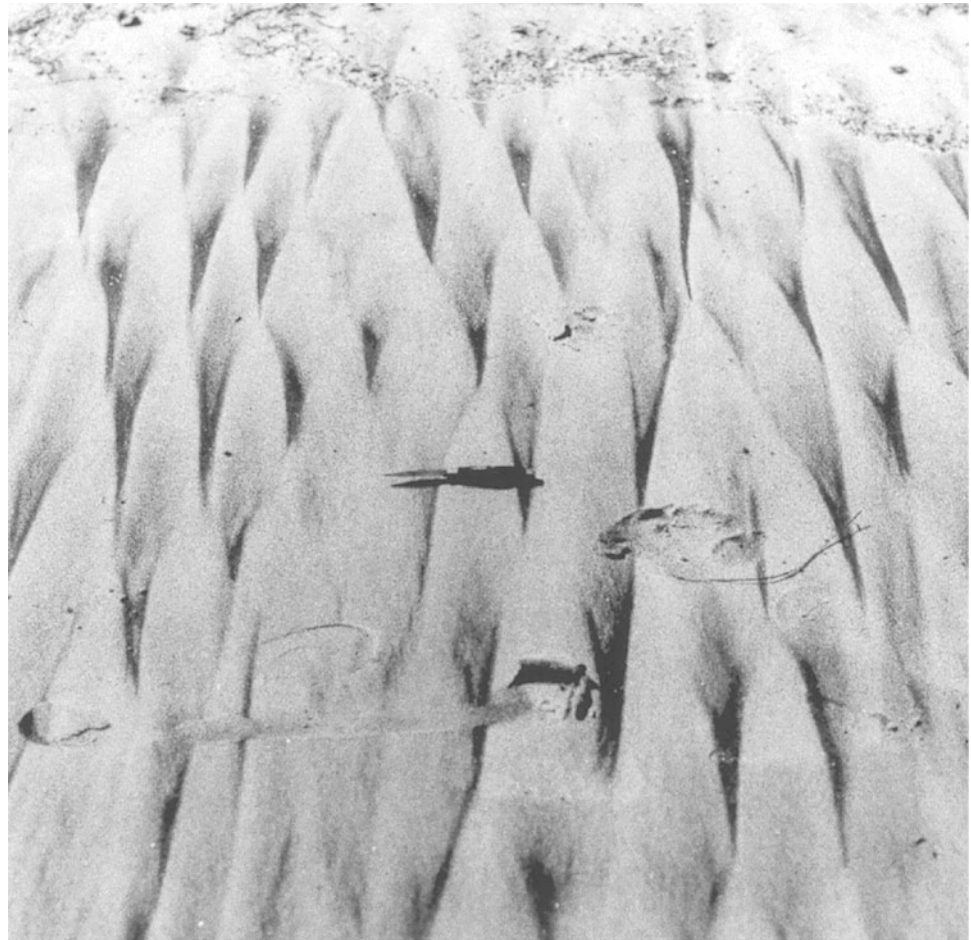
The flow regime over the beach face may be either subcritical ($u < \sqrt{gh}$) or supercritical ($u > \sqrt{gh}$) depending on the speed of the water u relative to the shallow water wave speed $\sqrt{gh}$, where g is the acceleration of gravity and h is the thickness of the flow. In either case, the height of small obstacles such as shells, pebbles, and the feathery antennae of filter feeding organisms that protrude above the bed are of the order of the flow thickness. Supercritical flow is readily perturbed by an obstacle on the bed and locally slowed to subcritical flow by small oblique hydraulic jumps (Henderson 1966) upstream of and extending downstream from the obstacle in a V-shaped pattern. The turbulence of the hydraulic jump scours a corresponding V-shaped erosion pattern around the obstacle, often made strikingly visible by exposure of dark minerals in the laminated beach sand similar to that shown in Fig. 6. Intersections of adjacent V-shaped jumps form the characteristic diamond pattern of the rhomboid ripple. These marks are distinguished by their long scour trail and because the vertex of the V-shape is always upstream of the obstacle, much like the crescentic scour of larger obstacles under

subcritical flow conditions. However, the large supercritical rhomboid marks are less common than would be expected because supercritical flow over sand beaches rapidly develop *backwash ripples*, small-scale sand waves that parallel the beach contours and obliterate the large, extensive rhomboid marks that are found on the otherwise flat beach face.

There are at least three thin-flow mechanisms that may lead to the formation of rhomboidal patterns in subcritical backwash flow. These include the bow and stern wave mechanisms of ordinary ship waves (e.g., Stoker 1957; Whitham 1974), and the formation of capillary waves. Bow waves may form at the stagnation point upstream of an obstacle on the beach, and travel outward with the traditional Kelvin angle of the classical ship wave pattern. In contrast, thinning flow over the obstacle may become supercritical, then revert to subcritical flow with the accompanying small hydraulic jump downstream of the object. The wakes from both bow and stern waves form V-shapes that trail downstream of the object, with the vertex of the V on the upstream side of the obstacle for the bow-wave case and just downstream for the stern wave. The rhomboid ripple marks shown in Fig. 6 were observed to have formed from the bow waves of the filter feeding beach crab *Emerita*. Sand crabs together with other beach-dwelling organisms, such as the bean clam *Donax gouldii* (Ricketts et al. 1985), play a far more active part in shaping bedforms on sandy beaches than is generally recognized.

Scour and Burial of Objects in Shallow Water, Fig. 6

Rhomboid ripple marks on beach face at La Jolla, CA. Diamond pattern associated with flow divergence around antennae of a field of sand crabs (*Emerita analoga*). Photo looking seaward, knife (including blade) 12 cm, swash mark at top. (D.L. Inman photograph)



Often, the system of ship waves formed by small obstacles on the beach have short-length scales where surface tension forces become important. In this case, the gravity–capillary waves (Whitham 1974) can propagate upstream as well as downstream of the object and produce a more parabolic-shaped wave and scour pattern near the object than for typical rhomboid marks.

Cross-References

- [Accretion and Erosion Waves on Beaches](#)
- [Beach Features](#)
- [Coastal Warfare](#)
- [Erosion Processes](#)
- [Ripple Marks](#)

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Seabed Roughness of Coastal Waters

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Definition

Coastal waters are subject to the continuous pounding of ocean waves – by short wind waves and swells and by long-period tidal actions. These regular processes are exacerbated during the episodic forcing of storm surge and tsunami. The resulting hydrodynamic processes caused by these forces face resistance along its boundary at the seabed giving rise to a layer of reduced flow velocity near the bed. The thickness of this boundary layer depending on the shearing actions of currents and waves is described as a logarithmic or power profile in currents and as a function of bottom orbital motion for waves. The resistance to the shearing fluid motion of currents and waves at the boundary caused by the roughness of seabed is defined in terms of granular characteristics and the irregularity of bed, as well as by their hydrodynamic response. The bed irregularities varying from small ripples to large dunes are distinctly different in current and wave regimes. In currents, they are asymmetric with a gentle upstream stoss face and a steep downstream avalanche face. In the wave regime, the bedforms are primarily symmetric.

From hydraulic perspective, the roughness of fixed bed and mobile alluvial bed is not expected to be identical – the former is likely to provide higher resistance than the latter, yet the two are treated the same in literature. Seabed roughness in oscillatory current regimes is primarily relevant in tidal channels and estuaries. In open coastal water including the surf zone, the roughness resisting the wave action is important. In

coastal water bodies acted upon by both the oscillatory current and wave action including the regimes during the periods of episodic storm surge and tsunami, the combined wave-current seabed roughness affects the near-bed hydraulics.

Hydraulic Fundamentals

Because of the fact that the resistance is relevant only when water is in motion, the bed resistance forces are reactive and can best be understood in hydrodynamic terms. In order to facilitate easy understanding of the complicated differential equations, these terms can simply be presented as in Eq. 1 – as a schematic representation of the Navier-Stokes equation (French engineer Claude-Louis Navier, 1785–1836; British mathematician George Gabriel Stokes, 1819–1903). The Navier-Stokes equation is the mathematical elaboration of the Newton's (Isaac Newton, 1642–1727) second law of motion. It can be thought of as a simple force-response system, with responses or accelerations on the left-hand side and the active and reactive forces on the right-hand side of the equation. Considering all different terms in the free surface flows of a homogeneous unit volume of water, the force-response system can be described as

$$A_L + A_C = \frac{1}{\rho} [F_{AC} + F_{AP} + F_{AG} + F_{AN} + F_{AW} + F_{RB} + F_{RE}], \quad (1)$$

where A_L = local acceleration representing unsteadiness, A_C = convective acceleration representing nonuniformity, ρ = water density, F_{AC} = active Coriolis (French mathematician Gustave de Coriolis, 1792–1843) force representing the effects of earth rotation on large water bodies, F_{AP} = active pressure gradient force, F_{AG} = active gravity force representing effects of change in bed-slope, F_{AN} = active wind forcing on the water surface, F_{AW} = active wave-induced force (radiation stress), F_{RB} = reactive bed resistance force, and F_{RE} = reactive eddy viscous force representing viscous and turbulent momentum transfer within the water column. The superscripts A and R represent active and reactive forces, respectively. The reactive bed resistance force F_{RB} in Eq. 1, defining the hydrodynamic roughness and dissipating some of the flow energy, is the term of interest in this chapter. As can be understood from this equation, the resistance of the seabed is a complicated response to many hydrodynamic forces acting on the water body and is not straightforward to determine accurately.

The Navier-Stokes equation is only solvable by numerical modeling; even so the effort depends on empirical formulation of the reactive forces. Therefore, understanding the hydrodynamic roughness of the seabed requires simplification and an analytical treatment of the problem. Literature on

seabed roughness and sediment transport is very elaborate and plentiful – to the extent of some confusion – yet the results of different methods are reasonably close. This chapter is primarily based on materials presented by Bagnold (1963), Bijker (1967), Vanoni (1975), Neilsen (1992), Van Rijn (1993), Soulsby (1997), and USACE (2002). In the following sections, the state-of-art analytical approach is discussed – for currents induced by long waves such as tide, storm surge, and tsunami, for short waves, and for the combined effects of both long and short waves. The difference between the long and short waves is hydrodynamically defined in terms of the local wave length and local water depth – those waves that have lengths longer than 20 times the local water depth are treated as long waves.

Roughness in Currents

The term F_{RB} in Eq. 1 comes down to defining the bed shear stress (hydrodynamic force per unit area exerted parallel to bed). For unidirectional flow, this stress is defined in terms of an energy gradient:

$$\tau_c = \rho g h S, \quad (2)$$

where τ_c = current-induced bed shear stress, g = acceleration due to gravity, h = water depth, and S = energy gradient or slope. This is the total bed shear stress formula known in literature (e.g., Vanoni 1975; Van Rijn 1993; Soulsby 1997). The energy gradient S is not relevant in an oscillatory current environment of coastal waters. The alternative is to redefine the total bed shear stress empirically. This empirical rendering rooted in the Bernoulli (Daniel Bernoulli, 1700–1782) theorem is known as the quadratic stress law:

$$\tau_c = \rho C_d U^2 = \rho u_*^2, \quad (3)$$

where C_d = total drag roughness coefficient and U = depth-averaged velocity. The second part the equation shows the current bed shear stress as a function of bed shear velocity u_* . The bed shear velocity is not a measurable quantity – it is a conceptual term to visualize bed shear stress on a velocity scale. Equation 3 leads one to define the bed shear velocity as

$$u_* = U \sqrt{C_d} \quad \text{or} \quad \frac{u_*}{U} = \sqrt{C_d}. \quad (4)$$

The total drag roughness coefficient C_d is related to the roughness defined in terms of granular texture as skin friction and irregularity of the bed. The pioneering work in this regard was done by Nikuradse (1933) who proposed an equivalent or effective roughness height k_s based on grain size and bed irregularity. The Nikuradse concept is elaborated in many textbooks including the ones by Vanoni (1975), Neilsen (1992), Van Rijn (1993), Soulsby (1997), and USACE (2002). The Nikuradse grain-related roughness

height is defined in terms of characteristic bed-material particle sizes such as the 50th (D_{50}) and 90th (D_{90}) percentiles. The roughness associated with bed irregularity is defined in terms of bedform height and length. Field measurements indicate that the equivalent roughness of alluvial beds decreases as the hydrodynamic forcing is increased (Van Rijn 1993).

Equivalence of the Current Roughness Coefficients

Apart from the drag resistance coefficient C_d , there are at least three other popular bed resistance or roughness coefficients in hydraulics. These are the Chezy C (Antoine Chezy, 1718–1798), Manning-Strickler n (Robert Manning, 1816–1897; Albert Strickler, 1887–1963), and Darcy-Weisbach f (Henry Darcy, 1803–1858; Julius Ludwig Weisbach, 1806–1871). It is important to relate C_d with these three coefficients. The Chezy C , Manning-Strickler n , and Darcy-Weisbach f are defined in terms of the depth-averaged velocity for a wide free-water surface flow in SI unit as in Eq. 5 (e.g., Van Rijn 1993):

$$U = \frac{1}{n} h^{2/3} S^{1/2} = C h^{1/2} S^{1/2} = \sqrt{\frac{8g}{f}} h^{1/2} S^{1/2}. \quad (5)$$

Using Eqs. 2 and 3,

$$U^2 = \frac{g}{C_d} h S, \quad (6)$$

and together with Eq. 5, one can obtain Eq. 7

$$U = \sqrt{\frac{g}{C_d}} \sqrt{h S} = \frac{h^{1/6}}{n} \sqrt{h S} = C \sqrt{h S} = \sqrt{\frac{8g}{f}} \sqrt{h S}. \quad (7)$$

Equation 7 leads to the equivalence of the four resistance coefficients in SI unit as

$$C_d = \frac{f}{8} = \frac{g}{C^2} = \frac{gn^2}{h^{1/3}}. \quad (8)$$

The equivalent total bed resistance coefficients defined by Eq. 8 can only be determined if one knows the hydrodynamic parameters. But once one of them is known, the others can be determined easily. The bed resistance coefficients C_d and f are unit less, while C ($\text{m}^{1/2}/\text{s}$) and n ($\text{s}/\text{m}^{1/3}$) are dimensional.

Division of the Total Current Resistance Coefficients

Contribution to the total bed shear stress by granular or skin friction and bedform resistance gives a theoretical clue as to how the roughness coefficients can be related. Assuming that the two are aligned, the total bed shear stress is taken as the sum of the two contributions as

$$\tau_c = \tau_{cg} + \tau_{cf}, \quad (9)$$

where τ_c = the total bed shear stress given as a summation of granular (τ_{cg}) and bedform (τ_{cf}) components.

Utilizing the definition of bed shear stress (Eq. 3), Eq. 9 yields

$$C_d = C_{dg} + C_{df}, \quad (10)$$

where C_{dg} and C_{df} are grain-related and bedform-related drag coefficients, respectively. The relationships for the other coefficients follow from Eq. 8:

$$\frac{1}{C^2} = \frac{1}{C_g^2} + \frac{1}{C_f^2}, \quad (11)$$

$$n^2 = n_g^2 + n_f^2, \quad (12)$$

$$f = f_g + f_f. \quad (13)$$

In these relations the subscripts g and f represent the grain-related and bedform-related roughness coefficients. The total Nikuradse effective roughness height is similar to C_d and f as in Eqs. 10 and 13. Bijker (1967) used the ratio of the total and grain-related roughness coefficients as an efficiency factor to estimate suspended load transport.

The estimates of the above hydrodynamic roughness coefficients are not straightforward, requiring rather an iterative process of computation. Among the various methods, the Engelund-Hansen (1967) iterative method is easier than the others, yet the results are comparatively close (see Van Rijn 1993). Table 1 shows the estimates of the roughness coefficients for four different classes of sand-sized particles using the Engelund-Hansen method and the Manning-Strickler

law. They represent a depth-averaged tidal current of 1 m/s at a water depth of 5 m. The near-bed hydraulic regime in each case represents rough turbulent flow, and an estimate following Soulsby (1983) suggests that the current boundary layer thickness asymptotes up to the water surface. In all the cases, the seabed roughness is dominated by bedform resistance, accounting for more than 75% of the total bed resistance. The bed shear velocity varies from 5% to 9% of the depth-averaged velocity.

The dependence of Eq. 3 on the depth-averaged velocity often poses practical difficulty in estimating the bed shear stress. This led Sternberg (1968) to propose

$$\tau_c = \rho C_{100} U_{100}^2, \quad (14)$$

where the drag coefficient and the measured current represent the water column height at 100 cm above the bed. Examples of this method can be found in Sternberg (1968), Heathershaw (1981), Barua et al. (1994), and Soulsby (1997). In the absence of specific information and as a first estimate, C_{100} can be taken 1.5 times the C_d .

Roughness in Waves

Compared to the current roughness, the literature on wave seabed roughness is rather limited. One reason is that in areas of vigorous wave action such as the surf zone, the energy dissipated by bed resistance is negligible compared to the dissipation of energy by wave breaking. Whatever bed resistance is present in such areas, it is dominated by skin friction of the granular bed material, often in the sheet flow regime (the sheet flow is a thin bed-material layer of very high concentration).

Seabed Roughness of Coastal Waters,

Table 1 The total, grain-related and bedform-related hydrodynamic roughness for four different sand-sized seabed granular textures in a sea-current environment at a water depth $h = 5$ m and a depth-averaged velocity $U = 1$ m/s, with seawater density = 1025 kg/m³, sediment density = 2650 kg/m³, and kinematic viscosity = 1×10^{-6} m²/s

	Case I	Case II	Case III	Case IV
	D ₅₀ = 1.00 mm	D ₅₀ = 0.50 mm	D ₅₀ = 0.20 mm	D ₅₀ = 0.11 mm
	D ₉₀ = 2.50 mm	D ₉₀ = 1.25 mm	D ₉₀ = 0.50 mm	D ₉₀ = 0.275 mm
Total				
C_d (—)	0.0072	0.0054	0.0034	0.0024
C (m ^{1/2} /s)	36.88	42.47	54.14	64.20
n (s/m ^{1/3})	0.0355	0.0308	0.0242	0.0204
f (—)	0.0577	0.0435	0.0268	0.0190
Grain related				
C_{dg} (—)	0.0012	0.0010	0.0007	0.0006
C_g (m ^{1/2} /s)	90.01	101.04	117.71	130.04
n_g (s/m ^{1/3})	0.0145	0.0129	0.0111	0.0101
f_g (—)	0.0097	0.0077	0.0057	0.0046
Bedform related				
C_{df} (—)	0.0060	0.0045	0.0026	0.0018
C_f (m ^{1/2} /s)	40.43	46.81	60.98	73.83
n_f (s/m ^{1/3})	0.0324	0.0279	0.0215	0.0177
f_f (—)	0.0480	0.0358	0.0211	0.0144

For the cases where bedform resistance appears important, the total wave seabed roughness is defined as the summation of the Nikuradse effective roughness heights similar to the situations with the current regime discussed earlier. This chapter briefly describes the wave roughness provided by bed-material skin friction. For more on the subject, the readers are referred to Neilsen (1992), Van Rijn (1993), Soulsby (1997), and USACE (2002). The shear stress exerted on the seabed as proposed by Jonsson (1966) is similar like the quadratic stress formula for current (Eq. 3):

$$\tau_w = \frac{1}{2} \rho f_w U_w^2, \quad (15)$$

where τ_w = wave bed shear stress amplitude, f_w = wave friction factor, and U_w = horizontal bottom orbital velocity amplitude. The exact estimate of the bottom orbital velocity for natural spectral waves and for asymmetric nonlinear waves requires advanced wave analysis. For simplicity one can apply the linear wave theory to estimate the first-order magnitude of horizontal bottom velocity amplitude. Following USACE (2006), this parameter can simply be estimated as

$$U_w = \frac{HgT}{2L}, \quad (16)$$

where H = wave height, T = wave period, and L = local wave length. For spectral waves, significant wave height, H_s is used in Eq. 16 together with the spectral wave period T_p . For bed resistance and sediment transport, the root-mean-square wave height is important, replacing H with $H_{rms} = H_s/\sqrt{2}$. The computation of the local wave length can either be performed

iteratively or by a simple yet accurate relation proposed by Hunt (1979).

The method for estimating the wave friction factor has been proposed by many investigators (e.g., Swart 1974; Grant and Madsen 1986; Myrhaug 1989; Neilsen 1992; Van Rijn 1993; Soulsby 1997; USACE 2002). For simplicity the friction factor proposed by Swart (1974) can be used:

$$f_w = \exp \left[5.213 \left(\frac{k_s}{A} \right)^{0.191} - 5.977 \right], \quad (17)$$

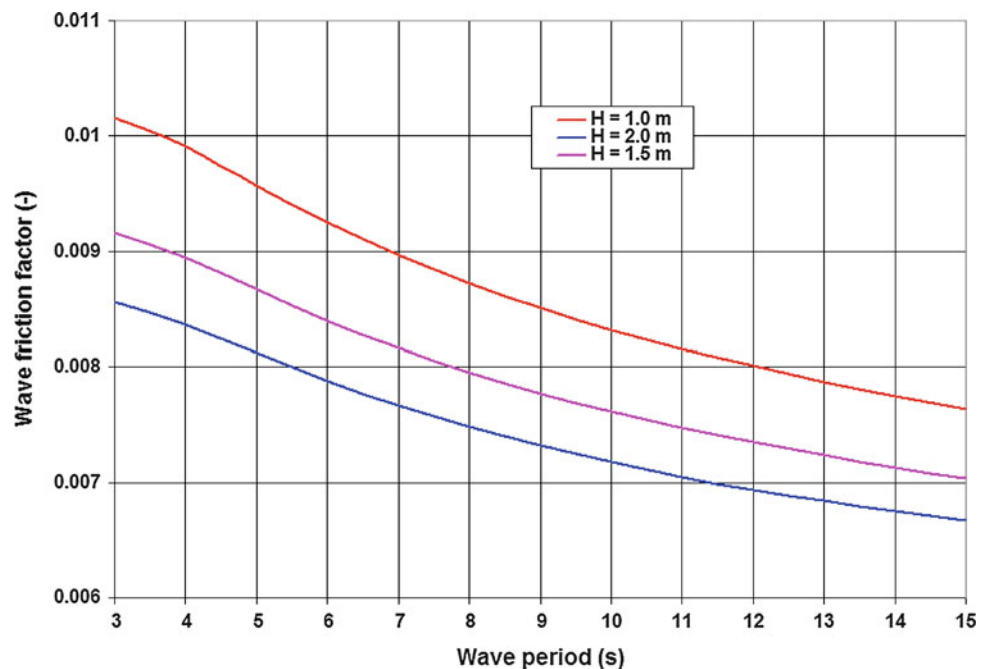
where A = amplitude of orbital velocity excursion at the bottom and k_s = Nikuradse effective bottom roughness height, which for skin friction resistance $\approx 2.5D_{50}$. For linear waves, the bottom orbital excursion amplitude is given by

$$A = \frac{U_w T}{2\pi}. \quad (18)$$

Figure 1 shows an illustration of the wave friction factors for three different wave heights 1 m, 1.5 m, and 2 m as a function of wave period at a water depth of 5 m outside the surf zone. The bed is characterized by fine sand bed-material texture with $D_{50} = 0.2$ mm. These estimates based on the simple method of Swart (1974) are surprisingly close to the complicated iterative method proposed by Grant and Madsen (1986). For the illustrated cases of rough turbulent flow, the estimated wave boundary layer thicknesses varied from 1.4 to 7.9 cm. The figure immediately demonstrates one very important characteristic of wave friction factor defining the bed shear stress – that the wave friction has a reciprocal relationship with both the wave height and the wave period.

Seabed Roughness of Coastal Waters, Fig. 1

The wave friction factor as a function of wave period for three different wave heights at a water depth $h = 5$ m. The bed material is characterized by fine sand seabed granular texture with $D_{50} = 0.2$ mm. Other properties include seawater density = 1025 kg/m^3 , sediment density = 2650 kg/m^3 , and kinematic viscosity = $1 \times 10^{-6} \text{ m}^2/\text{s}$



This behavior is similar to observations experienced in unidirectional currents where hydrodynamic bed resistance is found to decrease as the forcing is increased.

Roughness in Combined Currents and Waves

The methods for estimating the seabed roughness affected simultaneously by both currents and waves are also not robustly developed like those for currents. Instead of dynamic coupling of parameters, the available methods depend on determining the roughness and bed shear stresses individually for current and wave and then use them to estimate the combined bed shear stress. A method proposed by Soulsby (1997) is helpful in this regard to estimate the mean and the maximum combined bed shear stress as

$$\tau_m = \tau_c \left[1 + 1.2 \left(\frac{\tau_w}{\tau_c + \tau_w} \right)^{3.2} \right], \quad (19)$$

$$\tau_{\max} = \sqrt{(\tau_m + \tau_w \cos \phi)^2 + (\tau_w \sin \phi)^2}. \quad (20)$$

In these relations τ_m = mean combined bed shear stress, $\tau_{\max}$ = maximum combined bed shear stress, and ϕ = the angle between the tidal current direction and the direction of wave travel. Similar but more rigorous approaches were proposed by Grant and Madsen (1979) and Fredsøe (1984). For more on the subject, the readers are referred to Neilsen (1992), Van Rijn (1993), Soulsby (1997), and USACE (2002).

Summary and Conclusions

This chapter discusses state-of-the-art methods usually applied for determining seabed roughness of coastal waters for currents and waves and for the combined effects of currents and waves. The methods relying heavily on empirical observations and simplification of physics treat the total roughness as the contributed effects by granular skin friction and bedform resistance. Example estimates of current and wave resistance coefficients are provided to indicate their orders of magnitudes.

It is interesting to examine the discussed methods with Eq. 1. While Eq. 1 requires accounting for all different hydrodynamic terms to determine the bed resistance, the empirical methods can hardly afford to do so; instead they rely heavily on simplification and approximation of physics. The result is that the roughness remains indeterminate to a full level of accuracy. Therefore as an effort to calibrate hydraulic numerical models, this parameter is used as a tuning tool to adjust it to a suitable value. Moreover uncertainty propagation analysis presented by Barua (2015) suggests that the uncertainties associated with different primary variables such as the water depth, velocity,

density, particle size, wave height, and wave period propagate into the methods they constitute to cause such methods of estimates more uncertain than the primary variables.

Cross-References

- [Beach Features](#)
- [Beach Sediment Characteristics](#)
- [Coastal Currents](#)
- [Coastal Dynamics](#)
- [Coastal Sedimentary Facies](#)
- [Coasts, Coastlines, Shores, and Shorelines](#)
- [Cross-Shore Sediment Transport](#)
- [Nearshore Sediment Transport Measurement](#)
- [Sandy Coasts](#)
- [Sediment Suspension by Waves](#)
- [Shelf Processes](#)
- [Surf Zone Processes](#)

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Sea-Level and Climate Change

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Definition

Global mean sea level (GMSL, the height of the ocean's surface relative to land and averaged over the globe) rises when more mass is added to the oceans than is lost to glaciers, ice sheets, or other land-water storages or when ocean water warms and expands (thermal expansion). It falls when the net flow of mass is reversed or the oceans cool and contract. Changes in the shape of the ocean basins can also affect sea level. Regionally, sea-level change relative to the land (relative sea level, RSL) can be larger or smaller than the global average because of changes in ocean circulation (with associated changes in ocean temperatures and salinity) or vertical motion of the land, including local subsidence or uplift. Changes in relative sea level are also brought about by ongoing changes in the Earth's shape, gravity, and rotation resulting (in part) from altered distributions of water and ice in the recent and distant past. Locally, subsidence may occur as a result of compaction of sediments (particularly in deltaic regions and particularly following the withdrawal of groundwater). Geocentric sea level (also termed sea surface height, as measured by satellites) is the height of the ocean's surface relative to the center of mass of the Earth or a reference ellipsoid. Tides and storm surges affect local extreme sea levels.

Global mean sea-level change resulting from change in the mass of the ocean is called barystatic. The amount of barystatic sea-level change due to the addition or removal of a mass of water is called its sea-level equivalent. Sea-level changes, both globally and locally, resulting from changes in water density are called steric. Density changes induced by

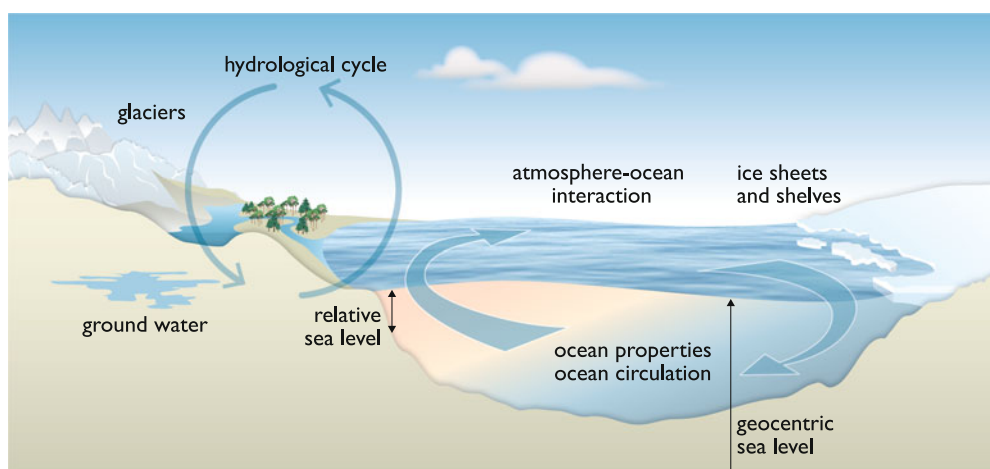
temperature changes only are called thermosteric, while density changes induced by salinity changes are called halosteric. Barystatic and steric sea-level changes do not include the effect of changes in the shape of ocean basins induced by the change in the ocean mass and its distribution.

Introduction

Globally, many of the world's mega cities are situated on the coast, with the order of 100 million people living within 1 m of current high tide level and over a trillion dollars of infrastructure in this region. With people continuing to live ever closer to the coast, rising sea levels are a serious issue. Particularly vulnerable are people living on deltas (e.g., millions in Bangladesh, Vietnam, and elsewhere) and on small low-lying islands in the Pacific and elsewhere.

Sea level changes over a broad range of space and time scales. This entry focuses on global to regional and local sea levels and on how mean sea level (averaging over waves and tides) has changed over the historical period and how it is likely to respond to natural climate variability and anthropogenic climate change over time scales from months to centuries and millennia. Past records of sea level over longer periods are also considered to help understand how sea level may evolve in our warming climate. Sea-level extreme events resulting from storm surges, tides, and waves are also briefly considered. See Church et al. (2010) for a thorough review of all of the relevant issues.

The height of the ocean surface can be measured relative to the surface of the Earth (relative sea level, RSL, Fig. 1) or relative to some geocentric reference level such as the center of mass of the Earth or a reference ellipsoid (geocentric sea level, also termed sea surface height, and sometimes referred to as absolute sea level). RSL is important for coastal impacts including coastal flooding, inundation, and erosion. Beginning in the seventeenth century, RSL was measured by tide gauges, mainly to support shipping services. Comprehensive databases of monthly average coastal tide gauge data are available from the Permanent Service for Mean Sea Level (PSMSL, Holgate et al. 2013), with higher frequency (hourly or more frequent) data available from the University of Hawaii Sea Level Centre. Prehistoric RSLs are also inferred by various geological indicators such as the heights of preserved coral reefs, sediment layers, and coastal rock platforms. Today, sea levels are also measured relative to the center of the Earth by satellites, giving near global coverage (see Stammer and Cazenave 2017 for a recent review). These data sets mostly start in 1993 and measure geocentric sea level. There are several websites where satellite altimeter data can be



Sea-Level and Climate Change, Fig. 1 Climate-sensitive processes and components that can influence global and regional sea level. Changes in any one of the components or processes shown will result in a sea level change. The term “ocean properties” refers to ocean

temperature, salinity, and density, which influence and are dependent on ocean circulation. Both relative and geocentric sea levels vary with position. (From Church et al. 2013, Fig. 13.1)

obtained, including NASA, AVISO, NOAA, ESA, the University of Colorado, and CSIRO.

Over periods of years to millennia, changes in global mean sea level (GMSL) are largely the result of changes in the volume of ocean water from expansion or contraction as it warms or cools (thermometric sea-level change) and changes in the mass of the ocean (barystatic sea-level change) as water is transferred between the ocean and ice sheets, glaciers, and other terrestrial water storages (e.g., lakes, reservoirs, aquifers).

Regional sea-level change may differ from the global average because of climate variability, changes in ocean currents and regional atmospheric pressure, the vertical movement of land, and changes in the Earth’s gravitational and rotational fields as a result of the redistribution of water, particularly from ice sheets. Local processes can also be important for changes in sea level relative to the land and may dominate at some locations. These processes include local subsidence resulting from the extraction of ground water or hydrocarbons and the compaction of sediments. Variations in sediment supply can affect local erosion/accretion of the coastline.

Past Sea-Level Change

Over the glacial cycles of the past 800,000 years, sea level varied by more than 120 m (Rohling et al. 2009) as major ice sheets waxed and waned over North America, northern Europe, Asia, and Antarctica. As a result of these larger ice sheets during ice ages, sea level fell to more than 120 m below present-day values.

During Warmer Periods Sea Level Was Higher

Of greater importance for understanding future sea levels is what height were sea levels in periods of a warmer climate. Here, the evidence is clear that in past warmer climates, sea level was meters higher than present-day values (Dutton et al. 2015). The best evidence comes from the last interglacial (warm) period (129,000–116,000 years ago). Here the evidence is for sea levels 6–9 m above present-day values at global average temperatures about 1 °C warmer than pre-industrial temperatures (present-day temperatures are close to this value) and with significant polar amplification of surface temperature increases. Evidence supports a partial but not complete retreat of the Greenland ice sheet during this period. The sizes of the estimated contributions from glaciers, ocean thermal expansion, and the Greenland ice sheet are insufficient to explain the higher sea levels (particularly their upper end values) implying a significant Antarctic contribution. Rates of rise during the period when sea level was above present-day values remain poorly constrained.

Sea-level data from the long interglacial period from 424,000 to 395,000 years ago also indicate higher sea levels with moderately warmer global temperatures (Dutton et al. 2015). The last time atmospheric CO₂ concentrations were similar to today’s levels was in the Mid-Pliocene warm period about 3.2–3 million years ago. Here, the evidence is less robust but indicates sea levels at least 6 m above present-day values with global temperatures about 2–3 °C above preindustrial values (Dutton et al. 2015). The upper end of the higher sea-level range remains poorly defined.

From the Last Glacial Maximum to the Holocene

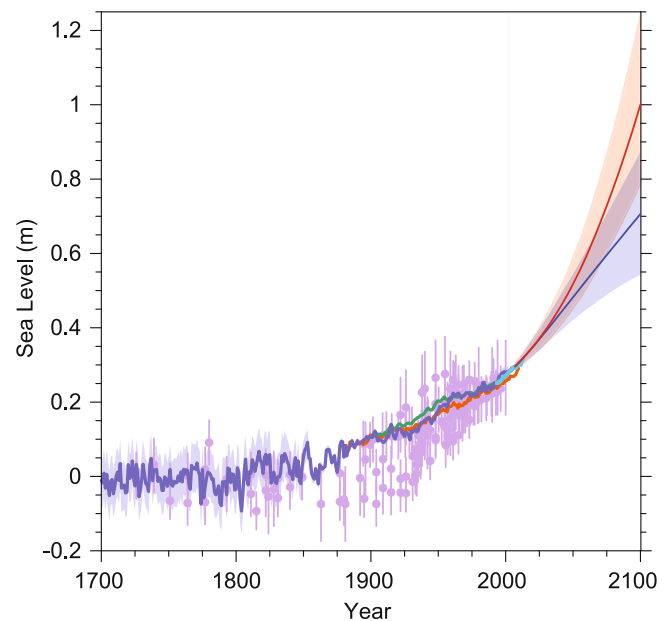
At the time of the last glacial maximum (29,000–21,000 years ago), sea level was estimated to be about 130 m

below present-day values (Lambeck et al. 2014). After this, sea level rose at rates of over 1 m/century for many thousands of years (with peak rates of about 4 m/century) as the Northern Hemisphere ice sheets decayed, with a significant contribution from the Antarctic ice sheet. From 8,500 to 2,500 years ago, the rate of rise gradually decreased with little change in global mean sea level until the rate of rise began increasing in the nineteenth century (Lambeck et al. 2014). In regions formerly covered by ice sheets, sea level fell significantly over the last several thousand years as the land rebounded (Dutton et al. 2015). In contrast, in the immediately adjacent region, sea level rose as the proglacial forebulge collapsed. The collapse also resulted in sea level falling at locations distant from the ice sheets. Sea-level estimates from salt marsh cores bordering the North Atlantic suggest a multi-century sea-level oscillation of the order of decimeters over decadal to millennial time scales (Dutton et al. 2015). However, it is unclear if this apparent oscillation has a global signature. The clearest signal in the recent paleo sea-level data is the tenfold increase in the rate of rise from low values (order tenths of a mm yr^{-1}) to modern-day rates (order 2 mm yr^{-1}) during the nineteenth and early twentieth century (Dutton et al. 2015).

The Instrumental Record

Estimates of historical sea-level change using tide gauge data (Fig. 2) reviewed in the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5) (Church et al. 2013) indicate a global averaged rate of rise of $1.5 \pm 0.3 \text{ mm yr}^{-1}$ from 1900 to 1990. Hay et al. (2015) improved the approach for estimating global mean sea-level change by incorporating regional “fingerprints” associated with loss of mass from glaciers and the ice sheets. Their estimate of GMSL rise for the same period (1900–1990) was about 20% smaller at $1.2 \pm 0.3 \text{ mm yr}^{-1}$, principally because of lower rates over the 1950–1970 period. However, these lower estimates remain controversial. Since 1970, the various estimates are in better agreement with each other and with the satellite altimeter data. Estimates of GMSL that start in the nineteenth century indicate an acceleration in the rate of rise (Church and White 2011; Clark et al. 2015).

The satellite altimeter record is available since the start of 1993 and continuing to the present provides near global and significantly more accurate estimates of sea-level change. After correction for an important bias in the first 6 years of the record, the rate of rise is now estimated at $3.1 \pm 0.3 \text{ mm yr}^{-1}$ and accelerated over this period at a rate of about 0.1 mm yr^{-2} (WCRP Sea Level Budget Group 2018). The satellite altimeter data (Fig. 3) indicates a faster rate of rise in the western Pacific and Indian Ocean than in the Eastern Pacific since the start of the record in 1993.



Sea-Level and Climate Change, Fig. 2 Global average sea level has increased from estimated preindustrial levels and is projected to rise at a faster rate during the twenty-first century. The blue, orange, and green curves up to 2010 are different estimates of global average sea-level change, relative to the preindustrial level (left-hand axis), based on historical tide gauge observations. The light blue curve is the satellite altimeter observations from 1993 to 2012. Projections, shown from 2006 to 2100, are relative to the average over 1986–2005 for high and low greenhouse gas emission pathways. The shading indicates the *likely* range of rise, but higher values are possible from a poorly understood contribution from the Antarctic ice sheet. (From Church et al. 2013, Fig. 13.27)

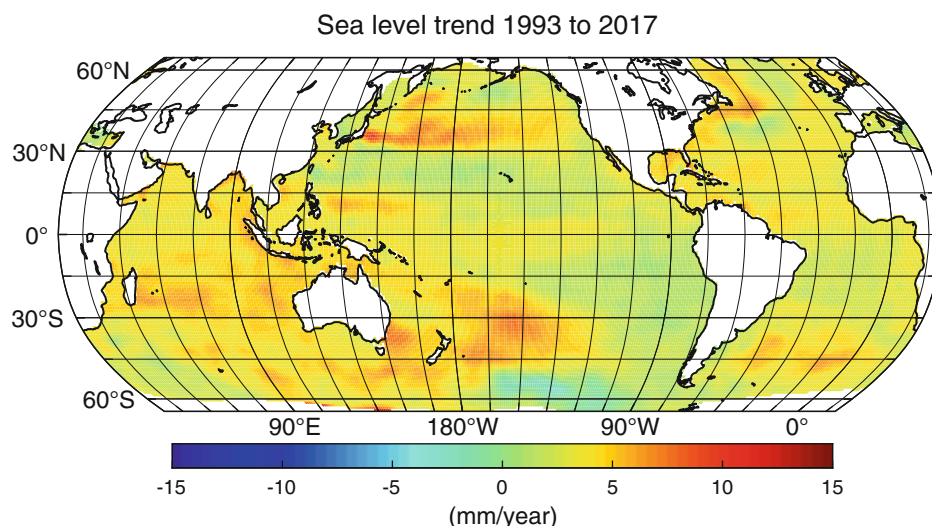
Understanding Observed Sea-Level Change

Understanding the reasons for the twentieth century sea-level change has been a challenge for many decades. The recent progress in understanding the rise since 1970 (Church et al. 2011), since 1900 (Gregory et al. 2012; Church et al. 2013), and since 1993 and 2005 (WCRP Sea Level Budget Group 2018) has been made by including all contributions to sea-level change and by using a combination of models and observations. The results indicate the two largest contributions over all of these periods have been the expansion of ocean water as it warmed and the addition of water to the ocean from the loss of ice from glaciers. Over the recent decades, there has been an accelerating contribution from the Greenland and Antarctic ice sheets.

Historical ocean observations have many gaps in the spatial coverage, particularly in the Southern Hemisphere, prior to the implementation of the Argo observational network in the early 2000s. Despite this, useful observational estimates of global average thermal expansion are available since the 1960s. Correcting instrumental biases in expendable bathythermograph (XBT) observations resulted in significant

Sea-Level and Climate Change,

Fig. 3 Rate of geocentric sea-level change (mm/yr) for 1993–2017, as measured by the TOPEX/Poseidon and Jason-n series of satellite altimeters. (Source: CSIRO Oceans and Atmosphere (2018). Figure 3 was drawn by Dr. Quran Wu, drawing on satellite altimeter data processed by CSIRO)



improvement in the agreement between observational estimates and model results. The observations indicate that much of the ocean thermal expansion comes from the upper 700 m of the ocean, with little global averaged change between 2,000 m and 3,000 m, and then additional warming in the abyssal ocean (principally below 4,000 m), particularly in the Southern Ocean. The Argo observations also indicate significant thermal expansion at depth of the order of 1,000 m and in the Southern Hemisphere at latitudes of 30–50 °S (Wijffels et al. 2016). Violent volcanic eruptions that inject aerosols into the stratosphere result in significant variability in the rate of ocean thermal expansion (several mm of sea level fall for about 2 years after the eruption followed by a slower recovery) and can bias model results for the twentieth century if the volcanic forcing for both the twentieth century and the preceding centuries is not considered.

Estimates of glacier contributions have improved significantly following the completion of the Randolph Glacier Inventory. A significant part of the early twentieth century glacier contribution is a response to the pre-twentieth century climate changes and potentially regional climate variability in the first half of the twentieth century. However, anthropogenic forcing dominates the glacier contribution since about 1970 (Marzeion et al. 2014).

The sum of storage of water in terrestrial reservoirs and the depletion of ground water have made a significant contribution to sea-level change during the twentieth century. Short-term climatic variations, such as the El Niño phenomenon, lead to changes in global sea level as water is stored in flood waters, or as lands dry during droughts, but these variations are short term.

Historical observations indicate that there was a significant contribution from the Greenland ice sheet in the early twentieth century that may relate to the pre- and early

twentieth century climate change. Since 1990, there have been further contributions from surface melting of the Greenland ice sheet. Much attention has been focused on the increased discharge of ice into the ocean from both the Greenland and Antarctic ice sheets since 1990 (see, e.g., the IMBIE Group 2018). The simulation of this increased solid ice discharge and its rapid evolution, particularly in Antarctica, remains challenging. It is directly related to increases in ocean temperatures underneath and adjacent to the floating ice shelves and glacier tongues that fringe the coast of Antarctica and Greenland. The mass loss from the ice sheets is the main reason for the acceleration of the rate of rise since 1993 (WCRP Sea Level Budget Group 2018).

Present-day model simulations of global mean and regional sea level only explain about 50% of the observed rise since 1900. However, when known additional non-anthropogenic climate change contributions (such as the early twentieth-century Greenland and glacier contributions) and the post-1993 dynamic response of the ice sheets are added, about 75% of the observed rise is explained, and essentially all of the rise since 1993 (Slangen et al. 2017; Meyssignac et al. 2017). Much of the regional variability is a result of land motion, particularly from the regional distribution of glacial isostatic adjustment. Anthropogenic climate change is the dominant reason of sea-level rise since about 1970, explaining about 70% of the observed rise (Slangen et al. 2016).

Note that climate fluctuations also lead to large local sea-level variability as water is moved around the ocean in response to changes in surface winds, atmospheric pressure, and transfer of heat and freshwater between the atmosphere and ocean. Much of the sea-level rise pattern present in Fig. 3 is probably associated with this natural climate variability. However, the high rates of rise at the poleward edges of the subtropical gyres (about 40 °N and

S) are likely associated with changes in surface wind patterns that relate at least in part to anthropogenic climate change signals. Note current climate models (as used in current predictions of climate variability) have significant and useful skill in predicting seasonal to interannual variations in regional sea-level fluctuations in the tropical Pacific Ocean and along the west coast of North and South America and Australia.

Projections of Future Sea-Level Change

Probabilistic projections of future sea-level change were evaluated using the same climate models and off-line models for glacier, ice sheet, and terrestrial water storage models that have been used to understand the historical rise. The major challenge in formulating the most recent IPCC projections (Church et al. 2013) was estimating the contributions for the solid ice discharge for both the Greenland and Antarctic ice sheets. For Greenland, there were dynamical simulations available, but these were not for the full range of future greenhouse gas emission scenarios. For Antarctica, no dynamic ice sheet simulations were available, so the ice discharge was estimated using a probabilistic framework scaled to available observations. However, no information on future greenhouse gas emission scenario dependence was available.

Projections of Global Mean Sea Level

For all scenarios considered, sea levels are projected to rise at a faster rate during the twenty-first century than during the twentieth century. By 2100, the IPCC AR5 projections were that sea level will *likely* (the central 66% of the probabilities) rise by a global average of 28–61 cm relative to the average level over 1986–2005 if greenhouse gas emissions are low (strong mitigation under Representation Concentration Pathway 2.6 (RCP2.6)) and by 52–98 cm if emissions are high (~business-as-usual, RCP8.5; Fig. 2). For the high emission scenario, the rate of rise increases to between 7.5 and 15.7 cm/decade over the last two decades of the twenty-first century, similar to average rates of sea-level rise at the end of the last ice age, and a factor of 5 larger than the early twentieth century rates. The largest contributions are projected to be ocean thermal expansion and the loss of ice from glaciers. Increased surface melting of the Greenland ice sheet and increased flow of ice from both Greenland and Antarctica will also contribute. The AR5 projections recognized that the relevant ocean and Antarctic ice sheet interactions are largely responsible for the control of the discharge of solid ice are poorly understood. As a result, they noted (with medium confidence) that the collapse of marine-based sectors of the Antarctic ice sheet, if initiated, could lead to an additional rise of several tenths of a meter by

2100. However, little is known about the probability distribution of projected sea-level rise, particularly the Antarctic contribution. Hence, projections of the tails of the distribution (as in some recent extrapolations) have little quantitative credibility.

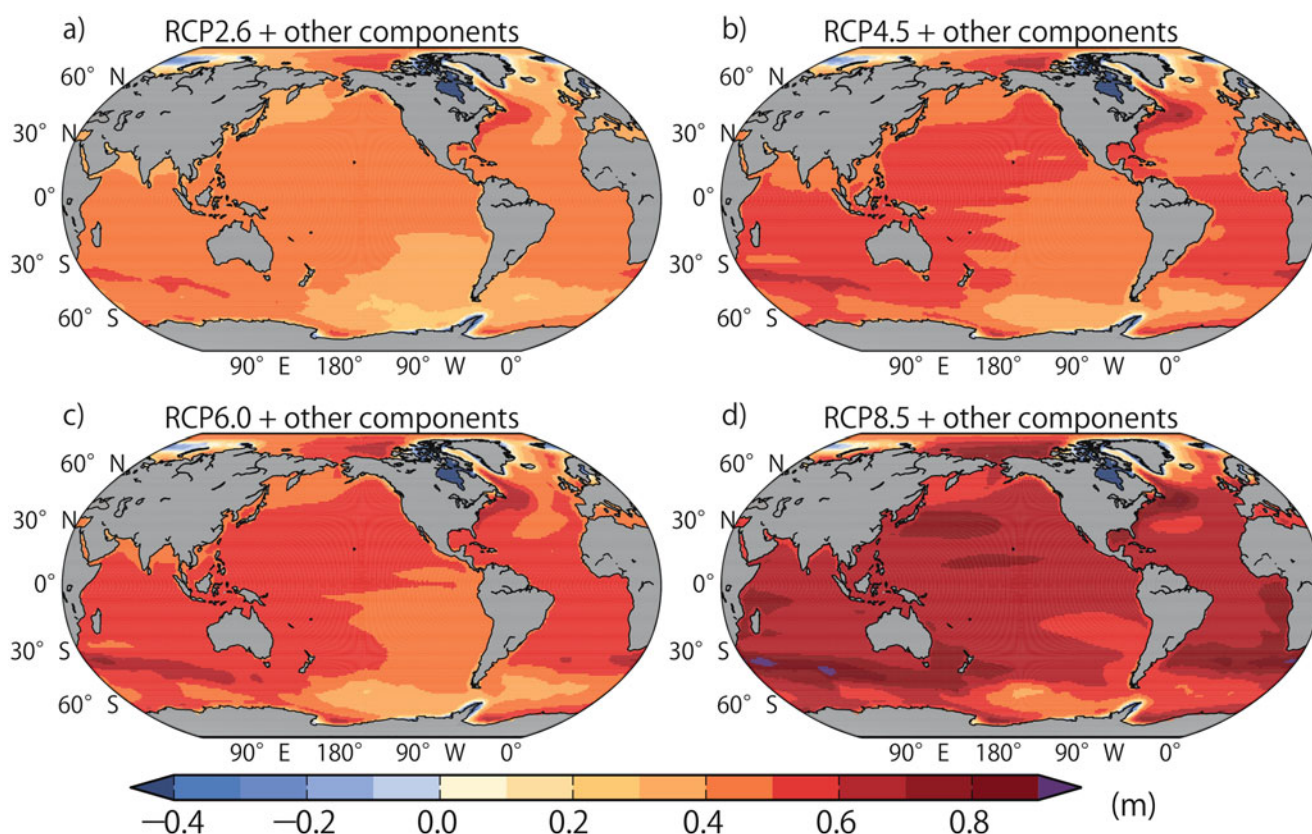
The most important weakness of these projections is the Antarctic contribution. Recent observations and models suggest that parts of the West Antarctic ice sheet that are resting on rock well below sea level may have entered a phase of ongoing mass loss. There have been a range of recent Antarctic ice sheet modeling studies. Some of these support the estimates used in the IPCC AR5 projections for a *likely* ice sheet discharge by 2100 of –1 to 16 cm, or only a moderate increase from them. However, one recent controversial study (DeConto and Pollard 2016) indicates an Antarctic contribution of the order of a meter by 2100.

Projections of Regional Sea-Level Change

Projections of regional sea-level change differ from the global average sea-level rise for several reasons. Firstly, changes in the air-sea fluxes of momentum, heat, and water vapor will result in changes in ocean circulation and regional sea-level change. Secondly, loss of mass from glaciers and ice sheets and changes in the mass of land-water storage result in a change in the Earth's gravitational and rotation fields and the surface loading of the Earth. As a result, the mass loss can contribute to a sea-level fall near regions of mass loss, while in the far field the sea-level rise may be larger (by up to about 30%) than the global averaged rise. Thirdly, there is ongoing vertical movement of the land as a result of changes in the mass of glaciers and ice sheets over past millennia. Fourthly, there may be local vertical land motion associated with sediment compaction (particularly following the withdrawal of ground water or petroleum) or other local tectonic motions. The first three of these processes were incorporated in the probabilistic regional projections of the IPCC AR5 (Church et al. 2013, Fig. 4), but the fourth was not as a result of its strong local and regional variability. The resulting projections to 2100 imply that about 95% of the global ocean will experience a sea-level rise, with only small regions near the current and former ice sheets potentially experiencing a sea-level fall. About 70% of the global coastlines are projected to experience a sea-level rise within 20% of the global mean change (Church et al. 2013).

Sea-Level Rise Will Continue for Centuries to Millennia

Because of the long time scales of oceans and ice sheets, sea level will continue to rise for many centuries under all emission scenarios (Church et al. 2013) and persist for millennia (Clark et al. 2016). For these long time scales, global mean ocean thermal expansion is estimated to be slightly more than 0.4 m for each degree of global averaged



Sea-Level and Climate Change, Fig. 4 Ensemble mean regional relative sea-level change (meters) evaluated from 21 CMIP5 models for RCP scenarios (a) RCP2.7, (b) RCP4.5, (c) RCP6.0, and (d)

RCP8.5 between 1986–2005 and 2018–2100. Each map includes effects of atmospheric loading, plus land ice, glacial isostatic adjustment (GIA), and terrestrial water sources. (From Church et al. 2013, Fig. 13.20)

warming. Glaciers contain less than 0.5 m of sea-level equivalent, thus limiting their long-term contribution. Sustained warming above a threshold would lead to the near-complete loss of the Greenland ice sheet over a thousand years or more, contributing up to about 7 m to global average sea-level rise. The warming threshold was estimated in the IPCC AR5 to be between 1 °C and 4 °C above preindustrial temperatures. Crossing this threshold, committing the world to this rise, could occur well before 2100 for unmitigated greenhouse gas emissions.

The largest scientific uncertainties in projecting future sea-level rise relate to the dynamic response of the Antarctic ice sheet to warming. Our understanding is currently insufficient to accurately constrain the timing or magnitude of such a multi-century contribution. With strong greenhouse gas mitigation, some projections indicate a relatively small multi-century sea-level contribution. However, there are some indications that parts of the West Antarctic ice sheet may have already crossed a threshold that would result in an ongoing decay and a sea-level contribution of meters. With significant parts of the East Antarctic ice sheet also grounded below sea-level, unmitigated emission, this century will

commit the world to sea levels tens of meters higher than today for millennia (Clark et al. 2016).

Impacts

Rising Sea Levels Result in a Greater Coastal Flood and Erosion Risk

Rising sea levels result in an increased intensity and frequency of coastal flooding, inundation, erosion, and impacts on coastal ecosystems, with the greatest impacts felt during extreme events (e.g., from storm surges and large waves; Wong et al. 2014). Rising average sea levels mean that extreme sea levels of a particular height are exceeded more often. This has already happened in many parts of the world and will continue. Even for the lowest projections, there would be substantially more than a tenfold increase in the frequency of extreme sea-level events by 2100 for most locations around the world, resulting in a much higher risk of coastal flooding and erosion during the twenty-first century and beyond. We will need to adapt to these impacts even for a low (strong mitigation) greenhouse gas emission

scenario. The increase in the frequency of coastal flooding events and their impact will be substantially larger for higher emissions.

Summary

Sea level changes on a wide range of time and space scales. Changes in global mean sea level are largely the result of changes in the volume of ocean water from expansion or contraction as it warms or cools (thermosteric sea-level change) and changes in the mass of the ocean (barystatic sea-level change) as water is transferred between the ocean and ice sheets, glaciers, and other terrestrial water storages. Regional sea-level change may differ from the global average because of climate variability, changes in ocean currents and regional atmospheric pressure, the vertical movement of land, and changes in the Earth's gravitational and rotational fields as a result of the redistribution of water, particularly from ice sheets. Local processes can also be important for changes in sea level relative to the land and may dominate at some locations.

Over the glacial cycles of the past 800,000 years, sea level varied by more than 120 m as major ice sheets waxed and waned over North America, northern Europe, Asia, and Antarctica. In past warmer climates, sea level was meters higher than present-day values at temperatures similar to or slightly above present-day values. After the last glacial maximum, sea level rose at rates of over 1 m/century for many thousands of years (with peak rates of about 4 m/century). There was little change in global mean sea level over the several millenia until there was a tenfold increase in the rate of rise from low values (order tenths of a mm yr^{-1}) to modern-day rates (order 2 mm yr^{-1}) during the nineteenth and early twentieth century and about 3 mm yr^{-1} in the satellite record since 1993. The two largest contributions over the twentieth century have been the expansion of ocean water as it warmed and the addition of water to the ocean from the loss of ice from glaciers. Recent increases in the contributions from Greenland and Antarctica have resulted in an acceleration in the rate of rise over the recent decades. Anthropogenic climate change is the dominant reason of sea-level rise since about 1970, explaining about 70% of the observed rise.

For all potential greenhouse gas scenarios considered in the IPCC AR5, sea levels are projected to rise at a faster rate during the twenty-first century than during the twentieth century. By 2100, it is currently projected that sea level will *likely* rise by a global average of 28–61 cm relative to the average level over 1986–2005 if greenhouse gas emissions are low (strong mitigation under RCP2.6) and by 52–98 cm if emissions are high (~business-as-usual, RCP8.5). For the high emission scenario, the rate of rise

increases to between 7.5 and 15.7 cm/decade over the last two decades of the twenty-first century, similar to average rates of sea-level rise at the end of the last ice age. The collapse of marine-based sectors of the Antarctic ice sheet could lead to an additional rise of several tenths of a meter by 2100. Projections of regional sea-level change to 2100 imply that about 95% of the global ocean will experience a sea-level rise, with only small regions near the current and former ice sheets potentially experiencing a sea-level fall. About 70% of the global coastlines are projected to experience a sea-level rise within 20% of the global mean change.

Because of the long time scales of oceans and ice sheets, sea level will continue to rise for many centuries under all emission scenarios and persist for millennia. Unmitigated emission this century will commit the world to sea levels tens of meters higher than present for millennia.

Strong mitigation is necessary to avoid the worst impacts of sea-level rise. However, we cannot avoid all sea-level change, and it will be necessary to adapt to the change we cannot avoid. Rising sea levels will result in an increased intensity and frequency of coastal flooding, inundation, erosion, and impacts on coastal ecosystems, with the greatest impacts felt during extreme events.

Cross-References

- Beach Erosion
- Isostasy
- Sea-Level Indicators, Biologic
- Sea-Level Indicators, Geomorphic
- Sea-Level Indicators: Biological in Depositional Sequences
- Storm Surge
- Tide Gauges

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Sea-Level Fluctuations Over the Last Millennium

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Definition

Sea levels fluctuate continually in response to numerous forcing functions on differing spatial and temporal scales. On long timescales (millennial and longer), sea levels are governed principally by the Earth's orbital variations which in turn alter temperatures in the atmosphere affecting long timescale mass balance of the world's glaciers, ice sheets, and oceans. In addition to these background processes operating over the last millennium, sea-level rise has also been fundamentally linked to increased accumulation of so-called “greenhouse” gases in the atmosphere since industrialization (post ≈ 1750) resulting in global warming.

Introduction

Longer-term fluctuations in mean sea level are related to natural cycles of warming and cooling of the planet resulting in the advancement and retreat of ice sheets and glaciers. These glacial and interglacial (G-IG) periods are strongly associated with cyclical variations in the orbital eccentricity, obliquity, and precession of the Earth around the Sun, pronounced in the published climate theories of astrophysicist Milutin Milankovitch (Milankovitch 1930) and confirmed in the late 1970s by Hays et al. (1976). Although the evidence of these cycles in paleoclimate records is clear, the internal mechanisms and climate cycle feedbacks triggering the G-IG change remain the subject of scientific conjecture (Broecker and Henderson 1998; Lambeck et al. 2002).

Over the past $\approx 1\text{Myr}$, the dominant cycle is the $\approx 100\text{ kyr}$ cycle associated with the orbital eccentricity of the Earth (Lisiecki 2010), resulting in associated sea-level fluctuations between G-IG cycles ranging in the order of $\approx 150\text{ m}$ over the past 0.5Myr (Rohling et al. 1998). The influence of each G-IG cycle is unique owing to combinations of different orbital

configurations (Tzedakis et al. 2012) with the last glacial maximum occurring around 20,000 years ago with sea level about 120 m lower than present.

Sea-level changes on long timescales result from fluctuations in the volume of the ocean water mass. Firstly changes in ocean temperature and salinity result in density (or steric) changes which are inversely proportional to volume changes (Leuliette 2015). Secondly, water exchange between the ocean and continents (such as glaciers, ice caps, ice sheets, and groundwater) results in variations to the ocean's total mass (Leuliette and Willis 2011).

The extensive analysis of deep and ancient ice core samples from East Antarctica, dating as far back as ≈ 800 kyr (Petit et al. 1999; Jouzel et al. 2007), provides strong evidence of the temporal correlations and interdependencies between sea level and temperature proxies and atmospheric gas concentrations such as CO_2 (Jansen et al. 2007; Masson-Delmotte et al. 2013), though phase relations (leads/lags) among the variables might be as long as 5 kyr (Mudelsee 2001).

These identified interdependencies give rise to the additional threat forecast from anthropogenically induced climate change evident since industrialization. Over the past 650 kyr of G-IG cycles, atmospheric CO_2 levels have varied in the range of 180–300 ppm (Jansen et al. 2007) with a level of approximately 280 ppm at the commencement of industrialization (Etheridge et al. 1996), rising sharply to exceed 400 ppm in 2016 at key recording stations located at Mauna Loa, Cape Grim, and the South Pole. The Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5) noted that the rate of sea-level rise since the mid-nineteenth century has been larger than the mean rate during the previous two millennia (IPCC 2013).

Measuring Mean Sea-Level (MSL) Fluctuations

Direct systematic sea-level observations are known to date back to 1682 in Amsterdam (Gehrels and Woodworth 2013) with the period post ≈ 1700 generally referred to as the “instrumental record” (Church et al. 2013). Mean sea-level estimates predating the “instrumental record” have been estimated using a range of sea-level markers estimated from various sources. In particular numerous studies analyzing salt marsh records have estimated mean sea-level records over the past 2–3 millennia by matching fossils and sediments of known age and elevation (e.g., Gehrels et al. 2012; Gehrels and Woodworth 2013; Kemp et al. 2014).

Additionally archaeological remains of maritime infrastructure around the Mediterranean Sea dating, from the Roman Era (Lambeck et al. 2004; Auriemma and Solinas 2009), provide other lines of evidence to corroborate those derived from sedimentological and fossil records.

The first self-recording tide gauge was installed in the 1830s at the entrance to the London Docks (Palmer 1831). By the end of the nineteenth century, similar instruments had been installed in most major ports (Gehrels and Woodworth 2013) becoming increasingly sophisticated to present day. Numerous continuously recording instruments using technologies such as acoustic, radar, and pressure sensors are now commonly installed around the major ports and throughout the coastal cities of the world.

With tide gauges fixed on the land, water level records are a combination of both the movement of the water surface and vertical land motion at the site of the gauge. Since the early 1990s, the advent of global navigation satellite systems has facilitated estimates of vertical land motion at an increasing number of tide gauge records to improve their utility for sea-level research (Santamaría-Gómez et al. 2012).

From around late 1992, the advent of satellite-based radar altimeter measurements has augmented the extensive array of tide gauge records around the world with nearly global (between 66° N and S) ocean coverage at 10-day intervals (Church et al. 2013). Since 2000, the deployment of a global array of Argo autonomous hydrographic profiling floats has sampled temperature and salinity across the upper 2000 m of the global ocean to measure steric changes with improved accuracy and coverage (Leuliette and Willis 2011). Some 3800 profiling floats were in operation through mid-2017.

Augmenting altimetry and the array of Argo profiling floats, the launch in 2002 of Gravity Recovery and Climate Experiment (GRACE), a pair of satellites monitoring changes in the Earth's gravity field, provides precision estimates of the mass change in the ocean (Leuliette and Willis 2011).

Key MSL Signatures Observed over the Past Millennium

By analyzing archaeological remains from the central Tyrrhenian coast of Italy dating back to the Roman era ≈ 2000 BP, Lambeck et al. (2004) estimated eustatic sea-level rise over the interim at 0.13 ± 0.09 m. This research established the onset of modern sea-level rise (due to climate change) at $\approx 100 \pm 53$ years BP.

Gehrels and Woodworth (2013) synthesized proxy paleo-sea-level markers from seven sites around the world with tide gauge records. This study concluded that the recent rapid rate of sea-level rise began to depart from the low background rate over the late Holocene at each site sometime ≈ 1905 –1945, similar to the findings of Lambeck et al. (2004).

Considering proxy relative sea-level reconstructions over the past 3000 years and tide gauge data, Kopp et al. (2016) observed that global mean sea level varied by $\approx \pm 8$ cm over the pre-industrial Common Era with the twentieth-century

rise extremely likely faster than during any of the 27 previous centuries.

Church et al. (2013) concluded that the most robust signal captured from salt marsh records in the southern and northern hemispheres alluded to a transition from relatively low rates of global mean sea-level rise during the late Holocene (order tenths of mm per year) to modern rates (order of mm per year).

Over the more recent satellite altimetry period, unprecedented knowledge has been gleaned concerning the nature of the sea-level rise observed (Leuliette 2015). Over the period from 2005 to 2013, one-third of the observed global mean sea-level rise has been attributable to steric contributions, while two-thirds can be attributed to ocean mass increases (Leuliette 2014; Llovel et al. 2014). Similarly, the spatial distribution of sea-level rise has not been uniform over the altimetry era, with rates of rise in the warm pool of the western Pacific up to three times the global average, while rates over much of the eastern Pacific are near zero or negative (Beckley et al. 2010; Rhein et al. 2013).

Future MSL Projections

The Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5) (IPCC 2013) provides the most authoritative and up-to-date global assessment of the state of climate science, including sea-level change, past, present, and future (Church et al. 2013; Rhein et al. 2013). These processes are underpinned by an extensive and continually evolving consortia of climate modeling products. The Coupled Model Intercomparison Project (CMIP) commenced in the mid-1990s as a comparison of a handful of early global coupled climate models. In response to a growing need to systematically analyze coupled ocean and atmosphere model outputs from multiple climate modeling centers, it has subsequently grown into a large program to advance model development and scientific understanding of the Earth system (Carlson et al. 2017).

CMIP Phase 5 (CMIP5), developed in conjunction with AR5, provides the means by which to assess the differences in future model projections of dynamical sea-level changes at fine-resolution scale for the benefit of climate research, policy setting, and adaptation planning (Hurrell et al. 2011; Taylor et al. 2012).

Global climate models used to predict the physical responses of the climate system (Taylor et al. 2012; Flato et al. 2013) to various greenhouse gas emission scenarios over the course of the twenty-first century (Moss et al. 2010), forecast increases in mean sea level of up to ≈ 1 m by 2100. In the longer term, Church et al. (2013) advise that less information is available on climate change beyond the year 2100; however, the ocean and ice sheets will continue to

respond to changes in external forcing on multi-centennial to multimillennial timescales.

Controversies with the Current State of the Science Concerning Sea-Level Rise

Over recent decades, significant scientific endeavor has been invested in understanding, measuring, and modeling the greenhouse gas-induced influences on our climate system (such as increasing mean sea level and temperatures) to differentiate them from natural background influences that include climate mode forcings on short timescales (≈ 10 – 30 year) through to cyclical variations of the Earth's orbit around the Sun that drive G-IG cycles (up to ≈ 100 kyr). The emerging nature of the science in these areas, coupled with the comparatively low rate of global mean sea-level rise evident over the past century (≈ 1 – 2 mm), provides difficulties associated with isolating the relative contributions of each of these forcings from the available data and building the same into projection models of the future.

These nuances have given rise to a range of conflicting theories that have surrounded the science of sea-level rise over the past decade. For example, there are differing views on the role that solar activity (as distinct from greenhouse gases) will have on climate over the twenty-first century. The extent of influence of the current unusually long solar minimum (Rozanov et al. 2012), through to a possible future grand minimum by 2050 (Ineson et al. 2015), could result in cooling sufficient to offset predicted anthropogenic warming ranging from a small fraction to something of comparable significance (Ineson et al. 2015; Rozanov et al. 2012; Stozhkov et al. 2017).

There is also considerable contention in the scientific literature concerning the presence of a measurable acceleration in mean sea level, commensurate with the consensus of climate change science (Church et al. 2013). Fasullo et al. (2016) note that among the major unanswered questions associated with climate change is why global mean sea level (GMSL) acceleration has not yet been detected in the altimeter record, given the increasing rates at which glacial and ice sheet melt is estimated to have occurred and as greenhouse gas concentrations have risen.

Fasullo et al. (2016) postulate that by using a combined analysis of altimeter data and specially designed climate model simulations, the 1991 eruption of Mt. Pinatubo likely masked the acceleration that would have otherwise occurred. Model simulations infer that the eruption had a profound and temporally complex influence on several contributors to sea level and especially the amount of heat stored in the oceans.

Fasullo et al. (2016) concluded that barring another major volcanic eruption, a detectable acceleration is likely to emerge

from the noise of internal climate variability in the coming decade in the satellite altimetry data set.

Alternatively, Watson et al. (2015) proposed the identification of significant nonzero systematic drifts that are satellite specific, most notably affecting the first 6 years of the GMSL record. This theory has been recently validated by others (Dieng et al. 2017; Tollefson 2017). Having accounted for the identified systematic drifts, Watson et al. (2015) observed an acceleration in sea-level rise over the altimetry period which was larger than the twentieth-century acceleration using quadratic model fits to infer acceleration. However, an extensive analysis of the world's longest tide gauge records around the USA (Watson 2016) and Europe (Watson 2017), using more robust techniques to estimate time-varying acceleration, provides no statistical evidence of an acceleration over the altimetry period within the spatial margins studied or one that is necessarily higher than occurring elsewhere throughout the historical records.

Another area of scientific contention over the past decade remains our ability to understand and close the associated sea-level budget based on measured and modeled components to improve the veracity of climate projection model products. Church et al. (2013) advised that the closure of the observational sea-level budget since 1993, within uncertainties, represented a significant advance since AR4 (IPCC 2007) in physical understanding of the causes of past GMSL change and provided an improved basis for critical evaluation of models of these contributions in order to assess their reliability for making projections.

However, a NASA study by Zwally et al. (2015) concluded that Antarctica was overall increasing in ice mass, challenging the conclusions of numerous other studies, including AR5 (IPCC 2013), which have consistently concluded that Antarctica was losing mass. The profound conclusions from the Zwally et al. (2015) study cast doubt on the apparent closure of the sea-level budget from known and measured contributions at this point in time, though these conclusions remain open to contention against the backdrop of the contemporary literature (Scambos and Shuman 2016; Martín-Español et al. 2017).

These examples serve to remind us that the science concerning climate change and sea-level rise is remarkably complex and will continue to evolve well into the future.

Conclusions

Over the past millennium, sea level has fluctuated in response to two key processes. Firstly, the primary forcing mechanism relates to Earth's orbital variations which govern G-IG cycles. At present, mean sea level is close to the high point of this cycle, varying by less than ± 10 cm over the past three

millennia. The second forcing mechanism relates to global warming during the "Industrial Era" resulting from the accumulation of greenhouse gases in the atmosphere. A consensus of scientific opinion suggests that at some time between ≈ 1850 and 1945, the rate of sea-level rise attributable to global warming began to exceed the low background rate over the late Holocene. Forecast projections of global warming suggest sea level could rise by close to a meter over the twenty-first century under certain scenarios (Church et al. 2013).

Under these circumstances, the rate of global sea-level rise is projected to increase from ≈ 3 mm per year at present (Dieng et al. 2017) to rates ≈ 8 –16 mm per year by 2100 (Church et al. 2013). The likely trajectory of sea-level rise due to global warming this century is anticipated to become much clearer across multiple lines of scientific evidence over the next 20 years.

Cross-References

- [Archaeology and Sea-Level Change](#)
- [Changing Sea Levels](#)
- [Coastal Changes, Gradual](#)
- [Eustasy](#)
- [Eustasy and Sea Level](#)
- [Geodesy](#)
- [Geomorphology and Sea Level](#)
- [Global Positioning Systems \(GPS\)](#)
- [Holocene Epoch](#)
- [Isostasy](#)
- [Last Interglacial](#)
- [Late Quaternary Marine Transgression](#)
- [Sea-Level and Climate Change](#)
- [Sea-Level Indicators, Biologic](#)
- [Sea-Level Indicators: Biological in Depositional Sequences](#)
- [Sea-Level Indicators, Geomorphic](#)
- [Thermal Expansion](#)
- [Tide Gauges](#)

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Sea-Level Indicators, Biologic

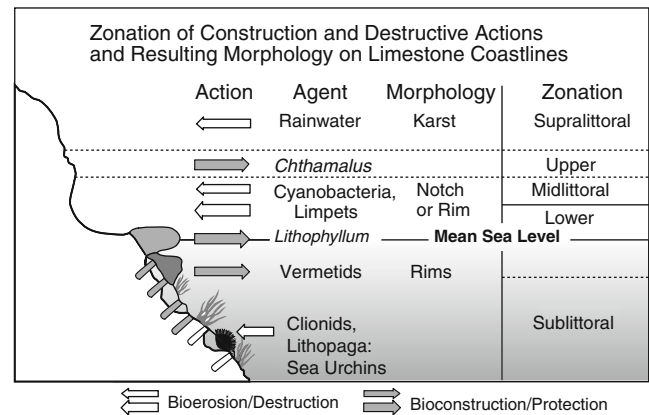
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Use of subfossil fixed biological remains as Biological Mean Sea-level Indicators (BMSIs), (Laborel and Laborel-Deguen 1994) or Fixed Biological Indicators (FBI) (Baker and Haworth 1999) were initiated about 40 years ago (Donner 1959; Van Andel and Laborel 1964; Fevret and Sanlaville 1966) and have gained recent impetus as the study of sealevel variations developed and took into account pluridisciplinary criteria. Use of FBI allowed reliable monitoring of recent sea-level variations along rocky coasts, stable or seismically active (Brazil, West Africa, Mediterranean, Japan, Australia).

Principle

On rocky shores, littoral fauna and vegetation currently develop in horizontal belts parallel to the water surface

Jacques Laborel: deceased.



Sea-Level Indicators, Biologic, Fig. 1 Division of erosion and construction on a vertical profile on limestones in temperate seas (modified from Laborel and Laborel-Deguen 1994)

(Stephenson and Stephenson 1949; Peres and Picard 1964) which define several zones, where various eroding and building biological factors are at work (Fig. 1). These zones are:

A *littoral fringe* (Stephenson and Stephenson 1949) or *supralittoral* zone (Peres and Picard 1964), wetted by surf where endolithic Cyanobacteria are dominating.

A *midlittoral* zone (Peres and Picard 1964), submersed at regular intervals by waves or tides, where parallel vegetational belts are more developed. Eroding Cyanobacteria, patellaceous gastropods (limpets), and chitons (lower zone) are abundant. Rock-building agents (coralline rhodophytes) also occur.

An *infralittoral* (*sublittoral*) zone, from mean sea level (MSL) down to 25–35 m whose upper part bears brown algae, coralline rhodophytes, vermetid gastropods, oysters, annelids, cirripeds, and eroding agents such as clionid sponges, sea-urchins and rock-boring pelecypods (*Lithophaga*). A few species are restricted to the upper margin of that zone, but many display a clear-cut population limit at that level. Some organisms build various reef or reef-like structures (bioherms, biostromes), or develop as an erosion-protecting cover.

On limestone coasts, the balance of bioerosion versus bioconstruction leads to various types of vertical profiles (Guilcher 1953) such as “tidal” notches and horizontal “benches” or “tidal platforms” on soft rocks.

Definition of a Biological Mean Sea Level (BMSL)

The limit between the midlittoral and infralittoral zones defines a (BMSL) marked by a large and sudden increase in species diversity (Boudouresque 1971). It also corresponds to such morphological features as the vertex of “tidal” notches or the inner edge of erosion “tidal” platforms (Focke 1978; Pirazzoli 1986).

Principal Groups of Plants and Animals Used as FBI

A small group of fixed plant and animal species such as the corallines *Lithophyllum lynoides* and *Lithophyllum onkodes*, vermetid gastropods of genus *Dendropoma* and *Spiroglyphus* (Laborel 1986) and annelids such as *Idanthyrus* and *Galeolaria* (Baker and Haworth 1999) enjoy very narrow depth ranges located at that limit (or a little above or below) so they are currently used as FBI and considered to be among the most reliable indicators. They are all the more interesting since they often develop into algal rims, cornices, and other reef-like bioherms whose erosion-resistant remains are easy to spot and sample. Most other species have a wider vertical range but may be successfully used as FBI taking into account the upper limit of their populations (for infralittoral cirrhipeds, like *Balanus*, largely used in the archaeological study of ancient harbors (Morhange et al. 1996), and *Lithophaga* holes). Scleractinian corals which develop in the infralittoral (sublittoral) zone belong to this category and the upper limit of coral construction is widely used as FBI in tropical waters (Hopley 1986). The use of coral indicators, generally sampled by coring or drilling methods, has generated a large specific bibliography of its own that we shall not develop here.

For midlittoral species such as the small barnacles *Chthamalus* or *Elminius*, or some coralline rhodophytes, it is the lower limit of the population which is taken into account.

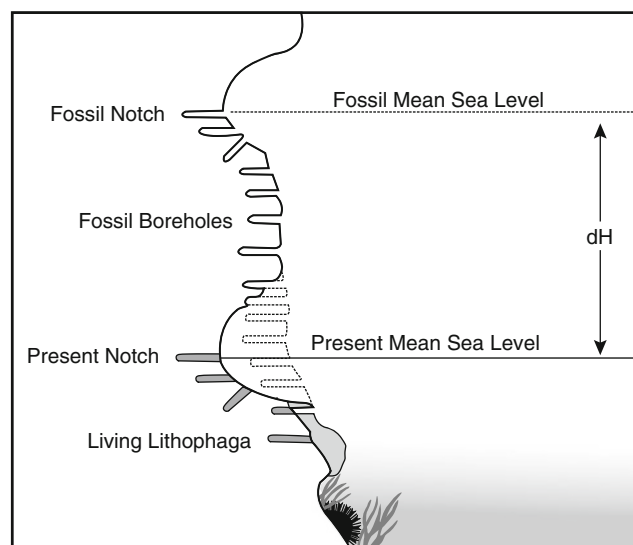
Population limits inside both midlittoral and sublittoral zones are considered as stable when MSL is constant. Aperiodic or seasonal short variations of sea level have little or no influence since most species have a long life and integrate sea-level variations on a yearly scale.

Field Use of FBI

The FBI must provide information upon the direction and rapidity of past displacements of MSL and allow easy and accurate radiometric dating.

The elevation or submersion of a displaced coastline is defined as the altitudinal difference between the upper limit of the displaced remains and that of their present homologs, measured with the local BMSL as datum (Fig. 2). No special reference (observed or calculated) to the actual water level is therefore necessary.

For species with a wide vertical range, best results are obtained when the uppermost limits of both fossil and living populations are well delineated. It must also be noted that biological sea-level marks are not perfect horizontal lines but may be warped, even on short distances, by local variations of hydrodynamism (Laborel and Laborel-Deguen 1994). Measurements must then be done on a single vertical profile



Sea-Level Indicators, Biologic, Fig. 2 Measurement for the elevation of an ancient sea level (modified from Laborel and Laborel-Deguen 1994)

including both the fossil specimen and its present equivalent. Selecting a unique temporary bench mark for several measurements (Jardine 1986), even at a horizontal distance of a few meters is therefore not recommended.

Discussion

An accuracy of ± 5 cm was obtained in Crete (Thommeret et al. 1981) for vermetid rims. In the western Mediterranean, a vertical accuracy of about ± 10 to 20 cm is common on submerged lines of *L. lichenoides* (Laborel et al. 1994). Lower accuracies (about ± 50 cm or less) were obtained for Brazilian vermetids (Delibrias and Laborel 1971) or for *Chthamalus* in surf-beaten crevices.

When sea level is falling slowly (about a few millimeters per year), the frailer sublittoral FBI species are killed, and their skeletons are eroded after a few years in the midlittoral zone. Reeflike structures, being stronger, may still be used as sea-level markers after long periods of midlittoral erosion. Preservation of frail details indicates rapid (even very rapid!) uplift (Thommeret et al. 1981).

Determination of the velocity of a submergence movement is sometimes difficult but submersed *Lithophyllum* or vermetid rims may be preserved underwater for long periods of time.

The limits of the FBI method have been tested by comparison of field observation and direct sea-level measurement in an area of high volcanic activity (Morhange et al. 1998). They proved to be reliable and sensitive in most cases, with the exception of short-lived oscillations of sea level (less than a few years) which are too rapid to be registered by biological growth.

Radiocarbon dating of aragonitic shells living in agitated surface sea water, is generally accurate. Calcareous algal bioherms can also be dated notwithstanding inner matrix and micritic cements. Direct isotopic counting methods make dating possible for small limestone fragments provided samples have suffered no contamination by alien carbon and the regional reservoir effect correction for sea water is known. Careful selection and cleaning of samples is always necessary.

Cross-References

- [Algal Rims](#)
- [Bioconstruction](#)
- [Bioerosion](#)
- [Bioherms and Biostromes](#)
- [Coral Reefs](#)
- [Littoral](#)
- [Notches](#)
- [Sea-Level Indicators: Biological in Depositional Sequences](#)
- [Sea-Level Indicators, Geomorphic](#)
- [Shore Platforms](#)

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Sea-Level Indicators, Geomorphic

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Introduction

Several geomorphic features, erosional or depositional, develop near sea level. Some of them may be preserved after a change in sea level and can be used, therefore, as indicators of former sea-level positions. Erosional indicators can be preserved only in hard rock, and occur in a vertical range which depends on site exposure. They include notches, benches, trottoirs, platforms, abrasional marine terraces, strandflats, pools, potholes, sea caves, honeycomb features, and tafoni. For accurate sea-level reconstructions (Pirazzoli 1996), it is essential therefore to refer the elevation of a former indicator to that of the active counterpart in the same place rather than to that of the present sea level. Depositional indicators include tidal flats, marine-built shore platforms and terraces, beaches, beachrocks, reef flats, and submerged speleothems. Erosional features are generally inadequate

Paolo A. Pirazzoli: deceased.

to date former sea levels, whereas marine deposits may include guide fossils or organic material liable to be dated radiometrically.

Erosional Indicators

Several types of notches can be distinguished (Van de Plassche 1986). Abrasion notches are often found near high-tide level at the boundary between a cliff and a shore platform, but they may develop at any level reached by wave action and give therefore only approximate indications of sea-level position. Much more precise are tidal notches, which are typical midlittoral erosional features, especially on limestone coasts. In sheltered conditions tidal notches occupy all the intertidal range, showing a vertex at about mean sea level (MSL) (Fig. 1).

The formation of erosional benches and shore platforms is usually ascribed to the removal by waves of the weathered parts of cliffs. The lowest level of possible weathering corresponds to that of constant soakage by sea water, probably in the intertidal zone. Wide benches of abrasional origin are called platforms or terraces; they usually develop in the intertidal range, sloping gently seawards.

Not to be confused with tidal notches are surf notches, which may occur at higher elevations on limestone coasts exposed to persistent winds and strong surf and spray action; surf notches delimit the boundary between a surf bench and a cliff. On exposed coasts, organic accretions often develop near the outer edge of a bench, protecting the substrate rock, while erosion can proceed higher than the accretion level, widening a surf bench and undercutting a surf notch on the cliff. The elevation of surf benches depends upon site exposure; elevation as high as 2 m above high tide-level has been reported locally, but it may decline rapidly with decreasing wave exposure. On the Levant coasts, in the eastern Mediterranean, surf benches (called trottoirs) are quite common at elevations of 0.2–0.4 m above MSL.

In high latitude, strandflats are low shore platforms where the rock material, disintegrated by freeze–thaw alternation, has been washed away by wave action. The reliability of erosional coastal benches, platforms and strandflats as sea-level indicators depends on the identification and understanding of the processes by which these features were produced.

Coastal pools are flat-bottomed depressions resulting from local lowering of limestone benches. On sheltered shores they correspond to intertidal levels, whereas on exposed coasts they may occur in any supralittoral area in the reach of waves. Potholes are rounded depressions, usually deeper



Sea-Level Indicators, Geomorphic, Fig. 1 Emerged tidal notches cut into a mushroom rock and, in the background, an emerged beachrock capped by green living algae, are consistent with the present-day

low-tide situation and indicate recent stability in the relative sea level. Miyako Island (the Ryukyus, Japan). Local mean spring tidal range is 1.5 m; water level is 0.4 m below MSL (photo 6149, Feb. 1981)

than wide, worn into solid rock by sand, gravel, pebbles, or boulders spun round by the force of waves. As they may be formed at various elevations, above or below sea level, they are not accurate sea-level indicators.

Sea caves are cavities excavated by erosion into a cliff in the range of wave action. They generally develop into weaker parts of the rock formation or are often of karstic origin. A sea cave open on both sides of a promontory is called a sea arch. Sea caves and arches are generally inaccurate sea-level indicators; however, in limestone formations, their floor, if regular and flat, may be related to a former low-tide position.

Honeycombs are small cavernous features, showing a cell-like structure, which may form on the surface of several granular rocks in the spray zone above high tide. Their development may be ascribed to temperature variations, chemical weathering, and wind corrosion. Honeycomb structures indicate the proximity of sea level below them, but with poor accuracy. Tafoni are cavernous weathering features of greater size (from a decimeter to several meters). Because they may occur also far away from the shore, tafoni should be avoided as sea-level indicators.

In uplifted areas of former ice sheets, low-tide positions corresponding to the marine limit (maximum recorded sea-level elevation) can be identified at the elevation where channels disappear from the surface of former glacio-marine deltas.

Depositional Indicators

The grain size of marine deposits depends mainly on the energy of the water environment, which varies with exposure to waves and currents. Marine muds and clays can be deposited only in very calm water, either offshore at depth, or in very sheltered coastal basins. Tidal flats are marshy or muddy land areas which are covered and uncovered by the rise and fall of the tide. They are usually studied by analyzing samples collected using boring techniques. The upper part of tidal flats, which may reach but not exceed the elevation of extreme high tides, is usually very flat and colonized by halophytic vegetation. Flora and fauna still in growing position, and the local tidal range and topographic configuration, will often help to indicate the kind of environment in which they formed and the depth of their deposition. Well-chosen foraminiferal assemblages from coastal paleomarine deposits may permit very precise (up to ± 10 cm) determinations of former sea level even in macrotidal areas. Botanical remains can also contribute to sea-level reconstructions, by determining the degree of salinity and therefore the position relative to mean high-water level. Assemblages of freshwater, brackish, and marine microflora (diatoms) can be used to infer marine transgressions and regressions (Van de Plassche 1986).

The main difficulties in reconstructing former sea-level histories from tidal-flat deposits are given by the estimation of sediment compaction and tidal changes. Wet sediments have a low density. After their deposition, compaction is rapid during the first centuries, then decreases gradually. In the Mississippi Delta, for example, contemporary silty muds contain water to 70% of their volume. At a burial depth of 24 m there is about 15% loss by compaction and at about 1,000 m apparent subsidence may approach 70%. The highest rates of compaction correspond to peat layers, which can reduce in volume as much as 90%. To survive, a tidal flat needs therefore rates of sedimentation high enough to compensate compaction phenomena. If sedimentation is inadequate, the tidal flat will be submerged; if it is excessive, its marginal, sheltered basin will be rapidly infilled. Tide characteristics depend narrowly upon the basin morphology, which is continually modified by sedimentation and compaction and eventually by sea-level changes. For example, in the Bay of Fundy, during the last 4,000 years, the high-tide level has increased 1.5 mm/year faster than MSL.

Sand deposits are common where wave energy is moderate and are found on beaches and intertidal shores near lagoon entrances. Pebbles are generally found only in high-energy shores, usually on exposed beaches. Raised or submerged beaches are however, seldom used as sea-level indicators because of the wide uncertainties in determining a clear relationship between a beach sample and the MSL. Depositional shore platforms produced by the supply of marine sediments (sand, shingle, or shells) to the shore have been described by Zenkovich (1967) and called improperly “shore terraces”; broad coastal plains characterized by many low shelly ridges that run roughly parallel to the shore, like the “cheniers” in coastal Louisiana (Shepard and Wanless 1971), also belong to the depositional shore platforms category. Their sea-level indications are similar to those provided by beaches.

Marine-built terraces develop near sea level and are made of accumulations of marine materials removed by shore erosion. Their landward part may be capped by alluvial deposits.

The term beachrock applies to beach sediments cemented by calcium carbonate in the intertidal zone (often in the upper part of it). Beachrocks are generally limited to areas of warm or temperate waters. Their upper surface usually shows a superimposition of beachrock slabs, each showing the same seaward slope as the nearby beach, delimited landward by baset edges. A beachrock is generally a good indicator of sea level, with a vertical uncertainty depending on the local tidal range. However, radiometric ages of beachrock samples (shells, coral debris) would provide only maximum dates for the sea level at which lithification occurred, because beach sediments may be reworked.

Reef flats are stony expanses of reef rock with a flat surface. They are generally situated in the lower part of the

tidal range and capped by calcareous algae, patches of sand and coral debris, and occasionally a few coral colonies living in shallow pools or depressions. In the absence of moating phenomena, the uppermost corals in growth position, often appearing as microatolls, correspond to the spring low-tide level. When the reef flat surface is made of coral conglomerate, its elevation may be at any level in the intertidal range or in the reach of regular waves. The low-tide level at the time of the conglomerate lithification can be determined by petrological analysis as the vertical boundary between the former marine phreatic and marine vadose environments for the first generation of intergranular cements (Montaggioni and Pirazzoli 1984). Speleothems, such as stalactites or stalagmites, are usually formed in caves above sea level by the evaporation of mineral-rich water. When submerged in sea-flooded karst systems, the depth of speleothems indicates minimum sea-level rise. In the fortunate case of a marine biogenic cover capping the speleothem, both the marine bioconstruction and the speleothem deposits can be dated radiometrically.

In glacio-isostatically uplifted areas, the highest marks of marine action may be identified from the lower limit of perched glacial boulders, till and continuous terrestrial deposition, or from the upper limit of shore deposits, beach ridges, and *in situ* marine fossils.

Lastly on the continental shelves, seismic stratigraphy gives a picture of the way the rock strata have been piled on one another. Each transgressive–regressive cycle removes some sediment and leaves piles of new sediment, producing recognizable patterns such as erosional unconformities. These features can be used to identify the best places to collect borings to interpret sea-level changes. In the laboratory, diagenetic products and intergranular cementation identified by analysis of depositional material may also provide evidence of former marine zonation related to sea-level positions.

Cross-References

- Beach Features
- Beachrock
- Cheniers
- Marine Terraces
- Muddy Coasts
- Notches
- Peat
- Salt Marsh
- Sequence Stratigraphy
- Shore Platforms
- Strandflat
- Tidal Flats
- Trottoirs

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Sea-Level Indicators: Biological in Depositional Sequences

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Introduction

Biological sea-level indicators have been discussed for *rocky intertidal coasts* but this section deals with biologic remains in a depositional sequence such as a marsh or *estuarine* (*q.v.*) deposit. Often the lithology will not provide sufficient information to delineate an accurate sea level from within a depositional sequence but biological remains in the form of fossils will greatly enhance the accuracy. Any good sea-level indicator must provide three essential elements: (1) accuracy, (2) preservability, and (3) be datable. Fossils within depositional sequences have the highest probability to provide all three of these elements. The type of fossil is very important because not all fossils will provide an accurate relocation of a former *sea level* (*q.v.*). The accuracy of a given fossil type depends mostly on the range of water depths that the organism occupies when living and it must be *in situ* when found in a sequence. However, even fossils with a large range may be useful to determine marine/freshwater transitions which may in turn sometimes be used as a sea-level indicator. There is a large range of organisms that serve as indicators; probably the most commonly used historically are “miscellaneous shells” which are not very accurate because they often are not in place and contain a large array of species with a broad band of water depth ranges. Some specific macro invertebrates such as mussels or oysters provide a range within ± 2 m but this is not adequate for determining the smaller movements of the late *Holocene* (*q.v.*). By far, the most useful are microfossils

because they occur in large numbers in small diameter cores which are often where the sea-level records come from in Holocene *submerged coastlines* (q.v.). In the following sections the various groups will be detailed in terms of where and how to use the various groups as well as a brief assessment of the accuracy possible with each group.

Plant Groups

Marsh Vegetation Generally

There are many cases in the literature where sea levels have been reported based on either “undifferentiated peat” (q.v.), meaning the authors did not know if it was freshwater or marine (Emery and Garrison 1967); in this case the only information that can be derived is that sea level was at about the depth (± 5 m) of the deposit at the time of formation or lower if it was freshwater. This determination is useful in the absence of any other data but not useful for the problems of measuring late Holocene movements which are well within the error bars mentioned above. Sometimes it is known that the peat is a salt marsh deposit in which case it narrows the range to the upper half of the *tidal cycle* (Chapman 1960, 1976). When one steps on a salt marsh it is visually obvious that there is a distinct plant zonation but the plants respond to more than just tidal exposure and often will have differing ranges even in adjacent marshes (Scott et al. 1981, 1988) so the macrovegetation alone is not a reliable indicator for subdividing salt marsh deposits in terms of sea level, especially in fossil peats.

On a broader scale macro-plant remains have been used to suggest trends of salinity change and hence *transgression regression* (q.v.) but there are many factors that have to be considered since plants respond to many variables (Behre 1986).

Many have also used buried forests as indicators of submergence (Heyworth 1986) and again this gives only the indication of the sea level being below the forest at the time it was living. This is a very powerful tool, however, when looking at rapid changes in sea level such as earthquakes (Atwater 1987) where the repeated submergence of a series of forests provides a sense of the periodicity of these events.

Microfossil Plant Remains

Apart from the variability of vegetation assemblages across the marsh, there is the problem of identifying macro-plant remains in a peat deposit since they tend to breakdown rapidly in the subsurface. However, some attempts have been made to use pollen which of course is the microscopic part of the sexual reproductive organs of all angiosperm plants. This meets with some success (Shennan et al. 1998) but pollen is inherently reworked. Probably, more successful is the use of diatoms which are one-celled algae which leave a siliceous

shell in the sediments (Shennan et al. 1998, 1999). These fossils appear to leave a record that can partially subdivide the marsh into zones with varying accuracies (± 20 –50 cm at best).

The most useful technique involving microfossil plants is determining sea levels in an indirect way where *coastal ponds* either become submerged or emerged (Palmer and Abbott 1986; Shennan et al. 1996). If they are emerged they go from marine to freshwater and *vice versa* for submergence; the key is to be able to determine the sill depth of the basin and then relate that to the radiocarbon-dated transition. This method is usually accurate to within ± 1 m depending on the depth and size of the basin (Laidler and Scott 1996). Either diatoms (Palmer and Abbott 1986) or dinoflagellates (Miller et al. 1982) can be used in this manner. Dinoflagellates are also microscopic algae but they are organic-walled as opposed to the siliceous shells of diatoms. Both of these groups occur in large numbers such that 1 cm³ of wet sediment is often sufficient to obtain a valid result (Haq and Boersma 1978).

Animal Groups

Macro-animal Fossils

As discussed briefly above invertebrate macrofossils are limited in their value as sea-level indicators by their vertical range in the water column. Some groups such as *corals* (q.v.) or attached biological indicators have extremely narrow ranges and have been used to produce some of the best and longest sea-level curves (Fairbanks 1989). However, these are limited in occurrence and rocky intertidal forms are often not well preserved, especially in a submergent regime. Other macro-invertebrate groups have been used in depositional sequences but their vertical ranges are usually quite high and hence accuracy is not good, especially the above-mentioned “miscellaneous shells.” Peterson (1986) details the use of marine molluscs which are by far the most widely used macro-fossil in Quaternary studies of raised marine deposits. He emphasizes the use of communities rather than single species to relocate sea level which is generally true for all organisms. Of all the species of molluscs perhaps *Mytilus* (the blue mussel) provides the highest accuracy (5 m) but even that range is high. However, in the raised deposits of northern Europe it is still very useful. However, as is the case with macro-vegetation, there is often insufficient material in small diameter cores that are often the basis for building sea-level curves, especially Late Holocene curves on submergent coastlines.

Micro-animal Fossils

There are two principal groups of animal microfossils that have been used extensively in sea-level studies, ostracodes

(Van Harten 1986) and foraminifera (Scott and Medioli 1986). Ostracodes leave a calcareous shell as a fossil and that in itself presents a problem because many of the best deposits for sea-level studies are not conducive to the preservation of CaCO_3 hence the fossils are not present. van Harten (1986) suggests the resolution with ostracodes to be within 100 m which is not useful at all for modern sea-level studies but they can be used as accurate salinity tracers and hence suggest transgression and regression (Haq and Boersma 1978).

On the other hand, foraminifera have both calcareous and agglutinated shells, agglutinated shells are resistant to dissolution in low pH, highly organic sediments, and often are very abundant in some highly organic deposits such as marshes ($>5,000/10 \text{ cm}^3$, Scott and Medioli 1980). Unlike plants foraminifera have been shown to be consistent in their vertical range in relation to tidal levels within a marsh sequence (Scott and Medioli 1978, 1980, 1986) such that the same 8–10 species inhabit the world's marshes and specific assemblages can always locate the upper one-fourth of the tidal range. In some cases the higher high level can be accurately located to provide an incredible accuracy of $\pm 5 \text{ cm}$ (Scott and Medioli 1978, 1980, 1986; Hayward et al. 1999); this is by far the most accurate indicator now available but it is limited to coastal marsh areas and isolated peat deposits that are sometimes found offshore in marine surveys (e.g., Scott and Medioli 1982). At the very least “undifferentiated peats” can now at least be determined to be either marine or freshwater which significantly increases their value as a sea-level indicator.

Archaeological Remains

Although many assumptions must be made to use paleo-human occupation sites as sea-level indicators they have been used by many workers, especially on emergent coasts (Colquhoun and Brookes 1986; Martin et al. 1986). The main assumption is that paleo-humans did not carry their food that they gathered from coastal *estuaries* and *lagoons* (*q.v.*) far from the coastline. The most commonly used remains are shell middens or dumps of usually molluscs that are exposed in many areas but are probably the most spectacular on the South American east coasts of Brazil and Argentina (Martin et al. 1986). Colquhoun and Brookes (1986) used a combination of shell middens and peats to reconstruct sea levels on the South Carolina (USA) coast.

Summary

The above is a small sampling of the most commonly used biological remains in depositional sequences for determining sea levels. It is clear that some areas will provide a higher probability of determining an accurate and extended sea-level

record than others and in some areas it may be impossible to obtain a sea-level record of any kind. Biological remains provide the best means of determining former sea levels because they are usually in a depositional sequence and they usually can supply the carbon required to obtain a ^{14}C date, without which you have a level but no way of knowing when the sea level was at that point. The key to determining accurate sea levels is being versatile in the approach taken to take advantage of what the record provides.

Cross-References

- Bogs
- Coral Reefs
- Estuaries
- Holocene Epoch
- Ingression, Regression, and Transgression
- Peat
- Rock Coast Processes
- Sea-Level Indicators: Biological in Depositional Sequences
- Submerged Coasts
- Tides

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Sediment Budget

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Definition

As it pertains to coastal sedimentary systems, the sediment budget can be defined as the balance between changes in the

volume of sediment stored in the system and the sum of the volumes of sediment entering or leaving the system. Examples of coastal sedimentary systems include estuarine areas composed of fine (mud-size) sediments and open-coast littoral systems most typically composed of sand-sized sediments. The focus here is on sedimentary systems of the littoral zone, the region of coast for which sediment transport is dominated by incident wave processes. This zone ranges from the intermittently dry beach (where wave swash dominates) to water depths of roughly 10–20 m (where the seabed first feels the impact of waves). A focus on the littoral zone reflects the objective of most sediment budget studies: to understand and/or predict longterm changes in the position of the coastline, the interface between the subaerial and subaqueous portions of the littoral zone.

Bowen and Inman (1966) provided one of the earliest sediment budget definitions for sandy coasts: “The procedure, sometimes referred to as the budget of sediments, consists of assessing the sedimentary contributions (credits) and losses (debits) and equating these to the net gain or loss (balance of sediments) in a given sedimentary compartment.” In essence, the sediment budget is a sediment volume continuity equation with three terms that sum to zero: sediment flux in, sediment flux out, and volume rate of change within the system. Solving the sediment budget is, in principle, simply a matter of determining two of the terms and solving for the third. Komar (1996) provides a recent review of modern sediment budget techniques.

Spatial and Temporal Scales of the Sediment Budget

Although there are no absolute limitations on the spatial and temporal scales considered for sediment budget investigations, in practice spatial scales are on the order of tens of kilometers of coast or greater and temporal scales are on the order of tens of years or longer. In general, these are the scales relevant to the long-term management of coastal erosion. The sediment budget is thus concerned with the large-scale and longterm net result of all processes that transport sediment in the littoral zone and affect the position of the coastline. Individual sediment transporting events, such as storms, may not be relevant unless they contribute to long-term change in coastline position.

The Littoral Cell

Fundamental to the determination of the sediment budget is the identification of the sedimentary system, and in particular the boundaries of the system. In terms of the continuity equation, this is analogous to the definition of a control

volume; for sediment budget studies the control volume is usually referred to as a littoral cell. No universally accepted standards exist for defining the littoral cell for all coasts – the coastal characteristics, the study objectives, and the method of sediment budget solution are all important factors in making this determination. One pragmatic, but broadly applicable, definition simply refers to littoral cells as “semi-contained entities where one can better develop a budget of sediments” (Komar 1996, p. 18).

The placement of littoral cell boundaries is typically guided by the focus of most sediment budget studies: to understand and/or predict coastline change. In principle, the littoral cell encompasses the region of littoral sediments for which sediment volume changes are closely linked to coastline changes, while regions external to the littoral cell act as sediment source or sink areas for littoral cell gains or losses, respectively. In practice, it can be very difficult to objectively place littoral cell boundaries, with problems unique to defining the longshore, landward, and seaward boundaries.

Ideally, the longshore boundaries of the littoral cell are defined to minimize sediment exchange with other littoral cells along the coast. A classic example is a pocket beach bounded by rocky headlands that are presumed to act as barriers to littoral transport (Fig. 1). Along many coasts, the longshore boundaries of the littoral cell are far less clear, especially on long stretches of sandy coast interrupted only

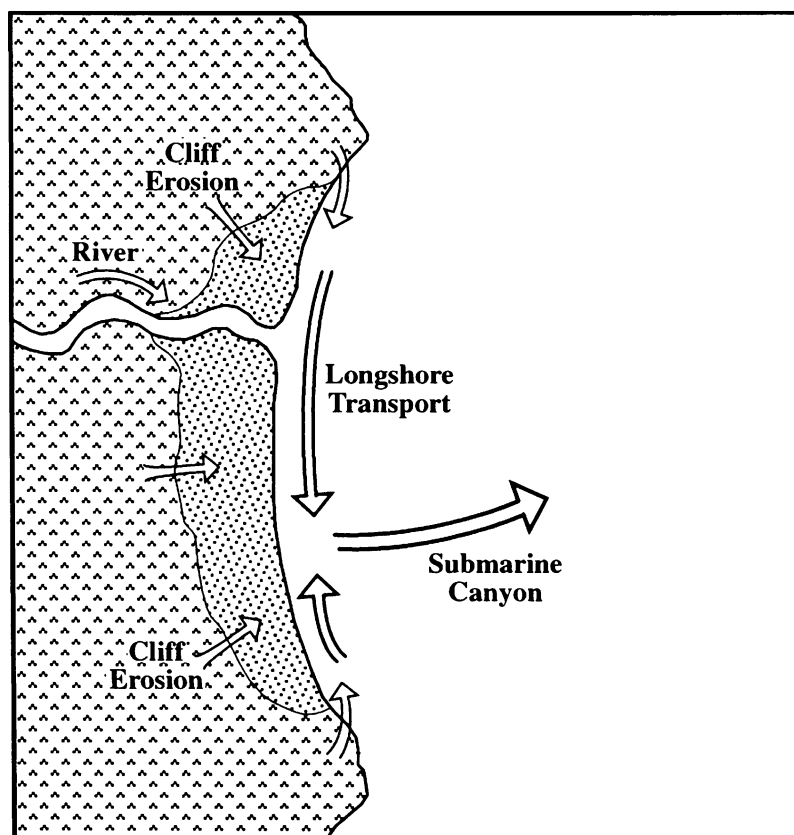
by tidal inlets. If the littoral cell is chosen to encompass a large enough area, an assumption of no sediment exchange with adjacent littoral cells may be realistic. However, if a littoral cell defined in this way includes sections of coastline with widely differing rates of erosion or accretion, the value of the sediment budget as a tool for understanding and/or predicting coastline change will be limited. In practice, these larger cells are usually broken into smaller subcells to isolate sections of coast with quasi-uniform rates of change (e.g., Jarrett 1991), with the underlying assumption that the processes acting within these smaller sections of coast are also quasi-uniform. Cell boundaries may thus lie within an uninterrupted section of coast, with the solution of the sediment budget requiring a simultaneous solution for all subcells with shared boundaries.

The landward boundary of the littoral cell is typically chosen as the base of an upland feature such as the toe of a dune or cliff, thus isolating littoral zone sediments from upland reservoirs of sediment that may serve as sources or sinks relative to the littoral cell. Although this is generally one of the most straightforward cell boundaries to define, significant erosion or accretion of the coastline over the time period of interest may require a moving cell boundary, with the associated complication that sediments may be repeatedly recycled between the littoral cell and upland features such as dunes.

Sediment Budget, Fig. 1

Schematic of a simple littoral cell consisting of a sandy pocket beach bounded by rocky headlands.

Arrows represent pathways for sediment transport (after Komar 1996)



The seaward boundary of the littoral cell can be especially difficult to define, in part because the processes of sediment transport in this environment are poorly understood. Typically, the seaward limit of the littoral cell is chosen on the basis of several simplifying assumptions. First, it is assumed that the littoral cell encompasses a coastal profile that maintains a constant form, generally known as an equilibrium profile (Dean 1991), as the coast advances or retreats. With this assumption, littoral cell volume changes can be equated to coastline changes through a simple geometric relationship (e.g., Jarrett 1991). It is further assumed that there is a seaward limit of significant sediment transport (Hallermeier 1981), at a depth roughly corresponding to the seaward limit of the equilibrium profile. This depth, typically judged to be 8–15 m on the ocean coast, is often referred to as the “closure depth,” a term derived from the depth at which sequential coastal profiles seem to “close,” or exhibit no significant change over time. Although some profiles have been shown to maintain a reasonably constant form over many decades and simple equations exist for determining the closure depth as a function of the wave climate and sediment grain size (Hallermeier 1981), profile data are typically inadequate for testing these assumptions rigorously in most study areas. Nevertheless, these simplifications provide an operational method for defining the seaward boundary of the littoral cell in the absence of more rigorous alternatives.

Quantifying Sediment Gains and Losses Within Littoral Cells

Regardless of the way in which the littoral cell is defined, a key challenge for all sediment budget studies is the determination of the sediment flux across cell boundaries. At each of the cell boundaries – longshore, landward, and seaward – there are many physical processes with the potential to add or remove sediment from the littoral cell. Although

few of these processes are understood well enough to permit the direct calculation of the cross-boundary sediment flux integrated over the sediment budget timescale, a critical first step for all sediment budget studies is to identify the important sediment transport processes in the region examined. In most cases it is also useful to identify the sources and sinks of sediment associated with these processes, that is, the areas outside the littoral cell that either supply sediment to the littoral cell or store the sediment leaving the littoral cell. Quantifying sediment volume changes within these source or sink areas is often the most reliable method of quantifying gains or losses to the littoral cell. Table 1 summarizes the principal natural and anthropogenic processes responsible for sediment flux across littoral cell boundaries, and the typical sources and sinks of sediment associated with these processes.

Longshore Transport

Waves breaking at an angle to the coast result in the longshore transport of sediment, also known as the littoral drift (see Komar 1990 for review). This transport can occur both on the subaerial beach under the influence of wave swash, as well as in deeper water where steady longshore flows are generated by breaking waves. Perhaps one of the most important components of the sediment budget, the net longshore transport is often readily apparent at coastal structures (e.g., jetties, groins) where sediment may be impounded (Fig. 2b).

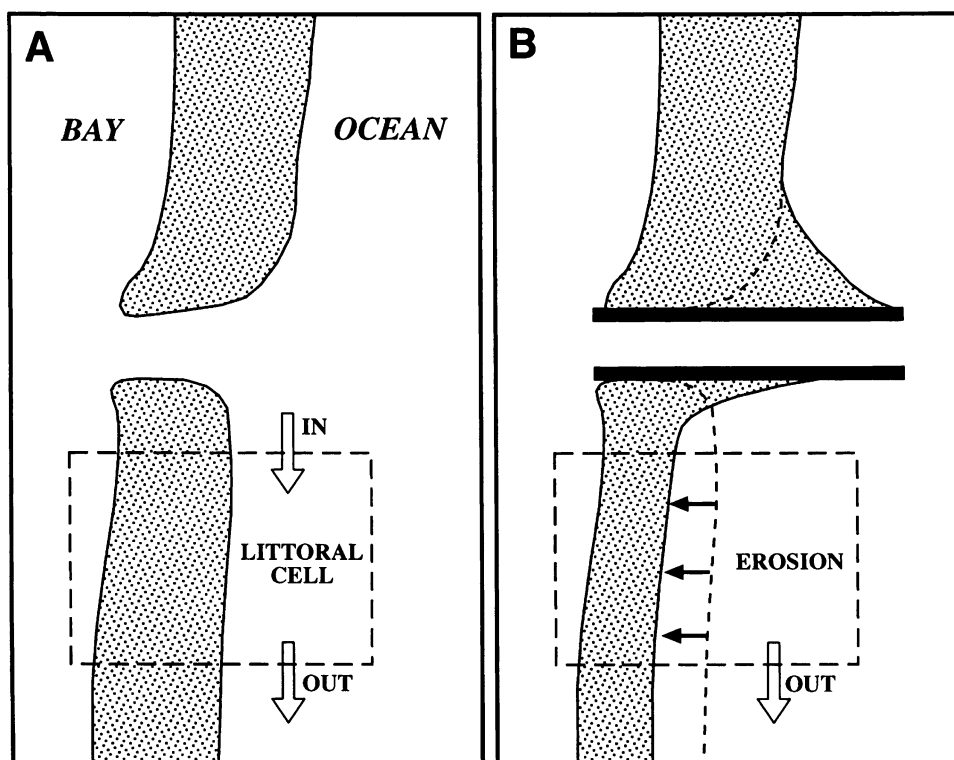
Two main approaches have been employed for estimating the littoral cell gains or losses due to longshore transport. A modeling approach employs empirically derived equations that relate the incident wave characteristics, in particular the wave height and direction, to the rate and direction of longshore transport (Komar 1990). To provide estimates relevant to the sediment budget, the longshore transport must be integrated over the timescale of the sediment budget. Both the gross sediment transport (total, irrespective of direction along the coast) and the net transport may be of importance to the

Sediment Budget, Table 1 Principal processes of sediment transport and associated sources and sinks of sediment relevant to the sediment budget

Sediment transport processes resulting in littoral cell gains or losses	Sediment sources (for littoral cell gains)	Sediment sinks (for littoral cell losses)
Longshore transport	Adjacent littoral cells; inlet sand bodies	Adjacent littoral cells; inlet sand bodies
Cross-shore transport	Offshore areas; cliffs, dune scarps; inlet sand bodies	Offshore areas; washover fans; inlet sand bodies
Riverine transport	Upland areas	
Aeolian transport	Coastal dunes; washover fans	Coastal dunes
Gravity flows	Cliffs, scarped dunes	Submarine canyons
Biogenic processes	Calcareous sand of biogenic origin	
Human interventions	Offshore and upland areas used as a source for beach nourishment; updrift sand delivered by inlet bypassing	Dredge disposal sites; industrial uses of sand from beach mining; sand bodies trapped at structures including groins, jetties, and dams; reduction in source from naturally eroding upland features (e.g., cliffs and dunes); artificial dunes constructed by beach scraping

Sediment Budget,

Fig. 2 Application of the sediment budget on a barrier island coast. In (a) the rate of sediment transport out of the littoral cell is balanced by the rate of sediment bypassing of the inlet and the coastline is stable. In (b) the construction of inlet jetties has eliminated the transport into the littoral cell, causing a net loss to the littoral cell and coastal erosion



sediment budget depending on the nature of the littoral cell boundary. For example, in Fig. 1, the gross longshore transport near the location of the submarine canyon is important if it is assumed that sand transported from either direction is trapped by the canyon. Alternatively, in Fig. 2 only the net transport at the longshore boundaries of the littoral cell may be relevant to volume changes within the cell.

Unfortunately, these transport estimates are generally highly uncertain due to uncertainties in the equations themselves or simply because quality wave information is lacking. Frequently, this approach is used only to estimate the difference in the longshore transport between the longshore boundaries of the littoral cell, with the absolute magnitudes of transport adjusted so that the observed volume changes within the cell are explained (e.g., Jarrett 1991).

As an alternative approach, littoral cell gains or losses due to longshore transport can be estimated, in some cases, by measuring changes in the volume of sediment contained in source or sink areas external to the littoral cell. For example, the growth of a terminal spit on a barrier island or the impounding of sediment updrift of a recently constructed barrier to longshore transport (e.g., Fig. 2b) may provide quantitative information on the volume of sediment transported, or prevented from being transported, across a littoral cell boundary.

Cross-Shore Transport

Waves and current also transport sediment in the cross-shore direction, with the potential for net transport across the

landward and seaward boundaries of the littoral cell. Although cross-shore transport can significantly influence the sediment budget, quantification of the gains and losses due to this process can be extremely difficult. Of the many important processes of cross-shore transport, few, if any, are understood to the degree that modeling or empirical approaches can provide quantitative estimates for the sediment budget. A further complication is that, for many of the relevant processes, the net transport across the littoral cell boundaries may be an extremely small fraction of the gross transport, yet this small net transport may still represent a major component of the sediment budget when integrated over the relevant timescale.

At the landward boundary of the littoral cell, wave swash – the up and down-wash of waves breaking on the subaerial beach – represents one of the dominant cross-shore transporting processes. Wave swash can transport large quantities of sand onshore or offshore depending on many factors, including the sand texture and the continually changing characteristics of incident waves. Because much of this transport is cyclic (with sediment typically carried offshore during storms and onshore during fair weather) and does not result in net transport across littoral cell boundaries, the sediment budget contribution from wave swash may be minimal under most conditions. With increasing storm intensity, however, wave and water levels may become high enough to intersect upland features such as dunes and cliffs, resulting in erosion of these features (e.g., Ruggiero et al. 1997). With the aid of gravity-induced slumping or debris

flows, this results in a direct transfer of sediment from the upland feature to the adjacent littoral cell, a process that is not readily reversible (with the exception of aeolian transport, see below).

Quantifying the sediment budget contribution from the erosion of upland features is usually accomplished through a source and sink approach. With data on the topographic change of an upland feature, the volume of sediment delivered to the littoral cell can be estimated (e.g., Komar 1983). However, as noted by Komar (1996) the volume of sediment delivered does not necessarily equal the volume of sand retained by the littoral cell if a component of the eroding sediment has a grain size incompatible with the littoral cell sediments (e.g., a silt or clay fraction will not remain in the littoral cell). In this case, only a fraction of the eroding upland feature volume can be assumed to contribute to the sediment budget.

When storm water levels and waves are high enough to overtop coastal dunes on barrier islands, offshore transport by wave swash may be replaced by onshore transport driven by both waves and steady currents associated with storm surge. The most common expression of this transport is a washover deposit or fan, consisting of sand removed from the littoral cell and deposited either on land or within the bay landward of a barrier island (e.g., Dolan and Hayden 1981). As the storm intensity increases, the waves and currents associated with washover fans may become focused at a particular point along the coast, opening a new tidal inlet through the barrier island. Although new inlets are typically reclosed rapidly by natural processes or human intervention, large quantities of sand may be transported onshore through these temporary openings, forming deposits in the back-barrier bay similar to washover fans. Washover fans and deposits associated with new inlets represent a loss of sand from the littoral cell, which can be quantified by measuring the thickness and areal extent of such deposits (e.g., Kochel and Dolan 1986). Estimating the long-term contribution from all storms over the timescale relevant to the sediment budget can be difficult, as washover and inlet deposits from past storms may be difficult to identify, both in terms of their volumes and their ages.

New inlets that remain open will subsequently develop a characteristic morphology consisting of ebb- and flood-tidal deltas on the seaward and landward side of the inlet channel, respectively. The processes of sediment redistribution associated with changes at tidal inlets are complex, involving both longshore and cross-shore transport, and may have a significant impact on the sediment budget of adjacent littoral cells (Rosati and Kraus 1999). Quantifying the sediment budget terms associated with tidal inlets can be exceedingly difficult. Perhaps the most straightforward approach is the use of bathymetric change to quantify the volume of sand trapped (or released) by the inlet over the time period relevant to the sediment budget being formulated. Unfortunately, adequate

bathymetric data are seldom available, and are of limited use for predicting the impact of major inlet changes (e.g., inlet opening, inlet closing, construction of inlet jetties, etc.). Frequently, assumptions are made of the efficiency of an inlet at trapping the gross longshore transport in order to arrive at an estimate of the inlet's sediment budget impact (Rosati and Kraus 1999).

At the seaward boundary of the littoral cell, determining the sediment gains or losses due to cross-shore transport is generally extremely difficult. At present, sediment transport knowledge is inadequate for quantitative predictions, although studies suggest, at least qualitatively, that offshore areas can act as both sources (e.g., Williams and Meisburger 1987) and sinks (e.g., Niedoroda et al. 1985) of sediment relative to the littoral cell. Because quantification is difficult, it is often assumed that sediment transport is negligible seaward of the predicted closure depth, making the determination of sediment budget gains or losses at this littoral cell boundary unnecessary (in principle). As an alternate approach, some studies have assumed that major imbalances in the sediment budget – after accounting for all the sediment budget components that can be quantified – are attributable to sediment flux at the seaward boundary (e.g., Inman and Dolan 1989). In almost all cases, however, this important component of the sediment budget remains a source of great uncertainty and warrants much further research.

Riverine Sediment Transport

The delivery of sand to the coast by rivers can represent a major component of the sediment budget. This is especially true on steep gradient coasts, such as typically found on the leading edge of continental plates, where rivers deliver their transport load directly to the coast. In contrast, the river valleys of low gradient, trailing edge coasts may be drowned, trapping sediment from rivers in estuaries before it can reach the coast.

A variety of techniques have been developed to estimate the sediment budget contribution from rivers, although all are associated with a high degree of uncertainty due to the difficulty of directly measuring the sediment load of a river integrated over the timescale relevant to the sediment budget. Komar (1996) reviews many of the methods that have been used, and makes the further point that, as in the case of sediment eroded from cliffs, only the fraction of delivered sand with a similar size as found in the littoral cell will contribute to the sediment budget.

Aeolian Sediment Transport

In many coastal areas, wind plays an important role in the flux of sediment across the landward boundary of the littoral cell. Upland features that are not well stabilized by vegetation, such as washover fans, may provide a source of sediment in areas dominated by offshore directed wind. In many areas,

however, aeolian transport results in the formation of coastal dunes, which by their existence indicate that the net effect of wind transport is landward – a loss to the littoral cell. As for most sediment transport processes affecting the sediment budget, the direct calculation of aeolian transport's contribution to the sediment budget is not feasible, and a source and sink approach, whereby changes in dune volume are inferred to represent a littoral cell loss, may be the best available method (e.g., Inman and Dolan 1989). Because coastal dunes are typically vulnerable to erosion by wave processes (see above), any sediment budget for which coastal dunes are important must take into account the possibility that dunes can represent a net source to the littoral cell, despite their earlier formation as a sink.

Gravity Flows

At both the landward and seaward boundaries of the littoral cell, sediment transport driven primarily by the direct influence of gravity may represent another significant contribution to the sediment budget. At the landward boundary, slumping and other forms of mass wasting of cliffs or scarped dunes can deliver sediment almost instantaneously to the littoral cell. The processes initiating this transport are varied and complex, but in many cases the cross-shore transport of sediment by waves plays a key role (e.g., Everts 1991). As described above, the contribution to the sediment budget can be estimated by quantifying the volumetric loss to the upland feature and accounting for the percent of eroded material expected to be size compatible with littoral cell sediments (e.g., Komar 1983).

At the seaward boundary of the littoral cell, submarine gravity flows can cause significant losses to the sediment budget, especially along narrowshelf coasts where submarine canyons extend landward into the littoral cell (e.g., Lewis and Barnes 1999). These gravity flows are initiated when longshore transport delivers sand to the head of the canyon; periodically the deposit becomes over steepened to the point where it slumps and flows down the canyon as a turbidity current. The most straightforward means of quantifying the volume of sand lost from the littoral cell is by determining the volume change in the sediment sink area at the bottom of the canyon. Often, however, the necessary bathymetric change information is lacking and inferences about the volume of sediment lost through this process must be made by considering the balance of other sediment budget components.

Biogenic Processes

In tropical and subtropical regions, a significant component of beach sand may be of biogenic origin, composed of calcium carbonate grains, rather than derived from terrestrial erosion (largely quartz or other silicate minerals). In these areas, the sediment budget may be strongly influenced by the biogenic production of carbonate material, such as corals, shells, and foraminiferal tests, which is reworked by physical processes

into sand of the littoral cell. As described by Komar (1996), estimating the volumetric contribution to the sediment budget is difficult, because the mere presence of carbonate sand does not give the rate at which it is being produced and lost due to abrasion. Although some techniques have been developed (see Komar 1996, for a description), much further research is needed to develop a reliable means of estimating this sediment budget component.

Human Interventions

Along many coasts, the influence of human activities is becoming increasingly significant and must be accounted for in balancing the sediment budget. Human activities take many forms, but can be broadly categorized into activities that interfere with natural processes, and activities that move sediment directly. Interferences in natural processes typically result in the reduction or elimination of a sediment source, representing a loss to the littoral cell (tabulated as the creation of new sediment sinks in Table 1). The construction of dams can trap sediment in upland reservoirs, greatly reducing the riverine transport of sediment to the littoral cell. Armoring of eroding upland features, such as cliffs and dunes, prevents transporting processes such as gravity flows and crossshore transport from delivering sediment that otherwise would have nourished the adjacent littoral cell. Groins and inlet jetties are designed to interrupt longshore sediment transport, creating new updrift sediment sinks for longshore transport (Fig. 2). Estimating the impact on the sediment budget as a result of these engineering structures is, in many cases, more straightforward than for purely natural processes, as the volume of sediment impounded (or prevented from being eroded) can often be directly measured.

Human activities that move sediment directly may also result in sediment losses or gains to the littoral cell. Sediment gains include beach nourishment, whereby sand is placed within the littoral cell using a source area external to the littoral cell; and inlet bypassing, whereby sand on the updrift side of a tidal inlet is pumped to the downdrift side (representing a loss to the updrift littoral cell). Sediment losses include inlet channel dredging when the disposal site is outside the littoral cell, the mining of beach sand for industrial uses (becoming rare), and beach scraping following storms, whereby sand is bulldozed off the beach foreshore to form artificial dunes. The best available estimate of the sediment budget contribution of these engineering activities is, in most cases, the contractor's estimate of sediment volume delivered.

Effect of Sea-Level Rise

Because sediment budget timescales are generally long, on the order of decades or longer, a rise in sea level relative to

land has the potential to significantly influence coastline position. Though neglected in most sediment budget studies, sea-level rise occasionally has been treated as an agent of coastal change operating independently of other processes affecting the sediment budget (e.g., Inman and Dolan 1989). The most common method of relating sea-level rise to coastline change is through a simple geometry-based model known as the Bruun Rule (Bruun 1962), in which a rise in sea level results in an upward and landward translation of the equilibrium profile with a redistribution of sediment from onshore to offshore. If the redistribution of sediment is assumed to occur within the bounds of the littoral cell, the amount of coastline change predicted by the Bruun Rule can be removed, in principle, from consideration when balancing the sediment budget (Inman and Dolan 1989).

Unfortunately, the approach of attributing a distinct component of coastline change to sea-level rise is not well founded by either field observations of the long-term profile response or through a processes-based modeling approach. In most cases, the uncertainty level in the overall sediment budget is enough to account for the component of coastline change that may be attributable to sea-level rise. Also, there is little justification for considering sea-level rise as an independent process, as sea-level rise by itself has no capacity to transport sediment. Rather, sea-level rise undoubtedly acts as a modifier of many transport processes influencing the sediment budget, although our understanding of these processes is currently too limited to quantify the effect of sea-level rise on coastline position. With the potential to be a significant and pervasive agent of long-term coastal change, sea-level rise clearly warrants much further research.

The Balance: Coastline Change

In principle, the sediment budget can be balanced by summing the volumetric contributions of all relevant sediment transporting processes, giving the net volumetric change within the littoral cell. Coastline change can then be predicted from volume change by invoking the equilibrium profile assumption, whereby the profile maintains a constant form out to the closure depth while translating landward (erosion) or seaward (accretion). In many cases, however, great uncertainties remain in most sediment budget components, as described above, and only the end result – littoral cell volume change or the coastline change itself – is known with any degree of certainty. For this reason, the observed volume or coastline change is frequently used to adjust or calibrate the most uncertain terms in the sediment budget. The sediment budget is then used to make predictions of coastline change given expected changes in specific sediment budget components, most often those induced by human activities. Several simple examples are given below.

In the case of Fig. 1, the most uncertain sediment budget component might be the down-canyon loss of sediment through gravity flows. By measuring the volume change within the littoral cell directly through bathymetric comparisons, or by estimating volume change from coastline change using the equilibrium profile assumption, the down-canyon loss can be estimated if the other important components of the sediment budget – here river input and cliff erosion – are reasonably well known. The utility of the sediment budget is then in making predictions of coastline change given a modification of a sediment budget component. For example, if cliff erosion is arrested by armoring a modification to the rate of coastline change can be predicted if it is assumed that the other sediment budget components remain unchanged after cliff armoring.

In another example, the construction of inlet jetties might interrupt the sediment bypassing a tidal inlet as in Fig. 2. If a preconstruction sediment budget has been established, including estimates of longshore transport and any other significant components, then the impact of jetty construction on coastline change can be estimated by assuming a total or partial interruption in the longshore transport reaching the littoral cell, and again assuming that other components of the sediment budget remain unchanged.

Despite their uncertain nature, the sediment budget often represents the best available tool for gauging the impact of human interventions, as well as changes due to natural processes, on the rate of coastline change. Also, as stated by Komar (1996, p. 25), “the formulation of a sediment budget can also serve to organize what is known about a coastal area or littoral cell, and where the chief gaps still exist in our understanding, thereby providing a guide for future research.”

Cross-References

- ▶ [Barrier](#)
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- Gross Transport
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- Sandy Coasts
- Sea-Level and Climate Change
- Shoreline Response to Littoral Drift Barriers
- Shore Protection Structures
- Storm Surge
- Wave-Dominated Coasts
- Waves

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Sediment Suspension by Waves

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Definition

Sediment suspension by waves is defined as sediment that is picked up (entrained) from the seabed and kept entrained by the motion of water. The water motion is imparted to the sediments on the seabed as fluid drag transmitted by the passage of waves.

Sediment suspension by waves occurs through the entrainment of loosely consolidated sediments, such as sand, by fluid drag (stress) imparted by water set in motion by the passage of waves. Once entrained, the moving water carries the sediments up into the water column away from the seabed. When the water stops moving, the sediments eventually settle back down to the seabed under the influence of gravity. However, moving water can counteract the effects of gravity by exerting fluid stress on the sediments and effectively dragging them up into the water column against gravity's pull. On the coast, waves arriving from deep water provide the necessary movement of water to lift sediments up off the seabed and transport them away from their initial location.

Sediment Transport Processes

When waves interact with the seabed, they mobilize sediment and progressively change the shape of the coastline over time. The mobilization of sediment by waves may be thought of as taking place in two principal ways (Bagnold 1963): (1) as a result of collisions between individual sediment grains and (2) as a result of fluid stresses on individual sediment grains, both of which may maintain the motion of the sediment. The first process is typically termed bedload and commonly occurs with large sediments such as sands, gravels, and pebbles. These large-sized sediments tend to bounce off one another because the water would have to be moving quickly for sustained periods of time to keep them suspended. The second process is typically termed suspended load and commonly occurs with sand and smaller-sized sediments. These sediments are small enough for the moving water to drag them into suspension. Sediments smaller than sand-sized material, like silt and clay, may be suspended for long periods of time and carried by the water over long distances before settling back to the seabed. These sediments form a subset of suspended material referred to as wash load because they are suspended for long enough periods that they are washed away. Waves reaching coastlines are typically energetic

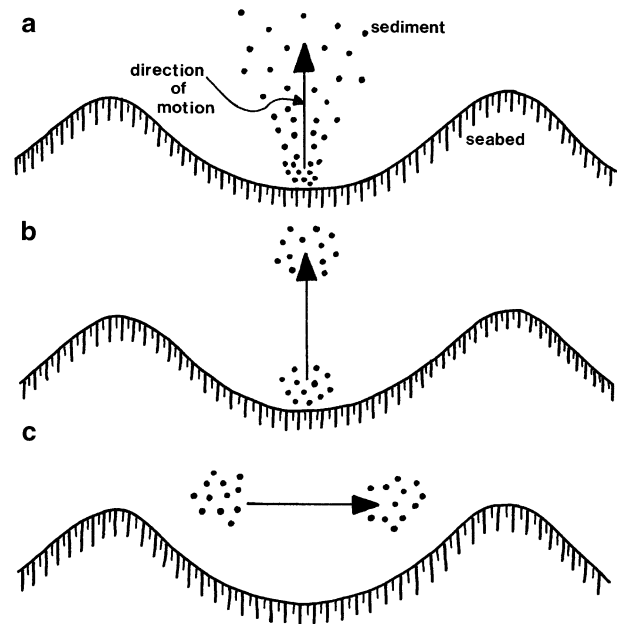
enough to suspend sediments smaller than sand size and wash them away into deeper water.

Historically, bedload was considered to be the principal process in sediment transport and to account for the volumetric majority of sediment moved since it included the movement of large sediments. Advances in understanding through the latter part of the twentieth century (e.g., Neilsen et al. 1979; Sternberg et al. 1989) indicated that suspended load is the major contributor to sediment transport. Sediment carried in suspension can account for upward of 80–90% of the volume of wave-transported sediment. So, even though bedload moves large sediments, the accumulated volume of smaller sediments moving in suspension comprises a significantly greater volume of sediment moved than bedload. Thus, the suspension of sediments under waves is an important factor in the redistribution of sediment and coastal evolution.

Sediment Suspension Mechanisms

In order to better understand the suspension of sediments under waves, several mechanisms have been proposed to account for the observed characteristics of suspension. Initially, diffusion, analogous to the mixing of concentrations at the molecular level in chemistry, was proposed as the mechanism for the suspension of sediment by Rouse (1937). By this mechanism, sediment grains behaved like molecules and moved down concentration gradients from areas of high concentration to areas of low concentration. Sediment suspension thus resulted from the movement of sediment away from the seabed (high concentration of sediment) into the water column (low concentration of sediment). Figure 1 represents the concept of diffusion showing the general pattern of motion of the water-sediment mixture. Currently, a molecular-type diffusion mechanism is commonly accepted for horizontal transport of sediments but is no longer accepted for vertical transport since the rate of diffusion in still water is not strong enough to overcome the force of gravity alone. However, diffusion can occur in moving water where the motion of the water counteracts gravity.

Additional mechanisms associated with the turbulent motion of the water were then proposed to supplant the molecular diffusion mechanism. These mechanisms, referred to as convection and advection, are represented in Fig. 1b, c, respectively. Convection and advection occur as a result of the fluid drag of moving water on stationary particles towing the particles along with the water. This fluid drag is referred to as a Reynolds stress and is a velocity-dependent force imparted on the sediment grains by the moving water. The faster the water, the greater the stress on the sediment. Convection occurs as a result of vertically directed Reynolds stresses that carry the water and sediment mixture upward away



Sediment Suspension by Waves, Fig. 1 Mechanisms of sediment suspension: (a) diffusion, (b) convection, and (c) advection

from the seabed. The sediment is maintained in suspension by these fluid stresses that counteract the force of gravity. Advection occurs as a result of horizontally directed Reynolds stresses that carry the water and sediment back and forth across the seabed. Again the sediment is maintained in suspension by fluid stresses that counteract the force of gravity. Over rippled seabeds, convective and advective suspension of sediment is an extremely important factor due to near-seabed turbulence as water passes over the bedforms.

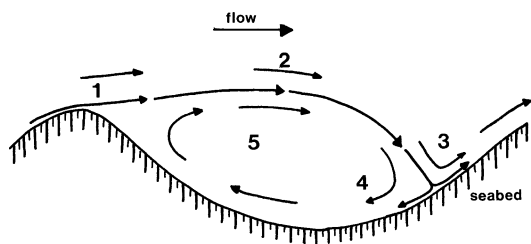
Flow Separation and Vortex Propagation

The near-seabed turbulence that causes convective and advective suspension of sediment is influenced by the presence of bedform-induced flow separation. Flow separation occurs as a direct result of the conservation of energy within the water. Water energy is commonly described using the Bernoulli equation, which equates the total energy of the water as equivalent to the sum of the pressure of the water, the kinetic energy (related to water velocity), and the potential energy (related to the elevation of the water and gravity). Figure 2 illustrates the dynamics of flow separation over a bedform under unidirectional flow (after Honji et al. 1980). In the figure, water is assumed to be flowing from the left to the right.

Immediately over the ripple crest (1), the flow of water just above the seabed expands as the water column deepens into the trough of the ripple. Flow velocity decreases into the trough since there is a greater depth of water through which

the same volume of water passing over the ripple crest must travel. In order to conserve energy in the direction of flow, the pressure of the water at the seabed increases to compensate for the decrease in velocity. If the water is moving fast enough, the increase in pressure at the seabed lifts the flowing water upward as a result of a pressure gradient away from the seabed, and the flow separates. If the water is not moving fast enough, the change in pressure is not sufficient to lift the water away from the seabed. The flow of water (streamline) lifted off the seabed by the pressure difference (1) begins to accelerate as it is no longer slowed by the friction of the seabed (2). As the flow accelerates, velocity increases and, to compensate and conserve energy, pressure decreases. The decrease in pressure reduces the pressure gradient away from the seabed, and the streamline reattaches (3). The streamline from 1–3 creates a free shear layer beneath which flow reverses (4) relative to the direction of the flow above the layer and forms a vortex of rotating water (5) in the lee of the ripple crest. The free shear layer has increased momentum relative to the surrounding water due to its larger velocity, thereby creating a momentum gradient toward the seabed. The momentum gradient traps the rotating flow in the lee of the ripple where it entrains sediment.

Under a wave-dominated environment, the free shear layer that traps the suspended sediment forms and decays twice each wave cycle since the horizontal component of the oscillatory flow becomes zero twice each wave cycle. When the free shear layer decays, the vortex at the seabed is no longer trapped. The vortex generally has larger velocities than the water above, and a momentum gradient is established away from the seabed. The upward-directed momentum gradient causes the vortex to rise away from the seabed, suspending sediment up into the water column (Tunstall and Inman 1975). The vortex is both convected and advected away from the seabed, and it diffuses over time into clouds of suspended sediment that are transported with the flow. The diffusion occurs down the concentration gradient from the vortex to the surrounding water. As the vortex decays, the Reynolds stresses suspending the sediment

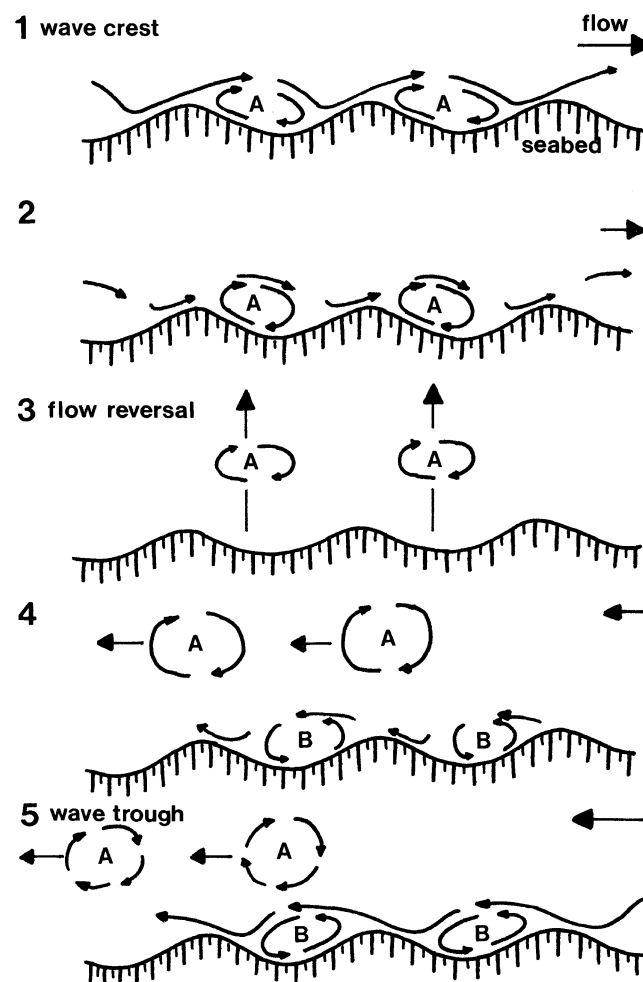


Sediment Suspension by Waves, Fig. 2 Schematic of flow separation and generation of a sediment suspending vortex (After Honji et al. 1980)

decay as well and the suspended material begin to settle out. The formation and subsequent ejection of vortices from the bed is an important mechanism for suspending sediment under waves and results in intermittent injection of suspended sediment into the water throughout a complete wave cycle.

Wave Cycle Model of Sediment Suspension

The intermittent nature of sediment suspension through vortex creation and ejection, or shedding, is significant because it leads to strong temporal variations in concentration and hence strong temporal dependencies in sediment transport. The temporal variations occur over the timescale of a complete wave cycle since the mechanisms controlling vortex shedding vary with the passage of each wave. The wave cycle model of sediment suspension is based on a model of



Sediment Suspension by Waves, Fig. 3 The wave cycle model of sediment suspension (After Sleath 1982)

periodic flow separation and vortex propagation over a rippled seabed that incorporates this intermittent suspension of sediment (Sleath 1982). Figure 3 represents this conceptual model over various stages, or phase angles, of a wave cycle. The motion of water under a wave at the water surface is generally orbital, molecules of water traveling through a 360° orbit between wave crests. The wave orbit becomes flatter toward the seabed as the vertical motion of the water is muted and the water generally flows horizontally first in one direction as a wave crest passes and then horizontally in the other direction as the wave trough passes. This change in flow direction is referred to as oscillatory flow. Given this oscillation, the phase angle of the wave cycle may be referenced relative to the 360° of one complete wave cycle. The start of the wave cycle may be set anywhere in the cycle but is commonly assigned to a maximum velocity in one direction under the wave cycle model. In Fig. 3, the start of the wave cycle has been arbitrarily set to occur with the passage of a wave crest (1).

At the beginning of the wave cycle (1), phase angle $\theta = 0^\circ$, the horizontal velocity of the flow is at a maximum to the right. Flow has separated over the lee slope of the ripple and has generated a vortex downstream of the crest (A). The vortex is trapped at the bed by the large momentum gradient directed toward the seabed from the free shear layer. The vortex entrains sediment from the seabed while trapped and reaches its maximum size and velocity. At this point, velocity is large, but the suspended sediment concentration in the water column is small since the vortex is trapped at the seabed. As the horizontal flow to the right decreases in velocity (2) with the passage of the wave crest, phase angle $\theta = 45^\circ$, the free shear layer decays resulting in a momentum gradient directed upward away from the seabed. The vortex carries the entrained sediment as it begins to rise. When the horizontal velocity reaches zero (3) at the midpoint in the wave between the crest and the trough, phase angle $\theta = 90^\circ$, the flow is on the point of reversal and the free shear layer has vanished. The vortex is ejected from the seabed by the upward transfer of momentum and carries sediment into suspension. At this point, velocity is small and suspended sediment concentration in the water column is large since the vortex has been released upward. With the collapse of the free shear layer, the vortex is no longer constrained, and the suspended material begins to diffuse under viscous and gravitational forces.

Once through the point of reversal, horizontal flow increases to the left (4) with the approach of the wave trough, phase angle $\theta = 135^\circ$. The vortex ejected during the previous phase (A) is transported in the opposite direction to the flow which generated it. This vortex is deflected higher into the water column by the generation of a new free shear layer on the ripple slope opposite to the slope above which it

formed. As the vortex (A) rises, it diffuses into the water column and is advected across the bed by the water that is now flowing to the left. At the seabed a second vortex (B) is generated under the newly developing free shear layer. When the horizontal velocity reaches a maximum to the left (5) with the arrival of the wave trough, phase angle $\theta = 180^\circ$, the flow of water is in the opposite direction to the flow at a phase angle $\theta = 0^\circ$ (wave crest) and the second vortex (B) reaches its maximum size and velocity while trapped at the seabed. The velocity is now large again, but suspended sediment concentration in the water is now dependent on the diffusing and advected vortices generated earlier in the wave cycle.

The second half of the wave cycle covering the passage of the wave trough and the arrival of the next wave crest, phase angles $\theta = 180^\circ$ to $\theta = 360^\circ$, go through these same five stages as the flow reverses from the left to the right. The second vortex (B) is ejected into the water column in the same manner as the first vortex (A) at the next flow reversal. The first vortex continues to diffuse and to be transported back and forth above the seabed so long as the fluid turbulence can maintain the sediment in suspension against the pull of gravity.

Not all waves will be large enough to generate flows at the seabed that will result in separation. The size of the wave needed to suspend sediment at the seabed will depend on the depth of water. If a series of waves pass by in succession, and each generates flows strong enough to develop vortices at the seabed, sediment suspended by the first wave may be kept in suspension until all the waves have passed. Such a series of waves is known as a wave group. Successive waves in a wave group inherit some turbulence from the preceding waves. This inheritance of turbulence leads to the suspension of sediment for periods of time longer than a single wave cycle. The extended suspension events result in the individual vortices coalescing into clouds of suspended material above the bed. Video observations (e.g., Dyer 1980) of sediment suspension under wave groups show the formation of these suspension clouds and their subsequent decay once the wave group has passed.

Modeling Sediment Suspension

Given the conceptual wave cycle model of sediment suspension, it is clear that the marked spatial and temporal variations in sediment suspension are critical in determining overall sediment transport. Bailard (1984) developed a model for sediment transport based on energetics principles proposed by Bagnold (1963) but recognized that the use of time-averaged terms was a significant weakness to the model's predictive ability. The weakness is due to the fact that time-averaged

terms do not represent the physical processes well (e.g., Osborne and Greenwood 1993). The process of vortex shedding acts as a strong control on the suspension of sediment into the water column and provides a significant influence on the resulting concentration profile. Additionally, the time at which concentration reaches a maximum is out of phase with the time at which velocity reaches a maximum. The energetics approach relies on the velocity and concentration maximums occurring together.

Ideally, any model that is developed to predict sediment suspension under waves should account for the suspension characteristics outlined in the wave cycle concept of suspension. Neilsen (1993) investigated several different models for predicting sediment suspension under waves and concluded that simple models worked better than complex models. Simple models work better than complex models because the contribution of many of the variables in the process of sediment suspension is poorly understood as are the interrelationships between the variables. However, the simplest model, where sediment suspension is predicted from the value of the flow velocity, also fails to provide reasonable predictions because suspension maximums occur at a different phase to velocity maximums. Neilsen (1993) concluded that the shift in phase between velocity and concentration had to be included in the derivation of a model.

Atkins (1993) used the wave cycle model of sediment suspension to suggest that a simple model for sediment suspension could be derived as a function of the flow velocity shifted with respect to the suspended sediment concentration by a time lag equivalent to the phase shift between the two variables. The model predicted total suspended sediment concentration as the sum of convective, advective, and diffuse components. The components were predicted from the flow velocity using simple sinusoidal relationships. In experimental tests, the model was able to explain nearly 80% of the variation in sediment suspension measured under a synthesized irregular wave spectrum. Both peaks and lows in suspended sediment concentration were reasonably well predicted both in magnitude and through time. The success of the model suggests that the wave cycle model of sediment suspension under waves is an adequate starting point for further research into suspension mechanisms.

Conclusions

Sediment suspension under waves is a complex interplay between the currently identified mechanisms of convection, advection, and diffusion. This interplay results in strong temporal relationships between water velocity and the concentration of suspended sediment. A group of

successive waves energetic enough to suspend material can keep sediment suspended for extended periods of time. The predictive successes of simple models based on the physical mechanisms provide strong evidence that simple functional relationships between velocity and concentration may be used to predict sediment suspension under waves. A better understanding of the processes controlling the mechanisms of sediment suspension, their time dependencies, and their interrelationships will lead to better predictive models of suspension and, ultimately, better models of sediment transport. Better models of sediment transport would then allow for better predictions of coastal evolution and increase our understanding of how and why coastlines change.

Cross-References

- [Beach Processes](#)
- [Coastal Dynamics](#)
- [Coastal Erosion](#)
- [Cross-Shore Sediment Transport](#)
- [Dynamic Equilibrium of Beaches](#)
- [Erosion of Coastal Systems](#)
- [Erosion Processes](#)
- [Longshore Sediment Transport](#)
- [Nearshore Sediment Transport Measurement](#)
- [Ripple Marks](#)
- [Shoreline Response to Littoral Drift Barriers](#)
- [Waves](#)

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Sedimentary Basins

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Regional subsidence in coastal areas is mainly due to the occurrence of sedimentary basins. Figure 1 indicates the major deltas of the world (Fairbridge and Jelgersma 1990); many situated on top of sedimentary basins. Before giving some examples of basins situated in the coastal zone some information should be given about the origin of those basins.

Sedimentary basins are places of prolonged slow subsidence in general, compensated by sediment input of rivers and the sea. In recent time, however, this sediment input in several deltas was strongly reduced due to the construction of dams and reservoirs in the upstream area of the river and to other human interference. This so called “sand starvation” results in surface lowering and consequently in shore erosion. The subsurface layers of sedimentary basins have been subject to intensive investigations, many studies have been made of their stratigraphic sequences and their relation to sea-level changes.

During the last decades the concept of plate tectonics has given a more detailed idea as to how these basins are formed. The lithosphere consists of a number of plates which are in motion. Accordingly, sedimentary basins exist in an environment of the motion of these plates. Basins can either be formed by collision of plates or by lithospheric stretching of the plates. The first mentioned activity gives rise to basins due to flexures, they are present as foreland basins at the foot of mountain belts. The Bengal basin of Bangladesh and the Po basin of Italy are examples of these flexures and will be discussed below. Lithospheric stretching of plates causes fault zones which become later, due to thermal cooling, important areas of subsidence. Examples of these rift basins are the

North Sea basin and the Niger basin. Another genetic class of rift basins are associated with strike-slip deformation. The Tertiary rift basins onshore and offshore of Thailand and Malaysia are associated with this phenomenon. A few samples of sedimentary basins present underneath important deltas are presented below.

The Po Delta

The river plain overlying the Po–Veneto sedimentary basin is enclosed by the Alpine mountain chain in the north and by the Apennines in the south. Collision of the African plate with the European plate in the Late Cretaceous was responsible for this landscape. The Po–Veneto basin came into being and due to the slow but continuous subsidence, thick layers of sediment were deposited. Mapping the base of the Pliocene–Quaternary sequence has indicated several fault zones in the southeast part of the plain. The Quaternary deposits reach a thickness of more than 1,000 m.

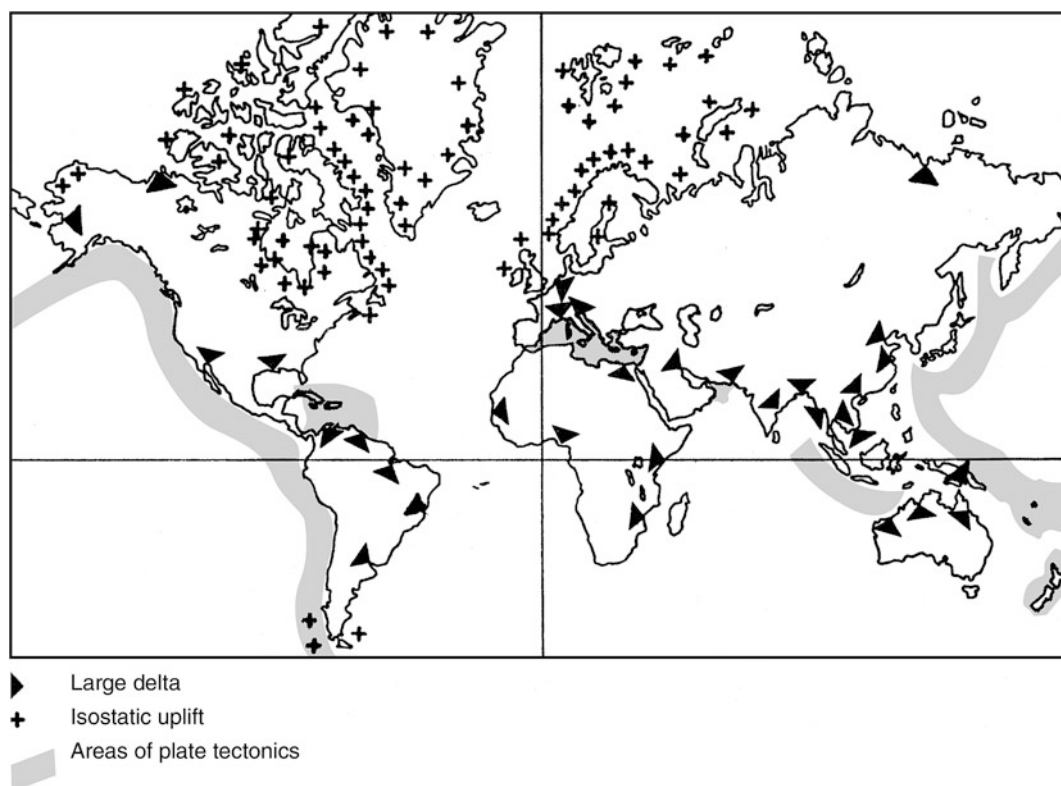
In geological terms, the lobate Po delta is a very young phenomena which came into being in historical time by an increase in sediment discharge caused by deforestation in the uplands and an increase in precipitation (Little Ice Age). In recent time reforestation, construction of dams and reservoirs, and human activities in the delta itself have caused sediment starvation, subsidence, and consequently shore erosion.

Ganges–Brahmaputra Delta

The Bengal basin is a product of the collision of the Indian plate with the Burmese and Tibetan plate. From the Cretaceous onwards thick layers of sediments have been laid down in this basin due to tectonic and isostatic subsidence. The uplift and erosion of the Himalayas mountain range are the main sources of the sediments filling the basin. The thickness of the sedimentary sequences is demonstrated in Fig. 2 by a contour line on the depth of the basin (Jones 1985; Alam 1996). Subsidence rates are thought to be 2–5 cm/century.

The Ganges–Brahmaputra delta and plain of India and Bangladesh have many fault zones which may have contributed to the shifting of river courses and the differences in surface lowering. Other tectonic phenomena are the occurrence of many earthquake epicenters. The Meghna fault zone, NE–SW direction, continues offshore in the Bay of Bengal and ends in the canyon of the Swatch of no Ground.

The alluvial and deltaic plain are subject to seasonal flooding by rivers and cyclonic surges. Due to the high tidal range the delta can be classified as tide-dominated.



Sedimentary Basins, Fig. 1 Location map of coastal instability. (From Fairbridge and Jelgersma 1990; with the kind permission of Kluwer Academic Publishers)

Niger Delta

The Niger delta has been developed since the Cretaceous when tectonic events culminated in seafloor spreading between the African and South American lithospheric plates. The post-drift period was characterized by instability and subsidence, due to thermal cooling, in the rift valleys and the rift margins. Due to this subsidence more than 8,000 m of Tertiary and Quaternary sediments are present (Fig. 3) in the Niger sedimentary basin (Whiteman 1982). Like all deltas in the world the recent delta sediments consist of a wedge of fine-grained sediments formed as a result of the post-glacial rise in sea level.

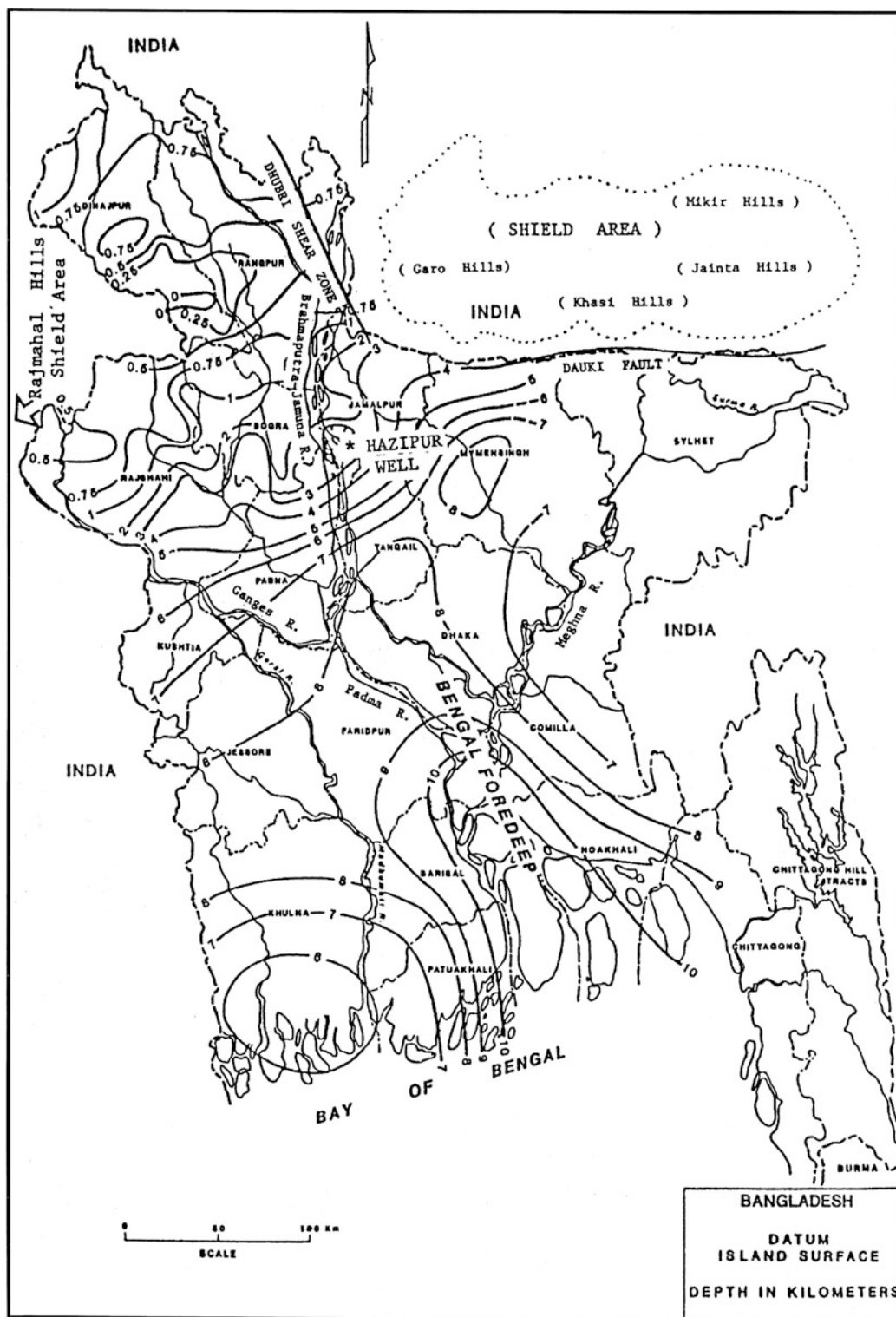
The coast of the delta is formed by small barrier islands separated by inlets and backed by an extensive area of mangrove swamps. The delta can be classified as wave-dominated with important input of river sediment. The latter has compensated for the subsidence but in recent time due to the construction of several dams and reservoirs the sediment supply is strongly reduced. This sediment starvation in the delta and the shore has resulted in important coastline retreat. Human activity in the delta,

mining of gas, oil and water, has also contributed to surface lowering.

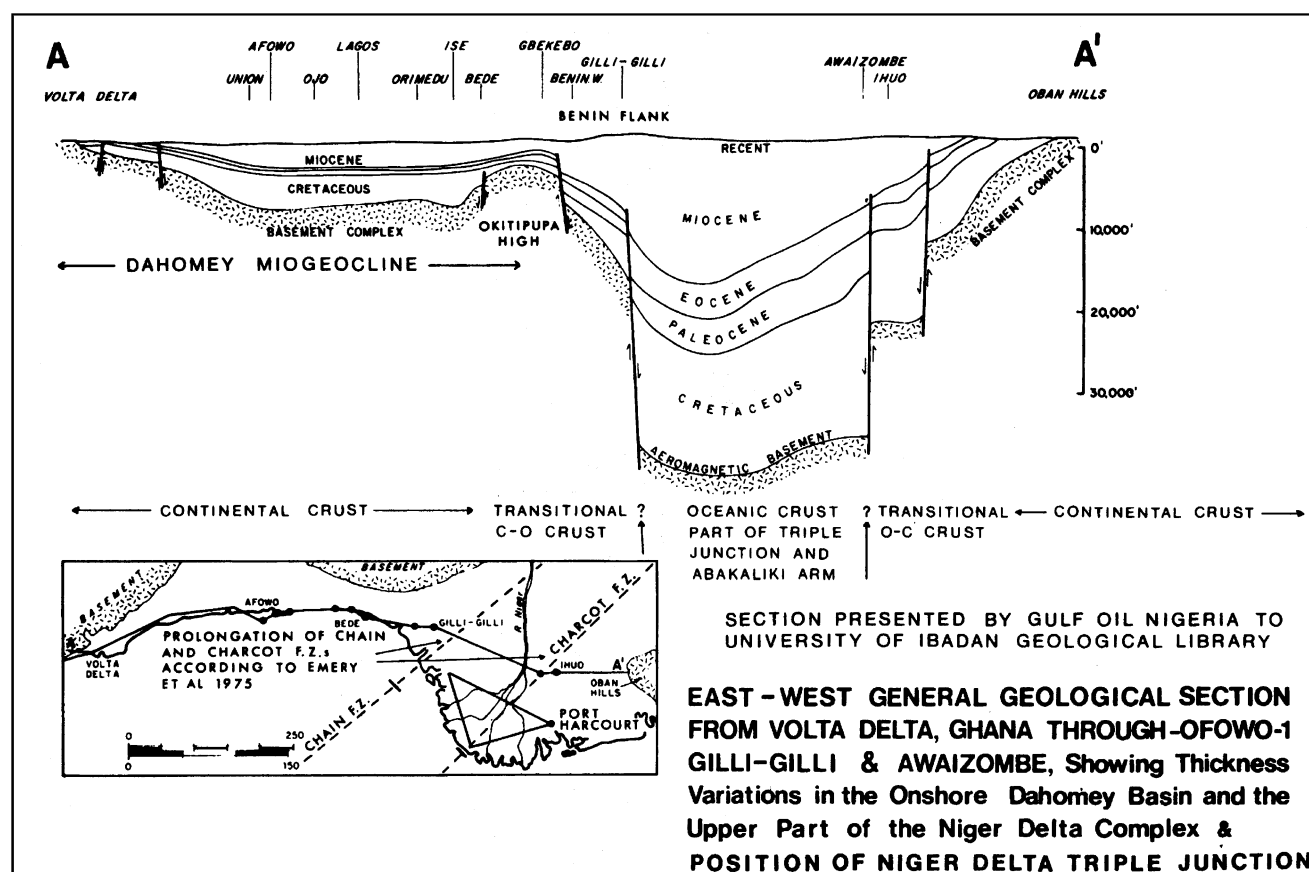
North Sea Basin

The North Sea sedimentary basin developed during Late Cretaceous and Early Tertiary rifting phases marking the onset of seafloor spreading in the Arctic North Atlantic between the European and the North American–Greenland plate. At the same time the early phases of the Alpine orogeny took place.

Thermal cooling in the rift systems of the Early Tertiary caused important subsidence. Consequently, thick layers of Cenozoic sediments could be deposited. As demonstrated in Fig. 4, the Cenozoic series reaches a maximum thickness of 3,500 m of which the Quaternary reaches a maximum of 1,000 m (Ziegler and Louwerens 1977; Jelgersma 1980). The Netherlands and NW Germany are situated on the edge of the subsiding North Sea basin. Tectonic subsidence occurs in these areas, contributing to a relative sea-level rise. The combined effect is



Sedimentary Basins, Fig. 2 Depth to basement in the Bengal basin of Bangladesh. (From Alam 1996; with the kind permission of Kluwer Academic Publishers)



Sedimentary Basins, Fig. 3 Cross-section of the Niger sedimentary basin. (From Whiteman 1982; with the kind permission of Kluwer Academic Publishers)

measured by tide gauges, its amount being between 15 and 20 cm/century.

Mississippi Delta

Sedimentation in the northern Gulf of Mexico has taken place since the end of the Cretaceous in a series of depocenters on the edge of the continental shelf. Sediment accumulation in the depocenters during the Neogene and the Quaternary reached high values as demonstrated in Fig. 5 by the depth contour line of the base of the Pleistocene and a cross section of the depocenter (Woodbury et al. 1973; Jelgersma 1996).

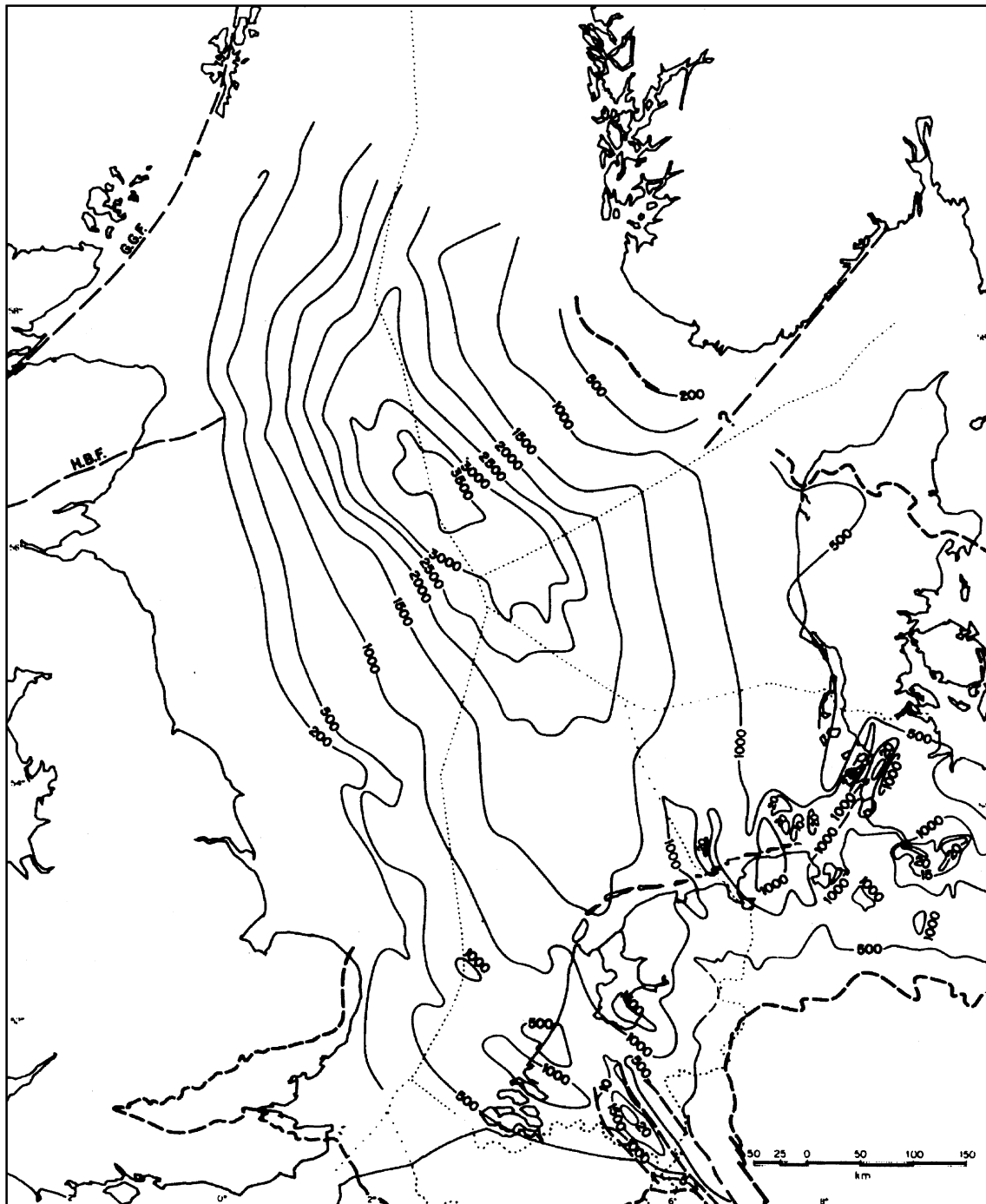
The present Mississippi delta consists of a wedge of soft sediment deposited during the post-glacial rise in sea level. The river delta can be classified as river-dominated; the tidal range is nearly absent and serious wave attack occurs only during hurricanes.

Subsidence rates in the delta are locally very great; this is mainly due to human interference. Accordingly, much of the

marshland surrounding the river delta itself has changed into lakes.

Nile Delta

Crustal movements in the western Mediterranean during the Late Miocene obstructed the connection between the Atlantic Ocean and Mediterranean in the Straits of Gibraltar. The interrupted inflow from the Atlantic led to a gradual desiccation of the Mediterranean. Due to this lowering of the base level a deep canyon was cut down in the area of the recent Nile delta. This Late Miocene channel reaches in the northern embayment of the delta to a depth of 4,000 m. In the beginning of the Pliocene, a renewed opening between the Atlantic and the Mediterranean took place. The inflow of ocean water restored the level of the Mediterranean, consequently the Nile canyon was flooded and a thick layer of Pliocene, Pleistocene, and Recent deposits could be formed and is present in the subsurface of the delta.



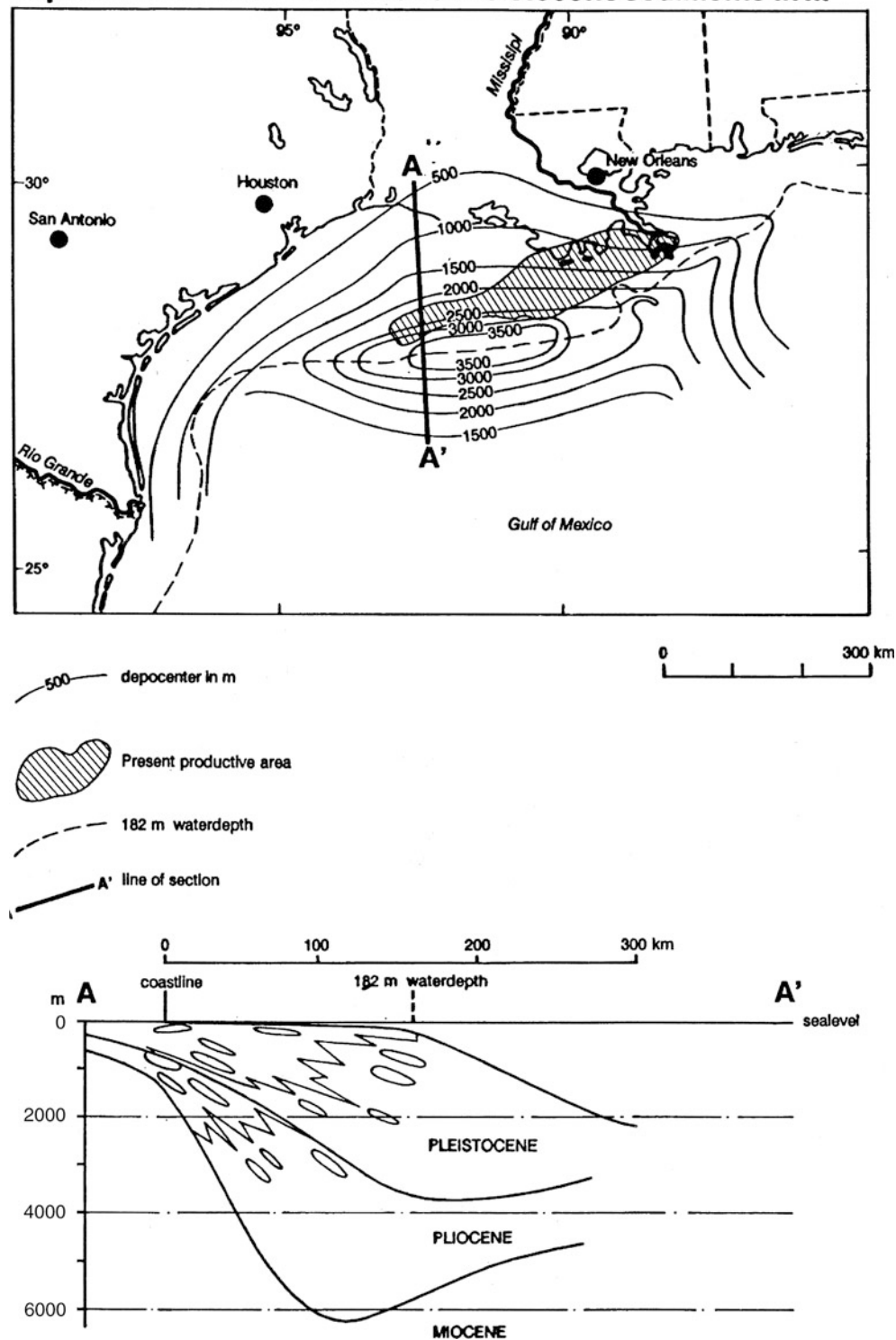
Sedimentary Basins, Fig. 4 Depth contour of the base of the Tertiary. (From Jelgersma 1980, reproduced with permission of John Wiley & Sons Limited)

In Fig. 6, a contour map of the base of the Pliocene is presented; the Pliocene and the Quaternary deposits dip northwards reaching a thickness of 4,000 m (Said 1981; Jelgersma 1996). It is evident that in the northern embayment the subsidence reaches high values, probably caused by fault zones. The Nile delta has been a river-dominated

delta; the input of river sediment compensating for the basin subsidence. During the last decades, due to the construction of the High Aswan dam, sediment input in the delta has been strongly reduced. This resulted in serious shore erosion and salt-water intrusion. The delta then changed from river- to wave-dominated.

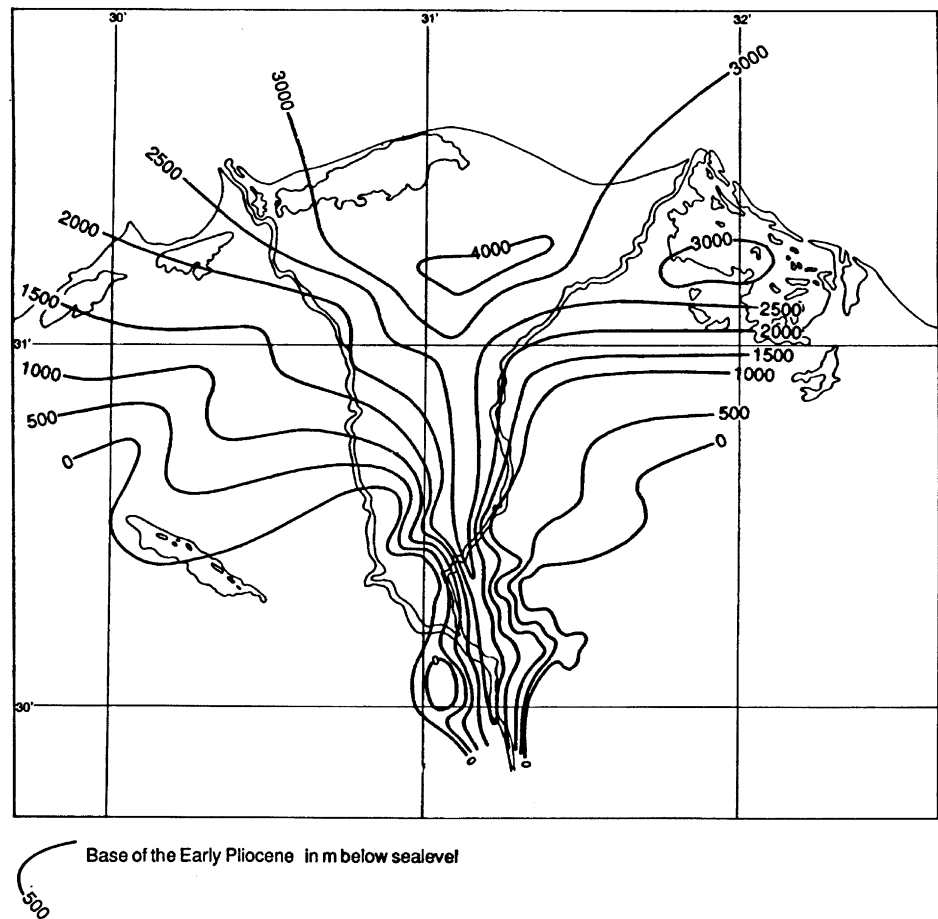
Sedimentary Basins,

Fig. 5 Depth contour of the base of the Pleistocene and a cross-section of the depocenter of the Mississippi. (From Jelgersma 1996; with the kind permission of Kluwer Academic Publishers)

Depocenter of Late Pliocene and Pleistocene sediments in m

Sedimentary Basins,

Fig. 6 Depth contour of the base of the Early Pliocene in the Nile delta. (From Jelgersma 1996; with the kind permission of Kluwer Academic Publishers)

**Cross-References**

- Coastal Changes, Gradual
- Coastal Subsidence
- Dams, Effect on Coasts
- Deltaic Ecology
- Deltas
- Sequence Stratigraphy
- Submerged Coasts
- Submerging Coasts

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Seismic Displacement

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When topographical changes are produced by an earthquake or a succession of earthquakes, a distinction is usually made between precursory displacements that occurred before the event (preseismic), the coseismic displacements during the event, and postseismic changes shortly later, often accompanying after-shocks.

Seismic displacements have a horizontal and a vertical component. The horizontal component is often clearly visible along fault lines activated by the earthquake, though more or less regular horizontal deformation may extend along the surface of nearby crustal blocks. In coastal areas, the vertical component is easily measurable and most important from a geodetic point of view, because it changes the relation to sea level.

As noted by Vita-Finzi (1986) earthquakes were being documented long before progressive uplift or depression attracted the attention of geologists. Several past events, important for their casualties and destructions, have been reported by ancient writers. Most past observers tend to say little or nothing about any ground movements accompanying an earthquake. Nevertheless, some exceptions exist that mention sudden seismic changes in the past, though seldom separating the effects of an earthquake from those of its after-shocks. According to several historical sources (Guidoboni et al. 1994) in the summer of BC 426, when an earthquake struck the Gulf of Malia, in Greece, the central part of Atalante, near Euboea, was split open to the extent that ships could pass through, and some of the plains were flooded as far as 20 stades (ca. 4 km). It is also reported that in a winter night in BC 373, after a violent earthquake, the city of Helice, in the Gulf of Corinth, was drowned by the sea with all of its inhabitants. (Schwartz and Tziavos 1979).

From ancient Japan, several examples of seismic displacements in coastal areas are reported by Imamura (1937). In AD 684, at the time of the Tosa earthquake in Shikoku Island, cultivated ground to the extent of 8.25 km² is said to have sunk beneath the sea. On September 4, 1596, at the time of a series of earthquakes, the island of Uryū-jima, slightly off the present city of Oita, with an area of 4 km N-S by 2.3 km E-W and a population of 5000, sank beneath the waves, where it now lies at a depth of 30–40 fathoms.

Seismic changes in land level are often mere effects of the tilting of crustal blocks. In Shikoku Island, during the 1707 earthquake, subsidence at Kochi reached nearly 2 m, and upheaval 2.1–2.4 m near Yoshiwara, while two belts of subsidence occurred in the western half of the province, indicating that two contiguous crustal blocks had tilted, both of which dipping north.

Important seismic vertical displacements have been reported also from coastal regions in other parts of the world. In the delta area of the river Indus, at the time of the Kutch earthquake, on June 19, 1819, a fault scarp 3 m high appeared, extending for a distance of 80 km parallel to the coast. The region SE of this locality, about 5200 km² in area began to sink, and within 3 h after the earthquake had changed into a sea.

In Chile, after the Valparaíso earthquake on November 19, 1822, the coast had risen 1 m, while at Quintero, some 24 km northwards, it was 1.2 m. According to Lyell (1875), the whole country from the foot of the Andes was raised on this occasion, the maximum rise being at a distance of about 3.2 km offshore.

Rigorous measurements of topographical changes produced by earthquakes started to be carried out toward the end of the nineteenth century. The geodetic investigation of ground deformation immediately after an earthquake, though often logistically problematic, offers the advantage that the area to be surveyed can be identified without much difficulty. The first post-event triangulation was done in Sumatra in 1892. The second survey was executed after the 1897 Indian earthquake and found a maximum vertical displacement of 3.6 m. Repeated leveling surveys carried out in seismic areas before and after a major earthquake, have helped to clarify the local pattern of crustal deformation. Tide gauge records, when located in areas affected by earthquakes, can also provide very precise, continuous information on all the preseismic–coseismic–postseismic sequence. Recently, complete control of land displacements can be furnished all over the world by satellite geodesy.

For the past, coastal coseismic displacements deduced from stepped elevated coastlines have been reported from several seismically active parts of the world. In some cases, detailed analysis of closely dated fossil coastlines (Fig. 1) promises to reveal regional forms of the geoid at successive time intervals and thus extend the brief period spanned by satellite data into the Holocene and possibly beyond.

Preseismic displacements are difficult to demonstrate for past events, as well as in historical times. This is because, in most cases, they are not reported. Preseismic movements are unexpected, may last for decades, and be sufficiently slow to escape notice, also in coastal areas. Such was the case, for example, for Cefalonia Island (Greece), before the 1953 earthquake that caused coseismic

Seismic Displacement,

Fig. 1 A crustal block approximately 200 km long, including all the western part of Crete and the nearby Antikythira Island (Greece), was uplifted and tilted coseismically by a great earthquake, probably on July 21, 365 AD. The upheaval reached 9 m in the southwestern part of Crete island (Pirazzoli et al. 1996). In this photograph (No. 3980, Oct. 1977), taken near Piper Eliá, on the east side of Gramvousa Peninsula, Crete, continuous erosional marks left by the sea level before the uplift are well visible on the limestone cliff at about 5 m in elevation



uplift of 30–70 cm along the south coast of the island. In the 1990s, however, a survey carried out along the uplifted coast (Pirazzoli et al. 1994) was able to demonstrate that, near Karavomilos, a slow 15 ± 5 cm preseismic submergence must have occurred within a few years before the earthquake, therefore preceding the coseismic uplift that reached about 50 cm at this site.

Cross-References

- [Changing Sea Levels](#)
- [Faulted Coasts](#)
- [Submerging Coasts](#)
- [Tectonics and Neotectonics](#)
- [Uplift Coasts](#)

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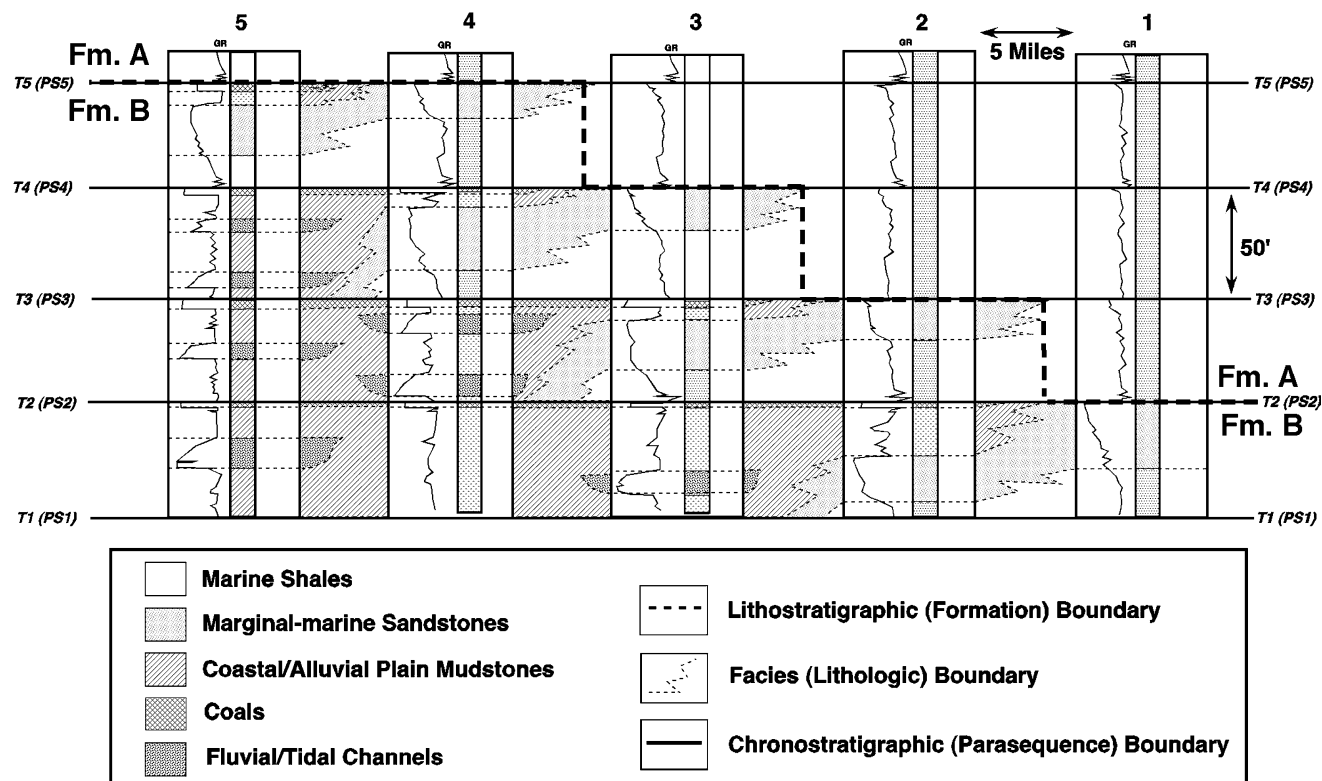
Sequence Stratigraphy

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Sequence stratigraphy is an informal chronostratigraphic methodology that uses stratal surfaces to subdivide sedimentary successions. Unlike most traditional lithostratigraphic units (NACSN 1983), which are defined as regionally mappable packages (members, formations, groups) of similar lithologies (rock types), sequence stratigraphic units trend across traditional lithostratigraphic boundaries (Fig. 1). This methodology owes its origins to the pioneering work of Caster (1934), Sloss (1963), Campbell (1967), and Asquith (1970). All of these workers documented that stratal surfaces trend across traditional lithostratigraphic boundaries, and concluded that stratal surfaces represent time-significant boundaries that can be used to define coeval packages of strata contained within different lithostratigraphic units (Fig. 1).

The original concept of a sequence, a stratigraphic unit bounded by unconformities and their correlative conformities, was outlined by Sloss (Sloss et al. 1949; Sloss 1963). Using the Phanerozoic sedimentary succession of the North American craton as an example, Sloss defined six interregional unconformities and six unconformity-bounded units which he termed sequences. While this methodology gained little interest among most geoscientists in the 1950s, 1960s, and early 1970s, it did find a niche following among

Facies, Lithostratigraphic, and Sequence Stratigraphic (Chronostratigraphic) Correlations



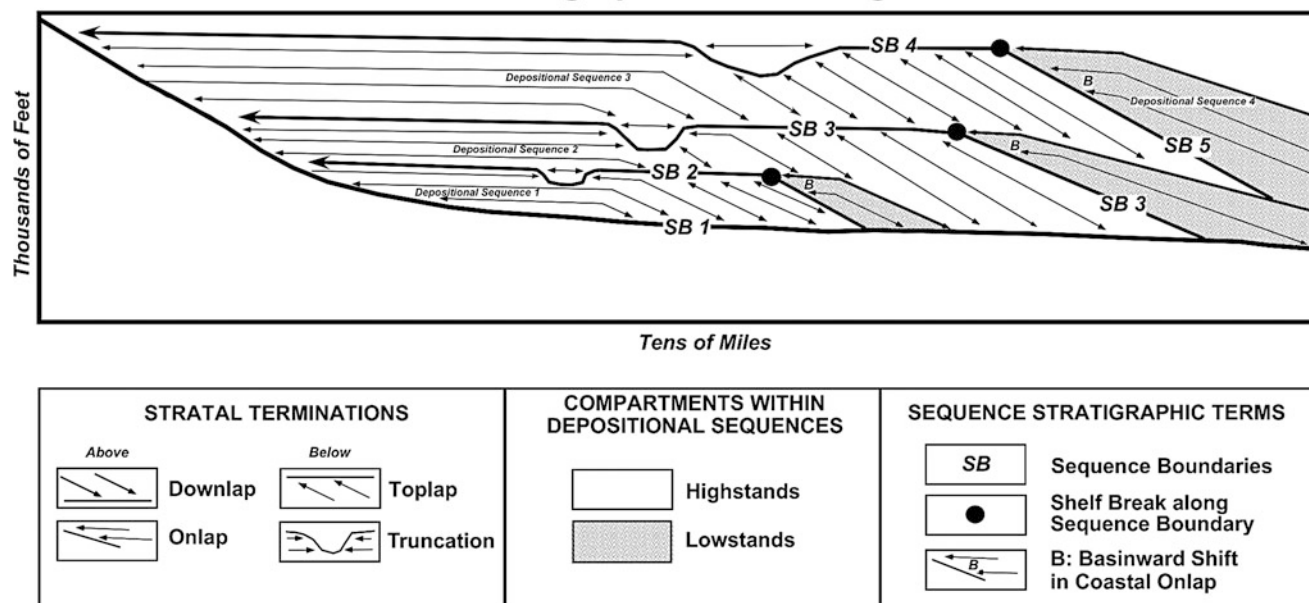
Sequence Stratigraphy, Fig. 1 Idealized well-log cross section that illustrates the differences among facies, lithostratigraphic, and sequence stratigraphic (chronostratigraphic) correlations. Based on detailed correlation of well-log markers, five (5) regional flooding surfaces (parasequence boundaries) were identified. The parasequence boundaries represent chronostratigraphically significant surfaces that separate older strata below from younger strata above. Each parasequence contains a coeval facies succession that is dominated by marine shales

downip and coastal/alluvial plain deposits updip. The chronostratigraphic framework provided by the stratal (parasequence) correlations document that the traditional lithostratigraphic (formation) boundaries are time-transgressive. In the example provided, the lithostratigraphic boundary between Formation A (marine shale) and Formation B (marine shales, marginal marine sandstones, coastal plain mudstones, coals, and fluvial/tidal channels) occurs at time T2 (PS2) in Well 1 and time T5 (PS5) in Well 5

petroleum geologists who were using seismic data to determine subsurface stratigraphy. With the publication of the landmark "Memoir 26" on seismic stratigraphy by the American Association of Petroleum Geologists (AAPG) in 1977 (Payton 1977), however, sequence stratigraphic methodology and concepts became an active topic of research and debate among geoscientists. Contained within this volume is the now classic 11-part paper written by researchers at Exxon Production Research Company (Vail et al. 1977a). This entry outlined the fundamentals of seismic stratigraphy and offered the Depositional Sequence as the basic unit for stratigraphic analysis. Please note that while the unconformity-bounded sequence of Sloss (1963) covered thousands of feet and were typically a hundred million years in duration, the depositional sequences of Vail et al. (1977a, b, c, d) were commonly hundreds of feet thick and a million years in duration. As defined by Mitchum (1977), the *Depositional Sequence* is "... a stratigraphic unit composed of a relatively conformable

succession of genetically related strata and bounded at its top and base by unconformities or their correlative conformities." The basic premise behind this methodology was the observation that "... primary seismic reflections are generated by physical surfaces in the rocks, consisting mainly of stratal (bedding) surfaces and unconformities with velocity-density contrasts." Furthermore, Vail and Mitchum (1977) concluded that since "... all the rocks above a stratal or unconformity surface are younger than those below it, the resulting seismic section is a record of the chronostratigraphic (time-stratigraphic) depositional and structural patterns and not a record of the time-transgress lithostratigraphy (rock stratigraphy)." Using datasets from the Cretaceous of North America and the Tertiary of South America, Vail et al. (1977d) illustrated that seismic stratigraphic surfaces trend oblique to traditional lithostratigraphic boundaries and define coeval facies successions among adjacent formations. Sequence stratigraphic analysis, as defined by Vail and Mitchum

Seismic Stratigraphic Methodologies



Sequence Stratigraphy, Fig. 2 Seismic stratigraphic analysis, as outlined by Vail et al. (1977a, b, c, d), was based on the identification of depositional sequences: a relatively conformable succession of genetically related strata. The upper and lower boundaries of depositional

sequences are surfaces defined by unconformities and their correlative conformities (sequence boundaries). Sequence boundaries can be objectively identified on seismic as through-going surfaces defined by stratal terminations both above and below

(1977), was based on the identification of stratigraphic units composed of a relatively conformable succession of genetically related strata termed depositional sequences. The upper and lower boundaries of depositional sequences are surfaces defined by unconformities and their correlative conformities. These surfaces are termed *Sequence Boundaries*. Vail et al. (1977a, b, c, d) believed that sequence boundaries (Fig. 2) could be objectively identified on seismic as through-going surfaces defined by stratal (reflection) terminations both below and above. Stratal (reflection) terminations beneath sequence boundaries included truncation and toplap, while stratal (reflection) terminations above sequence boundaries included onlap and downlap (Mitchum 1977). It should be noted, however, that in subsequent sequence stratigraphic publications (Van Wagoner et al. 1990) downlap was no longer utilized as criteria for sequence boundary identification. It is now believed that downlap more commonly occurs within sequences, and should be used to define compartments (systems tracts) within sequences.

Because the characteristics of sequence boundaries and sequences appeared similar in a wide variety of basins through geologic time, Vail et al. (1977b) concluded that cycles of relative sea-level change were the primary controlling factor in sequence development. In the seismic stratigraphic methodologies outlined in Vail et al. (1977b), relative sea-level rises were interpreted by (coastal) onlap, relative sea-level stillstands by (coastal) toplap, and a relative sea-level falls by a downward (basinward) shift in (coastal) onlap

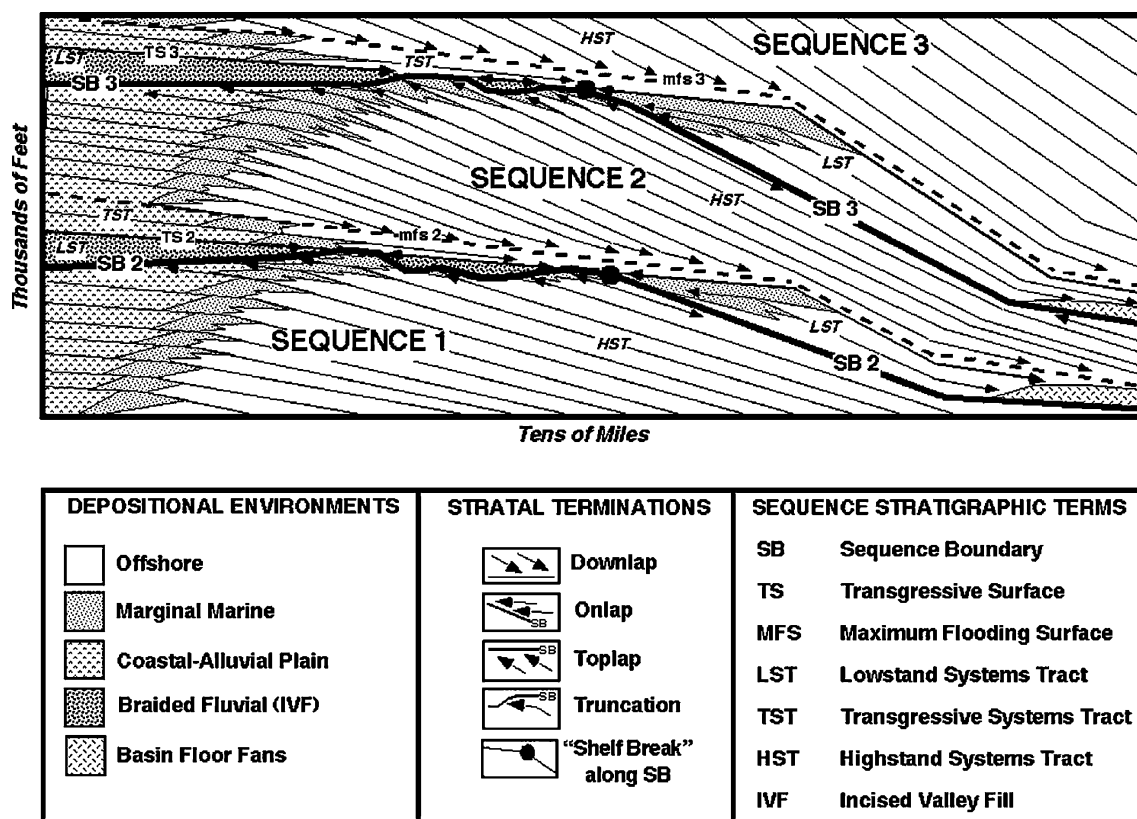
(Fig. 2). Within this paradigm, the physiographic break “shelf edge” along the basal sequence boundary serves as a major reference point for a given sequence. This shelf edge (Fig. 2) permits delineation of lowstands, the interval of time when sea level (coastal onlap) is interpreted below the shelf edge, and highstands, the interval of time when sea level (coastal onlap) is interpreted above the shelf edge (Vail et al. 1977b).

One final premise contained within the paper by Vail et al. (1977a, b, c, d) is that global cycles of sea-level change are evident throughout the Phanerozoic, and that these cycles are globally synchronous. This contention was based on the belief of the Exxon researchers that the depositional sequences they defined on different continental margins were coeval and of similar magnitude (Vail et al. 1977c). They concluded that these interpreted global cycles were records of geotectonic, glacial, and other large-scale processes and reflected major events of the Phanerozoic history. Based on this paradigm, Vail et al. (1977c) offered preliminary cycle charts to document the timing of the global (eustatic) events that controlled sequence development. Haq et al. (1987, 1988), as well as de Graciansky et al. (1998), have subsequently published updated versions of “The Cycle Charts.”

With the publication of SEPM Special Publication Number 42 in 1988 (Wilgus et al.), seismic stratigraphy evolved into sequence stratigraphy. This volume also contained numerous articles by Exxon researchers who presented updated sequence models based on more modern seismic, as well as the integration of outcrop, core, and well-log data.

A third publication by Exxon researchers (Van Wagoner et al. 1990) continued to refine sequence stratigraphic methodologies based on outcrop, core, and well-log data. In this volume, Van Wagoner et al. (1990) proposed a hierarchy of stratal surfaces (beds, bedsets, parasequence, parasequence sets, sequences) all of which had chronostratigraphic significance. While the depositional sequence continued to be the fundamental stratal unit in sequence stratigraphy, parasequences were identified as the building blocks of sequences. Parasequences were defined as relatively conformable succession of genetically related strata bounded by marine-flooding or correlative surfaces (Van Wagoner et al. 1990). Parasequence sets were defined as a succession of genetically related parasequences that form a distinctive stacking pattern (Van Wagoner et al. 1990). Based on outcrop and subsurface observations Van Wagoner et al. (1988, 1990) proposed

three basic compartments or *systems tracts* (lowstand, transgressive, and highstand) within sequences (Fig. 3). These systems tracts were defined by major stratal boundaries, parasequence set stacking, as well as temporal and spatial position within sequences. Surprisingly, they were not defined by interpreted sea-level variations as their names might suggest. The lowermost compartment the *lowstand systems tract* is bounded by the sequence boundary at its base and the transgressive surface at its top (Fig. 3). The *transgressive surface* is defined as first significant marine-flooding surface (parasequence boundary) across the shelf within a depositional sequence. In higher-relief basins (Fig. 2), where the basal sequence boundary displays a well-defined “shelf break” (slope of 3–6 degrees), the lowstand systems tract consists of a basin-floor lowstand fan, a basinally restricted lowstand wedge, and an incised valley system that extends



Sequence Stratigraphy, Fig. 3 Idealized Model of a Depositional Sequence for a basin with distinct physiographic relief (“shelf break model”). In this updated sequence stratigraphic model, first proposed by Van Wagoner et al. (1988), three basic stratal compartments (systems tracts) were proposed: lowstand, transgressive, and highstand. The *lowstand systems tract* is the lowermost compartment within a sequence. It is bounded by a sequence boundary at its base and a transgressive surface at its top. The *transgressive surface* is defined as first significant marine-flooding surface (parasequence boundary) across the shelf within a depositional sequence. The lowstand systems tract typically consists of a basin-floor fan, a basinally restricted lowstand wedge (aggrading parasequence set) that onlaps the “shelf break” of the underlying

sequence boundary, and an incised valley that extends updip across the “shelf-profile” of this sequence boundary. The *transgressive systems tract* is the middle compartment within a sequence. It is bounded below by the transgressive surface, and above by the maximum flooding surface. The *maximum flooding surface* is defined as the maximum marine incursion (transgression) within a sequence. Parasequences within the transgressive systems tract form a distinctive retrogradational parasequence set. The *highstand systems tract* is the uppermost compartment within a sequence. It is bounded below by the maximum flooding surface and above by the overlying sequence boundary. Parasequences within the highstand systems tract form an aggradational and then progradational parasequence set

updip across the shelf, landward of the lowstand wedge (Van Wagoner et al. 1990). In lower-relief “ramp-type” basins, where the basal sequence boundary lacks a well-defined “shelf break” (slope, 1 degree), lowstands lack basin-floor fans (Van Wagoner et al. 1990). The lowstand wedge consists of aggrading parasequences that onlap at or below the depositional shelf break of the underlying sequence boundary. The *transgressive systems tract* is the middle compartment within sequences. It is bounded below by the transgressive surface, and above by the maximum flooding surface. The *maximum flooding surface* is defined as the maximum marine incursion (transgression) within a sequence. Parasequences within the transgressive systems tract form a distinctive retrogradational pattern. The *highstand systems tract* is the uppermost compartment within a sequence. It is bounded below by the maximum flooding surface and above by the overlying sequence boundary. Parasequences within the highstand systems tract form an aggradational and then progradational parasequence set.

As outlined by Van Wagoner et al. (1990) sequence boundaries, sequences, and systems tracts can be interpreted in terms of cycles of relative sea-level change. Sequence boundaries are interpreted to be the result of relative sea-level falls (Van Wagoner et al. 1990). If slope failure, canyon development, and fluvial capture occur during this fall, mass transport complexes followed by submarine fans complexes can be locally deposited along the basin floor (Van Wagoner et al. 1990). As the rate of relative sea-level fall decreases, reaches a stillstand, and slowly rises, fluvial incision and submarine fan deposition ceases. Coarse-grained braided fluvial and/or estuarine deposits aggrade within the incised valleys, and a lowstand wedge is deposited below the depositional shelf break of the underlying sequence boundary (Van Wagoner et al. 1990). When the relative rate of sea-level rise increases toward its maximum, retrogradational parasequences of the transgressive systems tract are deposited (Van Wagoner et al. 1990). As the rate of relative sea-level rise begins to slow and reach a stillstand, aggradational followed by progradational parasequences of the highstand systems tract are deposited (Van Wagoner et al. 1990). As relative sea level falls once again, the next sequence boundary forms by fluvial incision, slope failure, and canyon development.

Since the publication of AAPG Memoir 26 in 1977, sequence stratigraphy has become a major research topic in sedimentary geology. Inclusion of industry, government, and academic researchers has produced healthy debate in a number of areas. From day one, a major issue of debate (Miall 1986, 1992) was the contention by Vail et al. (1977c) that depositional sequences are the product of eustasy, and therefore form globally synchronous cycles that can be used as a basis for global chronostratigraphy (The Cycle Charts). Discussions on the utility of defining and mapping sedimentary cycles (sequences) based on subaerial unconformities

(*Depositional Sequences*: Van Wagoner et al. 1990), maximum flooding surfaces (*Genetic Sequences*: Galloway 1989), or transgressive surfaces (*T–R Sequences*: Embry 1993) has also occurred. Even among practitioners of “depositional sequences,” differences in sequence stratigraphic methodologies has arisen. The classic Exxon papers in AAPG Memoir 26 and Methods in Exploration Number 7 (Mitchum et al. 1977; Van Wagoner et al. 1990) defined sequences and systems tracts purely by stratal geometries. Within this context, interpreted sea-level variations, facies, or increments of geologic time were not considered in sequence boundary or system tract definition. Posamentier et al. (1992), however, based sequence boundary placement and system tracts definition on interpreted sea-level variations. Plint and Nummedal (2000) defined sequence boundaries by stratal termination, but systems tracts by interpreted sea level variations. Mellere (1994) based system systems tracts on facies change, not stratal compartments. Erskine and Vail (1988) based sequence boundaries and system tracts on interpreted cycles of geologic time. Finally, while Van Wagoner et al. (1990) contended that transgressive erosion was not a significant process in the rock record, others such as Plint and Nummedal (2000), strongly argue that transgressive erosion can significantly erode and modify highstand, as well as lowstand, depositional patterns.

Cross-References

- ▶ [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
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Setbacks

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Coastal management strategies reduce the risks associated with coastal hazards and protect coastal infrastructure, habitat, water quality, recreation values, and aesthetic properties. A *setback* is one type of regulatory method used by all levels of government to mitigate risks to coastal structures and to protect coastal resources. For example, following Hurricane Luis in 1995, three island territories in the eastern Caribbean (Antigua and Barbuda, Nevis, and St. Lucia) developed a shoreline (defined here as the high water line) management strategy involving the establishment of setbacks (Cambers 1997).

What Is a Setback?

Taken literally, setbacks are a type of regulatory restriction that require coastal construction projects to “set back” a landward distance from a predetermined reference feature on the beach. This arrangement provides a buffer between a hazard area or natural area and coastal development. Setback lines often parallel (but are situated some distance inland from) a reference feature such as the high water line, vegetation line, dune crest, contour elevation, or cliff edge. Reference features are usually determined by the geomorphic and ecologic properties of a particular reach. The setback is the measured horizontal distance from the reference feature to the

setback line. The method used to establish the horizontal distance is a key component in the process of establishing setbacks because coastal structures are permitted landward of the setback line, but not seaward without a variance. Therefore, setbacks can have substantial implications for state and local governments, managers, planners, developers, and citizens.

The overall purposes of a setback are to:

- Provide a buffer between coastal development and coastal hazards, for example, flooding and erosion;
- Protect or establish natural resources such as conservation areas;
- Facilitate sustainable coastal development;
- Ensure public access to the beach;
- Maintain the aesthetic character of the coast.

Components and Types of Coastal Setbacks

Coastal setbacks require one or more of the following components:

- A reference feature
- A feature deemed worthy of protection (for example, a natural resource)
- A method for establishing a setback line.

Reference Features

Reference features provide a starting point from which to establish a setback distance. Reference features can be stationary (relatively permanent) or dynamic. Stationary reference features can include a surveyed elevation, dune toe or crest, vegetation line, road or other cultural feature (for example, lighthouse). Dynamic reference features typically include the shoreline (i.e., the intersection of the land and sea), a proxy for the shoreline (for example, high water line, HWL; mean high water line, MHWL; ordinary high water line, OHWL; normal water line, NHWL; etc.) or a cliff (also called bluff) edge. For example, Florida employs a setback from the seasonal high water line (SHWL) for coastal permitting. Wisconsin uses a statewide setback of 75 f. (22.9 m) from the ordinary high water mark (OHWM) and 75 f. (22.9 m) from the cliff edge to protect structures from cliff erosion.

Natural Resources

Setbacks provide one method of managing coastal resources including their features, processes, and characteristics (Clark 1974). Natural resources in the coastal zone can include coastal habitats and ecosystems (wetlands, estuaries, beaches, dunes, uplands), sand and gravel, small-town characteristics, visual openness and surf (Houlahan 1989; Nelson and Howd

1996; Klee 1999). According to the US Fish and Wildlife Service (1995), more than 45% of all species listed as endangered or threatened reside within coastal habitats.

Setback Line Methods: Fixed Setbacks

“Fixed” setback lines are rigid in the sense that they do not depend on *a priori* measurements, but rather on a uniform, standard, or fixed distance landward of a reference feature. In this case, setback distances are determined prior to a permit application (Houlahan 1989). For example, Delaware places setback lines 100 f. (30.5 m) landward from the seawardmost 7 f. (2.1 m) elevation above the national geodetic vertical datum (NGVD). In the Technical Belt of Poland, the Maritime Administration employs a setback of 200 m landward of the dune ridge or 100 m inland of the upper edge of a cliff. However, a case study conducted in Essex, England claimed that, along reaches of varied topography, a rigid setback line would have no topographical or planning relevance since the artificial boundary can bisect natural ecosystem and hydrologic boundaries (Bridge and Salman 2000). Additionally, Klee (1999) considers fixed setbacks a “weak” approach because coastal construction can occur at existing setbacks regardless of current potential hazards.

Setback Line Methods: Floating Setbacks

“Floating” or variable-rate setbacks use a dynamic (nonrigid) natural phenomenon to determine setback lines and can change according to the topography of an area or by measurements of shoreline movement over time. Thus, floating setbacks reflect site- or reach-specific coastal processes and, in theory, allow for natural erosion to occur. Floating setback distances usually are established after a permit request (Houlahan 1989). In most applications of variable-rate setbacks, setback lines typically are positioned in a landward direction of the reference feature by multiplying an “average annual erosion rate” (AAER in feet or meters per year) by a constant (time in years of setback protection). For example, the setbacks for single-family dwellings (defined as buildings with a floor area <5,000 ft² or 464.5 m²) in North Carolina are determined by multiplying the AAER by 30 with a minimum 60 ft (18.3 m) setback. The constants and minimum setbacks can vary with building size, anticipated life-expectancy of a structure, or duration of a mortgage.

The setback distance, obtained by the variable-rate AAER method, is based on a prediction of a future shoreline position. This method for producing a setback requires accurate assessments of historical shoreline changes, usually in the form of a shoreline rate-of-change statistic. In some cases, erosion rate data are averaged or “smoothed” along the length of a coast and/or grouped into blocks of similar rate values to produce a representative rate for a segment of coast. When using a dynamic setback approach, shoreline change data must be updated periodically to reflect current and historical

conditions. However, this approach may not explicitly account for extreme events such as storm surge.

Setback Line Methods: Combined Fixed and Floating Setbacks

Some US states require a combination of the two setback types to increase the setback a fixed distance landward of the distance given by the erosion rate. This strategy is used along the U.S.' Great Lakes (e.g., Michigan) to protect a cliff face and along some US Atlantic coastal states such as New Jersey and North Carolina to safeguard areas prone to severe erosion.

Methods Used to Establish "Erosion Rate-Based" Setbacks

Floating setback distances (in meters or feet) are obtained from extrapolating historical shoreline or cliff-line migration erosion rate or trend lines (in meters per year or feet per year) to a predetermined target date (in years). The goal of extrapolation is to provide the best estimate of a future shoreline or cliff-line position. In theory, the calculated setback distance reflects the actual distance the shoreline will migrate landward, year by year, over a number of years. The precision with which estimates of shoreline rates-of-change reflect actual changes and predict future changes depends on:

- The accuracy of the techniques used to identify and record historical shoreline positions (e.g., shoreline identification and rectification on historical maps, vertical aerial photographs, or Geographic Positioning Systems, GPS),
- The quality of the data base (e.g., map and photograph quality, number of maps and photos of shoreline positions available and used in the computation, the time period between measurements, the total time span of the record),
- The mathematical and statistical method used to calculate an historical shoreline rate-of-change – especially when faced with nonlinear, cyclic, or chaotic shoreline migration trends,
- The temporal variability of shoreline movement,
- The proximity of the observations to actual changes in the trend (sampling bias),
- The quantitative techniques used to predict a future shoreline position based on historical trends, and
- An analysis and understanding of the coastal processes and human-induced shoreline modifications that govern shoreline migration trends.

The data used to determine erosion rate-based setbacks consist of a number of shoreline positions obtained from historical maps, maps compiled from aerial photography, aerial photographs and GPS. Analyses of historical shoreline

movement trends begin by overlaying a set of georeferenced maps and/or rectified aerial orthophotographs for a coastal area. Next, transects or monuments of known geographic coordinates are established perpendicular to a baseline and, preferentially, to the shoreline (i.e., estimate error is minimized as the baseline and shoreline become more parallel). At each transect or monument, a data set is produced of shoreline positions at specific dates from which trends or shoreline rates of change can be delineated.

Shoreline changes occur over a wide range of nested time scales. At one extreme, shorelines move in response to eustatic sea-level fluctuations and tectonism; at the other, changes result from the constant fluctuations of wind, wave, and tidal action. The shoreline position at any given time (and the ensuing direction and magnitude of shoreline movement) is a cumulative summary of all long- and short-term processes. However, aerial orthophotographs record instantaneous positions of the shoreline at a specific time, and maps often record shorelines surveyed from instantaneous to longer, but unknown time periods. Consequently, each recorded shoreline time/position data point possesses a degree of gross, systematic, or random error arising from attempts to locate precisely the shoreline datum (typically the HWL) from photographs and maps. This situation leads to erosion rates that contain various degrees of uncertainty. Historical shoreline rate-of-change statistics are, therefore, relatively long-term, linear summaries of several instantaneous, error-prone geographical shoreline positions. Dolan et al. (1991) discussed four methods used to calculate shoreline rates of change and showed how the potential sources of error can bias the final statistics.

Erosion-rate based setbacks are obtained by extrapolating the slope of a trend line (a rate-of-change statistic) which passes through a few (usually <10) historical shoreline position/time data points. In most cases, when setback distances are predicted from a line calculated by a linear regression, the y-intercept is omitted, the slope (rate-of-change) is retained, and the line is adjusted to pass through the most recent data point (not needed for predictions based on an endpoint rate).

The accuracy of a shoreline prediction depends on the accuracy of both the data and the historical trend estimates based on those data (Fenster et al. 1993; Douglas and Crowell 2000; Galgano and Douglas 2000). The challenge of establishing accurate setback lines comes from (1) using a limited number of error-prone shoreline position data to delineate actual historical trends and (2) knowing when to rely on a 30 year or more projection of a shoreline positions based on an historical *linear* erosion rate estimate. A rate-of-change statistic computed from shoreline position/time data sets implicitly assumes that shoreline movement is constant and uniform (i.e., linear) through time – often not the case in reality. In fact, shoreline movement can be linear, nonlinear, cyclic, or (perhaps) chaotic (Eliot and Clarke 1989; Fenster

et al. 1993). The popularity of linear extrapolation for setback determination is due chiefly to its simplicity. As with any empirical technique, no knowledge of or theory regarding the sand transport system is required. Instead, the cumulative effect of all the underlying processes is assumed to be captured in the position history. Thus, an assumption implicit in the procedure for delineating erosion-rate setback lines is that the observed historical rate-of-change is the best estimate available for predicting the future. Regardless of the potential limitations of this assumption, the use of linear models to predict the future avoids problems (such as overfit, underfit, and/or accelerations) that arise when using nonlinear models to extrapolate. Thus, it is best to constrain projections of shoreline position to a line or series of lines.

The NRC (1990), Morton (1991), and many others have discussed the highly nonlinear nature of processes that transport sediment in the coastal zone. For shorelines that experience short-term fluctuations in shoreline migration or longer-term migration nonlinearity, linear estimation methods will vary in their ability to approximate the long-term historical trend. Fenster et al. (1993) developed and demonstrated a method to employ nonlinear models, when called for by statistical significance tests of historical shoreline trend data, as an intermediate step to find the best line, or range of lines, with which to estimate future shoreline positions. Moreover, they investigated techniques for extrapolating shoreline behavior and discuss what constitutes a significant change in shoreline trend (especially in the absence of structures or processes known to have altered the underlying system). The goal of this shoreline prediction technique was to allow assessment of the stability of a long-term trend relative to intermediate (the 50 years of aerial photography) and short-term (decennial) trends, thereby best relating past shorelines to those which can be expected in the future. In this way, *linear* predictions using a pair of lines representing the range of probable future shoreline positions (and the degree of prediction uncertainty) can be based on linear *or* curvilinear historical shoreline migration trends. This approach, used in concert with knowledge of the process-response system (if known), was designed to avoid problems associated with using nonlinear models for predictions and to provide a probabilistic basis for predicting future shoreline locations. Crowell et al. (1997) conducted a thorough review of this technique and found, using average annual sea-level data as a surrogate for a few instantaneous shoreline positions, that a simple linear regression surpasses other forecasting methods unless known physical changes have occurred to the system. Douglas and Crowell (2000) and Galgano and Douglas (2000) have examined errors associated with common shoreline forecasting methods.

Although process information for a coastal reach is critical to understanding long-term geomorphological changes, often these data are not available. In addition, the relative

contribution of all the process variables that contribute to shoreline erosion or accretion may be difficult to determine. For these reasons, coastal management strategies have relied mostly on extrapolating shoreline migration data that use time as a surrogate for the processes that produce shoreline change. Other deterministic shoreline modeling routines lack universal reliability (e.g., Theiler et al. 2000). Research efforts should strive to produce reliable process-based models that can predict shoreline erosion from the synergistic processes that produce changes to the sediment budget.

Several attempts have been made to identify and isolate the processes responsible for producing shoreline changes (the signal) and deviations from the signal (noise). For example, Leatherman et al. (2000) asserted that global eustatic sea-level rise drives coastal erosion and that a relationship exists between long-term average rates of shoreline erosion and the rate of sea-level rise (i.e., the shoreline retreat is 150 times the rate of sea-level rise). However, arguments posed by Pilkey et al. (2000), Sallenger et al. (2000), and Galvin (2000) that question this relationship underscore the difficulty involved in isolating specific oceanographic processes with their coastal (i.e., shoreline) responses.

The fundamental difficulties involved in relating hydrodynamics to sediment transport are further illustrated by recent research dealing with the relationship between tropical and extratropical storms and the long-term shoreline migration history of a coastal reach. Douglas and Crowell (2000) and Honeycutt et al. (2001) contend that, despite the fact that large storms can cause enough beach erosion to require a decade or more for recovery, it is appropriate to eliminate storms from shoreline change data bases. On the other hand, based on analyses of a reach along the wave-dominated Outer Banks of North Carolina, Fenster et al. (2001) concluded that the exclusion of storm-influenced data points is neither warranted nor prudent because such values do not constitute outliers, and they do not increase substantially the range of uncertainty surrounding predicted future shoreline positions. In addition, the added value of reducing uncertainty with the inclusion of more data points outweighs the potential advantages of excluding storm-influenced or storm-dominated data points.

Spatial Considerations

The spatial considerations involved in establishing erosion-rate based setbacks stem from the need to project the setback line some landward distance from a continuous reference line. Therefore, extrapolations must occur at some predetermined spacing (transects) along a shore-parallel baseline that contains an infinite number of points. This situation raises a question, namely, where along a shore-parallel baseline, that is, at what spacings, should extrapolations occur? Although Geographic Information Systems (GIS) and high-speed digitizing techniques allow investigators to assemble almost unlimited samples for assessments of along-the-coast patterns

of variation, there is value in knowing the relationship between sample size (e.g., number of digitized points) and the accuracy of an average rate-of-change estimate for a particular reach or an unspecified location along-the-shore. Rate-of-change values will be more useful for erosion-rate based setback delineation and are less likely to be contested when obtained by a scientifically and statistically sound approach. Although many coastal management agencies use arbitrary transect spacings (e.g., 50 m, 100 m), Dolan et al. (1992) used geostatistics (i.e., the theory of regionalized variables) to provide nomograms by which optimal sample sizes can be determined. In this case, optimal sample size is defined as the amount of data required to support the inferences of a particular study. Too few data produce inconclusive results; oversampling is inefficient. This study showed, in part, that the confidence in estimation of rate-of-change values in the spatial domain (due to spatial continuity) far exceeds the confidence in rate values calculated at a sample location in the temporal domain because along-the-shore rate-of-change values exhibit a high degree of spatial autocorrelation (nearest neighbor effect). In fact, the 50 m spacing of transects along Hatteras Island, North Carolina provides an excellent estimate of rates between transects (accurate to ± 0.5 m/yr).

Administration and Setback Implementation

In addition to differences in the type of setback used by a particular agency, variations exist among nations and states with respect to how setbacks are administered and who administers them. Additionally, the technical standards and methods for establishing those standards vary widely. For example, some US states apply setbacks only to new construction, while others apply to both new and old development. Also, some states differentiate among classes of buildings (e.g., single-family dwelling, multi-family structures, and industrial/commercial buildings). For example, North Carolina uses data averaging of similar shoreline recession rates at neighboring transects to produce a single shoreline recession rate for a “block” of shoreline. On the other hand, Massachusetts calculates recession rates every 100 m along the shore.

Twenty-three of the twenty-nine US coastal states and territories (including American Samoa, Guam, Marianas Islands, Puerto Rico, and Virgin Islands) utilize regulatory setbacks as a coastal zone management tool (Bernd-Cohen and Gordon 1998). According to the Heinz Center Report (2000), ten US states use fixed methods (43.5%), five use floating methods (21.8%), and four use a combination of the two (17.4%). All five US territories used fixed setback methods. Most of the US coastal states use state-controlled setbacks, although two states exclusively

use local setbacks (California and Washington) and five states have no setbacks (Heinz Center 2000). In most cases, states with setback programs use local government agencies to administer the program. These setback programs can be administered on either a mandatory or a voluntary basis.

In Europe, the use of setback policies is increasing in erosion-prone coastal nations. In a study of national policy and legislation in selected European Union and Baltic States, Bridge and Salman (2000) discuss the use of setback lines in the management of coastal zones for nine selected nations and list three types of setback methods in practice within Europe: (1) shore-parallel linear (fixed); (2) contour (a distance from a shore reference feature based on elevation); and (3) Exclusive Economic Zones (offshore protection). Of the nine listed nations, eight have established fixed setbacks: Denmark, Finland, Germany, Norway, Poland, Spain, Sweden, and Turkey. The buffer zone varies widely from 5 m to 3 km. England stands as the exception with no formal coastal setback policy in place. As in the United States, the types and implementation of setbacks within the European Union and Baltic nations vary. Among the cities and regions located along the Russian Baltic Sea coast, a lone local statutory regulation establishes a 1 km wide setback for the economically and militarily strategic Kaliningrad region. No other setback policy is in place within Russia's coastal zone.

Advantages and Disadvantages

The advantages of using setbacks as a regulatory method include:

- Prevent structural loss and damage from erosion
- Protect coastal habitat and water quality
- Provide open space for natural shoreline environment
- Provide recreation and beach access
- Allow natural erosion/accretion cycles to occur
- Can contribute to sustainable management of coastal systems.

The disadvantages are:

- May not provide adequate protection, for example, fixed methods may not account for topography or variations in coastal erosion processes, buffer zone may not mitigate impacts
- May limit tax base (economically restrictive)
- Erosion-rate based data have accuracy limitations
- May not address existing structures
- A strategy is needed to deal with structures located near “old” long-term predictions

- Shore-parallel, linear setback lines do not include the marine zone
- Enforcement may depend on cultural attitudes and administrative context.

Cross-References

- [Caribbean Islands, Coastal Ecology and Geomorphology](#)
- [Changing Sea Levels](#)
- [Coastal Management Practices](#)
- [Coasts, Coastlines, Shores, and Shorelines](#)
- [Erosion Processes](#)
- [Managed Retreat](#)
- [Meteorological Effects on Coasts](#)
- [Natural Hazards](#)
- [Sea-Level and Climate Change](#)
- [Shoreline and Coastal Terrain Mapping](#)
- [Storm Surge](#)

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Sharm Coasts

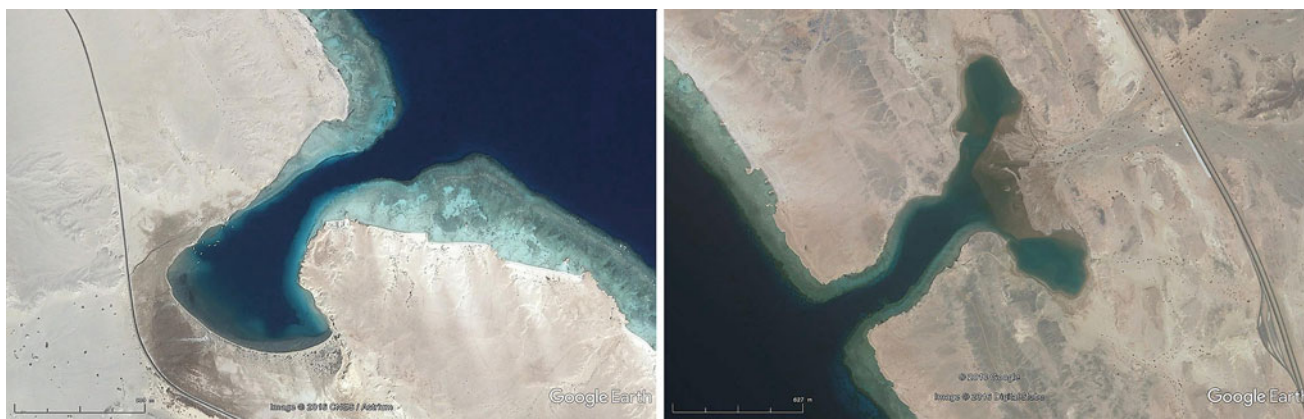
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Definition

Sharms are a type of irregular but mostly broad indentations with a limited extension inland, found along the coastlines of the Red Sea in Egypt (Sharm el Sheikh) or in the middle part of the Saudi Arabian shoreline (El-Abd Y, Awad M 1992a, b; Schmidt 1923).

Sharms do interrupt coral reef belts and sometimes exhibit small reef benches (Fig. 1). The inner parts of the sharms may show beaches (often with beachrocks) at the end of dry valleys (Wadi). Sharms also exist without a connection to valleylike forms inland. There are several interpretations for the formation of sharm embayments, and probably several factors are responsible: lack or interruption of coral reefs by sediment discharge and suspension at the mouth of Wadis or too much freshwater in the form of groundwater seepage. Other explanations contain cutting of a former reef body by fluvial erosion during sea-level low stands in glacial times or over-deepening of the seabed in front of valley mouths where coral reef growth could not come up with the rising sea level in postglacial times (Behairy 1982). Some of the sharms are simply ria-like forms, i.e., drowned lower parts of former valleys, others formed by slumping or sliding of over-steepened fringing



Sharm Coasts, Fig. 1 Sharms at the Red Sea coast. *Left*: from Egypt (1.3 km long), no valley to inland; *right*: from Saudi Arabia, 1.5 km long, with a Wadi outlet and a small delta (Credit: Google Earth)

reef bodies on submarine slopes, but no evidence for this process has ever been checked out in the Red Sea.

Cross-References

- [Coral Reef Coasts](#)
- [Desert Coasts](#)

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this is largely true, the specific modes of origin are often complex. The processes of marine sedimentation and erosion are episodic and strongly related to the water depth. Seaward of the zone of breaking waves, these processes are generally related to storm activity. As a consequence, the timescale of change is long compared to the timescale at which significant changes in sea level are known to occur. In many areas, the timescale of tectonic change is also similar. This means that the continental shelves of the world have been alternately submerged and exposed throughout the numerous sea level excursions of the Pleistocene and earlier geologic epochs. The overall topographic characteristics and local relief features of most continental shelf areas result from a combination of subaerial and submarine processes.

The large-scale geologic setting and the long-term history determine the overall configuration of the shelf. The processes of the erosion and sedimentation can be effective in modifying features caused by tectonic warping or faulting. Depending upon the intensity of the wave-current climate, range of water depths, and degree of consolidation, the shelf processes can be effective in forming both broad scale and local relief.

Shelf Processes

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Introduction

The low slopes and gentle relief of most continental shelves immediately suggest that they were formed through some combination of marine erosion and sedimentation. Although

The Hydrodynamic Regime

A discussion of shelf processes logically begins with a description of the behavior of waves and currents, with special attention to how they bring about marine sediment transport. The actual processes of marine sediment transport are complex, and a detailed discussion is beyond the scope of this entry. Briefly, we parameterize the entrainment of bottom sediment by the fluid shear stress acting on a unit of bed area. This shear stress is proportional to the vertical gradient of fluid velocity acting immediately above the seabed. Thus, both the magnitude of the flow and its vertical structure

contribute to its ability to entrain bottom sediment. Shelf currents arise from various combinations of tide, surface wind stress, and internal pressure gradients. Normally, the current bottom boundary layer is meters to tens of meters in thickness. Waves, on the other hand, have boundary layers that are generally less than 10 cm in thickness because the flow reverses during each wave cycle. The wave boundary layer is thin and it readily produces large shear stress acting on the seabed.

There are many factors that distinguish the marine environment from other environments with active sediment transport. Unlike rivers, the magnitude and direction of currents on the shelf are constantly changing. At the same time, the size and period of the waves are also varying. The orbital motion of the waves decreases exponentially with water depth and becomes insignificant at depths equal to half the deepwater wavelength. Inside of this depth, both waves and currents contribute to the shear stress acting on the bottom sediments. In many cases, especially during storms, the major stress results from the waves. A common condition is that the waves are primarily responsible for entraining the bottom sediment, and the currents determine the magnitude and direction of the sediment transport. There is no distinct depth limit for shelf sediment transport due to waves and currents. Instead, the frequency at which storm events are powerful enough to cause sediment entrainment simply diminishes with increasing water depth across the whole width of most shelves.

When waves and currents occur together there is an interaction because turbulence is generated by the intense shear in the wave boundary layer which interacts with the turbulence in the bottom boundary layer of the current. The effect is that the current experiences greater drag than would occur without the waves. Entrainment and transport of bottom sediment are also enhanced.

Marine Sediment Dynamics

Several factors related to the condition of the seabed influence how sediment is entrained. Local bottom roughness, which may result from the presence of sand ripples, worm burrows, or shell fragments, produce more drag on the current. Part of this enhanced shear is applied to the sediment surface and part results from drag that does not act directly to entrain particles. Thus, identical currents passing over adjacent seafloor segments with different bottom roughness characteristics will apply different magnitudes of bottom shear stress.

The bottom sediments may be granular or cohesive. Granular sediments are stabilized in their “at-rest” positions only by gravity. As a general rule, sediments with less than 10–15% clay particles exhibit granular behavior (Metha

et al. 1989). Sediments with higher concentrations of clay particles, or significant amounts of organic material, tend to resist entrainment because of binding effects.

Empirical relationships have been developed to relate the entrainment of granular bottom sediment to the fluid shear stress acting on the bed. At present there is no method to predict the onset of entrainment of cohesive sediments without direct measurements, either in the field or in specially equipped laboratory flumes.

Once the waves, currents, or a combination of both entrains the bottom sediment, it is subject to transport and deposition. As in rivers, the sediment is transported as suspended load, bedload, and dissolved load. But because waves play an important role in marine sediment dynamics, and because current conditions change rapidly, the sediment transport processes are noticeably different in shelf environments. Pure bedload transport is restricted to medium and coarser sand. These particles settle rapidly and are rarely a significant fraction of the suspended load outside of the nearshore zone.

There is no physical limit to the water depth at which bedload transport occurs. There are some places where bedload sand transport happens in mid-shelf and outer-shelf environments. However, the necessary high fluid stress rarely occur in the ordinary conditions in shelf environments. Only in certain places are tidal or wind-driven currents fast enough to cause this form of sediment transport.

Suspended sediment transport is the dominant mode in most continental shelf environments. Under most conditions, active deposits of medium and coarser sand are restricted to the beach, surf zone, and shoreface environments (Dean 2001). Mud, silt, and fine sand sediment, entrained from the seabed, are dispersed upward into the shelf water column by turbulent eddies in the boundary layer of the currents and waves. In a stable transport condition, the turbulence and the vertical profile of the concentration of suspended sediment particles act to cause an upward flux that balances the settling speed of the suspended sediment. If the flow intensifies, more sediment will be entrained and both the concentration and height of the suspended layer will increase. The reverse of these processes cause deposition.

The presence of suspended particles alters the structure of the bottom boundary layer. The vertical concentration gradient acts to suppress the turbulence which, in turn, reduces the stress applied to the seafloor. The presence of a vertical density gradient, brought about by the vertical gradient in the concentration of suspended particles, can severely limit the thickness of the suspended layer. Both numerical model results and scale model observations have shown that the boundary layer of large storm waves can become saturated with suspended sediment. A sharp density gradient develops at the top of the boundary layer and prevents the sediment from escaping upward into the mean flow.

In most shelf environments suspended sediments are repeatedly entrained, transported, and redeposited until they are finally buried and protected (Niedoroda et al. 1989). Cohesive deposits generally undergo an aging process after their burial that strongly affects their susceptibility to reentrainment. New deposits have high water contents and are weak. Over periods of days to months these sediments dewater and are bioturbated. Both of these effects make these sediments more resistive to reentrainment.

The intensity of boundary layer turbulence is also significantly reduced where the seafloor is covered by a smooth mud layer. The absence of boundary roughness allows a laminar boundary layer to develop to a significant thickness. This tends to suppress the shear stress transmitted to the bottom sediment. The entrainment of the sediment is thus suppressed.

Shelf Sediment Transport Processes

The combined effects of waves and currents dominate shelf sediment transport. In most places, currents alone are inadequate to entrain the bottom sediment. Thus, under calm conditions there is little to no sediment transport. Shelf sediment transport is mainly accomplished during storm events. The magnitude of the transport is a nonlinear function of the intensity of the bottom currents and size of the waves. The magnitude of sediment transport rises sharply as these two forcing parameters increase. A major storm can be responsible for more sediment transport than occurs in many smaller storms. Because patterns of waves and currents do not repeat themselves exactly in a succession, the net transport of sediment over long time intervals results from the sum of individual displacements during each individual storm event. This makes the dispersion of sediment particles in most shelf environments quite complex.

There are cases where sediment is entrained and transported primarily by shelf currents. The shear stresses that are imparted to the bottom sediments on the open shelf by hurricane-driven currents can be large enough to affect significantly entrained sediment (Murray 1970). In some locations around the British Islands strong tidal currents routinely cause sediment transport (McCave 1972). In other places, such as the shelf off southeast Florida, a strong offshore current provides sufficient stress to continuously transport sediment.

There continues to be some dispute about whether density-driven bottom currents, similar to turbidity currents, occur on continental shelves. It has been proposed by Hayes (1963), that as a storm-surge return-flow floods back over barrier islands, it becomes sufficiently sediment charged to persist as a density-driven flow as far as the mid-shelf region. The Texas shelf, where these storm deposits have been observed, has been re-examined by Snedden and Nummedal (1989).

They found that the beds associated with Hurricane Carla are better explained as resulting from a more normal downwelling flow in the latter portion of the storm.

A form of gravity-driven sediment-laden bottom flow has recently been observed. Using a combination of direct measurements from an instrumented tripod and acoustic images, it was possible to show that fine sediment was entrained within the wave boundary layer during major storms on the Eel River shelf. The sediment was suspended by the intense shear in the wave boundary layer at a concentration that inhibited vertical mixing across the top of the boundary layer. Gravity, acting down the bottom slope, propelled the sediment suspension across the shelf to near the shelf edge. This is a wave-entrained bottom flow rather than a classic turbidity current because the source of intense turbulence is the waves rather than the mean flow.

Mass gravity flows in the form of mudflows and related turbidity currents are known to develop in shelf environments off major river deltas (Coleman and Prior 1982). These can be associated with the venting of natural gas, or as a result of liquefaction due to excitation by surface storm waves. In either case large pockets of mud become unstable and flow downslope (Bea et al. 1983). Mudflows of this type are capable of triggering turbidity currents (Niedoroda et al. 2003b).

Hyperpycnal flows are density currents caused by sediment-laden flows from a river mouth entering still water bodies. These have been observed in reservoirs (Syvitski et al. 1998). These authors show that hyperpycnal flows are rare in continental shelf situations because the density contrast between saltwater and river water causes the river plumes to be buoyant.

Coastal plumes are an important sediment transport and dispersal mechanism on shelves. Depending on the size of the river watershed, the flood discharges may or may not be associated with local storm conditions. In small watersheds the correlation of storm conditions and a river flood plume is stronger. Under these conditions the plume is often driven along the shore by the storm winds, and temporary deposits form beneath it. Subsequent storm events that are not associated with river floods then shift the temporary deposits from the inner-shelf to the outer-shelf. This process has been well documented on the Eel River shelf (Wheatcroft et al. 1999).

Sediment-rich flood plumes from large river systems often arrive at the river mouth during ordinary weather conditions. Because of the effects of the earth's rotation these discharges tend to turn cum sole (to the right in the Northern Hemisphere) and become dynamically coupled to the coastline. These plumes of brackish water with high concentrations of suspended sediments commonly persist for hundreds of kilometers along the Louisiana-Texas coast, the coast of the central United States, the south coast of the northern and many other places.

Temporary mud patches have been observed in the mid-Atlantic Bight off New Jersey and Long Island. These form during the calmer summer months on the mid- and inner-shelf. It is thought that under prolonged calm conditions an initial mud deposit begins to collect in a sheltered deeper area within the ridge and swale bottom topography. The presence of even a thin layer of mud makes the bottom hydrodynamically smooth. This inhibits the production of turbulence in the bottom boundary layer and promotes additional settling of suspended particles. The initial mud patch can thus grow. These persist until the stormy winter season.

Thicker, more persistent and extensive deposits of soft mud have been observed on the inner- and mid-shelf off Louisiana and east Texas (Suhayda 1977). The water content of the mud is high so that it develops wave motions when forced by storm waves in the overlying water column. The wave energy is dissipated in the moving mud to the point that shelf areas behind these mud deposits are sheltered from the storm waves. Local fishermen are known to utilize these refuges in storms.

At high-latitudes ice is an agent of shelf sediment transport. The Canadian Atlantic shelf, the shelves of some of the large islands in the High Canadian Arctic and the shelf around Antarctica (Anderson 1989) are places where a significant amount of ice-rafted sediment is deposited. This sediment comes from the melting of icebergs that have calved from glaciers. Stranded ice produces one of the oddest modes of shelf sediment transport that is important only very locally. Ice keels, either from icebergs or from the undersides of sea ice ridges, rake through the shelf sediments and produce distinctive gouges (scours) that can be several meters deep, tens of meters wide, and kilometers long (Niedoroda 1991; Weeks et al. 1991).

Shelf Sedimentary Sources

Sediment that is transported on the shelf originates either from land sources or from the reworking of previous deposits. Shelves in tectonically active areas tend to derive most of their sediment from the land. Shelves in stable tectonic areas are generally sediment-starved because rising sea level has caused the river sediments to be trapped within large coastal plain estuaries (Swift et al. 1991).

On most coasts the coastline provides a line source of sediments to the adjoining continental shelf. The sediment is moved alongshore in the surf zone, in coastal plumes and in the accelerated flows of the upper shoreface. Coastal engineers have adopted the concept of a depth-of-closure (Hallermeier 1978) for use in planning beach nourishment programs. This concept is based on the observation that the changes of beach and nearshore profiles are constrained within an envelope. The upper and lower envelope limits

converge towards each other at the seaward extent suggesting that all of the sand simply exchanges between the beach and offshore bars without net loss. Actually, there generally is a net loss, but it is small compared with the large profile changes brought about by the cycle of storm flattening and recovery during quiet periods. Along coastlines undergoing net erosion this small loss to the upper shoreface provides a line source of sediment to the shelf.

Many of the discrete coastal sediment sources (rivers, tidal inlets, estuaries) have been discussed above. There is often a perturbation in the shelf sedimentary regime adjacent to these sources in the form of shoals, tidal sand waves or shifting deposits of sediment. In most cases, shelf processes spread the sediment delivered from these sources. Because shelf flows forced by the tide and wind are dominantly shore-parallel within a few kilometers of the coastline, this dispersal of sediment is also dominantly in the coast-parallel direction.

Much of the sediment on broad, sandy continental shelves results from the reworking of preexisting deposits. Where coastal systems are retreating in the face of rising sea level the mid-shoreface exposes older strata (Swift et al. 1991). As the coastal system migrates landward these exposed strata become buried by lower shoreface deposits. This is the origin of a transgressive sand sheet that extends across the entire shelf. It is a common feature of broad sandy shelves and is formed by a combination of shoreface retreat and storm the reworking over the period since the last glacial maximum.

Shelf Morphodynamics

Most continental shelves and their major features are composed of unconsolidated sediments. As these are exposed to considerable fluid power, in the form of waves and currents, it is predictable that the morphology of shelf features, and the shelves themselves, represent a timeaveraged morphological adjustment of the climate of waves and currents. Niedoroda and Swift (1981) have shown that, although all unconsolidated shelves have a concave-upwards profile, the steepness of the inner shelf profile and the distribution of depths in the mid-shelf correlate with the general intensity of the wave-current climatology. The profile of shelves in stormy areas have a steeper inner-shelf shelf and overall greater depth.

The shoreface is defined according to the concept that it represents the time-averaged response of the shelf surface to the controlling hydrodynamics acting on a sloped surface (Niedoroda and Swift 1981; Niedoroda et al. 1985a, b). It has been recognized that the shoreface profile translates landward in the face of sea-level rise with little change to its overall shape (Bruun 1962; Cowell et al. 1995). Moody (1964) measured shoreface adjustments to severe storms off the Delaware coast and showed that the profile alternately steepened and flattened about a mean shape. Niedoroda and

Swift (1981) argued that the shape of the shoreface was maintained by a time-averaged balance of sediment transport processes that involved the action of downwell and upwelling storm flows, the corresponding timing of maximum wave conditions during the storm events, the offshore bias caused by the bottom slope and diminished sediment suspension with depth and an onshore bias to sand transport during the period between storms. Measurement programs by Niedoroda et al. (1984), Wright et al. (1994), and have demonstrated the action of these components, but a complete balance of processes has yet to be successfully demonstrated.

The mode of origin and maintenance of some of the major shelf features has recently been studied. Sand waves are very common in the mid-shelf ridge and swale morphology in the North Sea, Canadian and American Atlantic shelf, and the northeast Gulf of Mexico, among other areas. Recent work by Nemeth et al. (2001) has examined these features with stability analyses. A system of equations covering both the shelf hydrodynamics and sediment dynamics has been developed. These are used to show that, under a range of current and wave conditions, a flat shelf surface is not stable. Differential sediment transport tends to produce a “bumpy” surface due to an interaction between the vertical velocity and turbulence profiles and the shape of shelf surface. Over time, only the bottom shape perturbations (bumps) within a narrow range of spacings (wavelengths) will grow. These become the sand waves. The characteristic timescale for producing these features in an energetic mid-shelf location is on the order of centuries to millennia. Once produced, the sand waves persist because they are the stable configuration of the shelf surface to the time-averaged wave-current climate of the area. Nemeth et al. (2001) have shown that, in areas where there is a net bias to the shelf flows the sand waves slowly migrate while maintaining their forms. Without this bias, sand waves can remain fixed in location and form.

Calvete et al. (2001) have used a similar mathematical method to explain the existence of shoreface connected sand waves. Their analyses show that these features are the stable response mode of the seafloor in the lower shoreface region to the wave–current climatology weighted according to the scale of the induce sediment transport events. The timescale for developing these features is in the range of centuries to millennia. They can be stable or they can migrate longshore (usually in the direction of the apex of the angle their axes make with the shoreline) and shoreward as the shoreface retreats.

The time-averaged morphodynamics of the shelf and shoreface profiles have been represented in numerical models (Niedoroda et al. 1995, 2001, 2003a; Stive and deVriend 1995; Zhang et al. 1997, 1999; Carey et al. 1999). In these studies the physics of shelf sediment transport are represented in time- and space-averaged forms so that individual events, such as storms, are not resolved. Both depth-averaged cross-

shelf (1D-V) and plan-view (2D-H) representations have been developed. These models are in part based on the concept proposed by Clarke et al. (1983). This states that provided the characteristic sediment particle transport excursion distance in individual events (typically storms) is small relative to the scale of the shelf, then the time-averaged sediment transport can be represented as a Fickian process similar to turbulent diffusion, but at very large time- and length-scales. Thus the effects of individual storm flows, tidal currents, currents due to shelf waves and internal waves, and other fluctuating currents are represented by time-averaged diffusion coefficients, and all results are limited to terms that are long compared with the timescale of characteristic shelf sediment transport events.

These models have been able to reproduce the dependence of the shelf profiles on the major parameters that include the storm climate, sediment supply, and rate of sea-level rise (or fall). The characteristic responses to changes of these parameters are shown to vary from hours in the surf zone to millennia in the mid- to outer-shelf zones. The controlling parameters are the same as the “Sloss parameters” identified by Swift et al. (1991) based on more empirical observations.

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Cross-References

- [Changing Sea Levels](#)
- [Coastal Currents](#)
- [Continental Shelves](#)
- [Cross-Shore Sediment Transport](#)
- [Deltas](#)
- [Depth of Closure on Sandy Coasts](#)
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- [Offshore Sand Sheets](#)
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- [Physical Models](#)
- [Sequence Stratigraphy](#)
- [Shoreface](#)
- [Wave–Current Interaction](#)

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Shell Middens

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Shell midden refers to anthropogenic deposits containing noticeable amounts of shell, that is, calcareous invertebrate tests. Such deposits are common in marine coastal areas from subarctic to tropical latitudes throughout the world. They also occur along rivers and lakes, where freshwater molluscs comprise the shell constituents.

Shell midden is not an analytically rigorous term in the sense of human activity, but a descriptive label that identifies the most superficially recognizable constituent of the deposit. *Shell-bearing site* is a more accurate label, although cumbersome and not likely to be adopted (Claasen 1991). *Midden* originally meant domestic refuse deposited around a house, but shell-bearing sites in the label further obscures the importance of other constituents of the deposit. These sites do not solely represent shell-gathering activities by people. Other food remains such as fish or plants may be very abundant and may, in fact, represent a more significant economic activity at that location, but such remains generally require screening or microscopic analysis of samples to identify.

One sense in which shell middens or shell-bearing sites are a meaningful category is in terms of the depositional and preservational environment represented. As deposits, shell middens share the attributes of having increased porosity, permeability, and alkalinity (Stein 1992). The alkaline environment created by the leaching of calcium carbonate from the shell and the neutralizing of groundwater, which is generally weakly acidic, creates excellent preservation for organic materials such as shell, bone, and antler. Also, because shell middens are located near aquatic habitats, they frequently are affected by saturation by water (Stein 1992).

History of Research

Investigations of shell middens in Europe, North America, and other continents began in the mid to late 1800s by naturalists, but once they were recognized as anthropogenic, became largely the domain of archaeologists in the twentieth century (Stein 1992). In the early part of the twentieth century, it was not uncommon for shell middens to be excavated simply to recover artifacts for determining culture historical sequences, essentially disregarding the shell constituents. Most research, however, has focused on the unique aspects of shell middens, their rich invertebrate content, their relationship to coastal resources, and the unique depositional and postdepositional processes that affect them. By far the greatest amount of work has been expended on reconstructions of subsistence and diet, and relating this to the development of marine adaptations among human groups. Closely related is methodological research concerning appropriate sampling and quantification strategies for characterizing midden contents and methods for determining seasonality of deposits. Interest in how shell midden deposits were formed, in terms of the human behavioral components, began with the earliest observations, but studying formation and diagenesis from a geoarchaeological point of view is a more recent development. Middens have always been viewed as informing on ancient environments and habitats, from the position of the coast to the nature of the neritic environment,

but the integration with studies of coastal ecology, geomorphology, and neotectonism has increased dramatically in recent years. Studies of middens have been applied in legal cases concerning the traditional subsistence practices and territories of native peoples.

Behavioral Component of Shell Deposition

A century of research at shell-bearing sites, combined with ethnohistoric and ethnoarchaeological studies around the world (Wessen 1982; Trigger 1986; Waselkov and Shiffer 1987; Claasen 1991) has resulted in the recognition of a great deal of variability in the cultural activity that created shell-bearing sites. Layers with dense shell may be deposited in association with major village settlements, annually re-used seasonal encampments, temporary wayfaring encampments, or specialized nonhabitation resource procurement sites. Shell middens commonly have more than 20 different taxa of marine invertebrates, each of which is deposited as the result of a distinct chain of activities. Multiple harvesting methods for multiple taxa might be employed at a specific location, and several different marine habitats might be exploited from a single base camp on foot or by watercraft. Shellfish might have been collected regularly, only seasonally, or incidentally to the procurement of other resources, and again, this is different for each taxa represented. They may have been consumed immediately or preserved for storage or trade, or used as bait rather than as food (Claasen 1991). Processing ranged from eating shellfish raw to a variety of methods of roasting, steaming, boiling, drying, and smoking. These activities leave a wide variety of thermal features from pits to surface pavements of stone used as heating elements.

Shell refuse was disposed of in various ways. It was sometimes burned, and might be discarded adjacent to the habitation area, into the intertidal zone, or over a riverbank. In a short-term occupation, living areas and refuse discard areas might be discreet, but when a location was re-used over hundreds or thousands of years, these areas might shift, so that houses and other structures were built into former refuse areas or refuse disposed of in former house areas. In a shore setting the midden accumulation itself might make a significant contribution to accretion and become the most desirable place for habitation as it built up above the water level. Midden areas were sometimes later used as cemeteries, which adds to the impetus for protecting these resources against erosion and development.

Industrial, in contrast to domestic, uses of shell also create or contributed to shell accumulations. Marine shells have been widely used for making ornaments, such as beads and pendants, or as an inlay materials. They have also been used as musical instruments, for example, Aztec conch shell horns, and as a trading standards (e.g., wampum). Use of shell to

manufacture tools is less common, but includes the manufacture of mussel shell knives and whaling harpoon heads in northwest North America, fishhooks in the Polynesian areas, and heavy pounders and adzes made from huge *Tridacna* shells in the Pacific. Tools and ornaments of marine shell were made not just for local consumption, but were traded far inland on all continents. Ground shell was used as temper in some prehistoric pottery wares. Shell also was used as a raw materials in construction (Blukis Onat 1985). Crushed, it makes a permeable substrate, and in Mesoamerica was ground to make lime for plaster and cement. In the Pacific Islands, it was common for people to extend their land by building bulkheads, and then fill behind them with a mixture of refuse and materials dredged up from the reef flats (thus aiding in keeping canoe channels clear as well) which included coral and shells. Shell industries that involved mining shell for construction, or processing shells for non-subsistence trade items, likely created distinctive types of shell middens.

Anthropogenic Versus Natural Shell Accumulations

Recognition of the anthropogenic origin of some shell beds began with observations by naturalists in the nineteenth century. Criteria for distinguishing cultural shell accumulations from other fossils shell beds has continued to be an area of interest as archaeologists have continued to encounter regionally specific nonanthropogenic shell deposits and develop additional criteria to aid in field recognition. Locally dense accumulations of shell can result from activities of animals such as birds or otters that repeatedly break and leave shells in localized areas, or through beach processes that sort sediments by size, shaped, and density. Natural death assemblages in subtidal marine deposits are less similar, but can be confusing when these are elevated relative to contemporary sea levels. A further problem is the recognition of cases where cultural middens are redeposited by shore processes or by later cultures (Ceci 1984).

A typical shell midden is likely to have associated artifacts such as tools, ornaments, and manufacturing waste, although these are often quite sparse. Thermally altered rocks, a ubiquitous byproduct of many food processing activities, are often abundant. Other evidence of thermal processes can include ash lenses, charcoal, lenses of oxidized sediment, and calcined shell and bone. The fine matrix in a midden is typically very dark, from charcoal and decomposed organic materials, although there may also be lenses of wood ash, or oxidized soil. The microstratification present in most middens reflects different dumping events as well as constructed features such as surface hearths, roasting or steaming pits, pavements of rock used as heating elements under drying racks, or

postholes from large structures such as houses or small structures such as drying racks.

The faunal assemblages in most prehistoric shell middens are diverse, and at the same time, highly selected. The diverse array of marine invertebrates may well represent multiple habitats. Yet not all the locally available invertebrates were collected, and generally a limited size range of individuals. Selectivity is also indicated in the varying compositions of the strata and lenses of the deposit, which are dominated by different associations of taxa representing specific harvesting events. Numerous species of birds, fishes, and mammals can also be present. Flotation to recover carbonized plant remains has rarely been applied, but can reveal a diverse array of wood charcoal and edible plant tissues.

Shells in middens may be whole or fragmented, depending on how they were collected and processed, and depending also on postdepositional alterations such as trampling, or deliberate use for engineering purposes. Fragmentation often follows specific patterns that indicated particular processing methods, such as splitting chiton plates to remove them. Also, the fragmented edges are sharp and angular, not rounded. Attachments of other animals are not found on the interiors of shells, indicating the organisms were alive shortly before deposition. Pieces of weathered shell may occur, but as minor constituents.

Cultural middens can be found in a wide range of geomorphological conditions because their deposition is not causally related to shore processes in the way that beach deposits are. Humans may transport shells to a location elevated above the shore, or toss refuse shells into the subtidal zone. Size varies, from small lenses that resulted from a single episode of harvesting and processing, to sites hundreds of meters in length, and thicknesses of over 10 m.

Of the traits described above, associated artifacts and constructed features are the most conclusive for recognizing in situ cultural midden deposits, but also the least likely to be observable in a field survey situation without extensive exposure. When observations are limited to shell scattered over a surface or a limited vertical exposure in a cut bank, the criteria of matrix color presence of thermally altered rock and charcoal, and diversity of faunal remains are often more useful. The fragmentation, weathering, rounding, and other alternation of the shells is particularly useful in distinguishing different transportation mechanisms, and the length of time elapsed between death of the organism and deposition. Natural death assemblages in marine deposits exhibit the least degree of movement, often including articulated shells with little fragmentation or mechanical damage. They also typically represent a single habitat, and include abundant juveniles. Beach deposits may incorporate some very fresh shells, but generally contain a significant amount of shells that died some time ago and are essentially part of the sediment load. These will be whole or fragmentary shells that are surface

abraded and rounded on the exterior and exposed interior anatomical surfaces, as well as on fractured edges. Older shell may also have been used as substrate for other animals, indicated by attachment of animals such as barnacles, sponges, or bryozoa, on both interior and exterior surfaces. Animals with indications of death by predators, such as drill holes, are more common than in a cultural deposit. The diversity of taxa in a shell midden will be underestimated in field observations (compared with screened samples), but this criteria is usually sufficient to distinguish middens from accumulations of shells created by animals, which tend to focus on a more limited range of species.

Recent shell middens, that is, those formed in the historic period, are potentially quite different from prehistoric ones. Colonization of many areas by people of European descent and changes in traditional diets have led to very different uses of shellfish. Disposal practices are different, too, including incinerating and burying materials in back yards, and use of municipal garbage dumps. The practice of aquaculture, mass production for distant markets, and the use of mechanical equipment results in a different signature characterized by high selectivity, perhaps even a single species. Historic-era middens are most definitively identified by the presence of introduced shellfish species or associated artifacts of recent vintage. In some areas, prehistoric shell middens were mined for shell for use in road construction, for making lime, or as soil amendment, creating extensive secondary deposits. Microstratification and the pressure of constructed features are the best criteria for showing that the deposit is *in situ* and has been redeposited.

Research Issues

Shell middens have played a major role in terms of attempting to understand the evolution of maritime adaptations around the globe, yet the degree to which humans exploited or relied on marine intertidal resources at different points in human development is less well known than for terrestrial resources because changing sea levels have biased the archaeological record, drowning much of the Pleistocene and early Holocene record (Bailey and Parkington 1988). The extensive record at the Klasies River Mouth site in South Africa shows use of marine resources, including molluscs, beginning 60,000 years ago (Thackeray 1988), and a number of shell-bearing sites older than 20,000 years have been documented in Europe, Asia, and Africa (Claassen 1991). Yet most recorded shell middens are Holocene in age. This is interpreted as evidence of increasing resource intensification in the last 10,000 years, correlated with increased population growth and decreasing mobility and territory size during which human groups turned from high-ranked resources such as large herbivores and incorporated more

resources that are low-ranked in terms of energy return such as molluscs and other invertebrates into their subsistence (Bailey 1978, 1983).

Reconstruction of prehistoric subsistence through the quantitative analysis of midden contents has been a major theme since the mid-twentieth century. Much of this research suffers from methodological problems such as inadequate sampling and assuming all shellfish represent food, and estimates of the human population derived from shellfish quantities are particularly flawed (Claassen 1991). Related efforts have focused on determining the seasonality of harvesting shellfish, and by extension, the season of occupation. A number of studies have estimated season of death for species with annual growth rings by determining the proportion of the annual growth that had been achieved, using thin sections or surface measurements of ring width. The validity of the estimates of season of death suffers from reliance on modern samples, and the extension of these seasonality estimates to estimating the seasonality of habitation is weakened by inadequate attention to sampling and assuming that shellfish were harvested for immediate consumption as food, ignoring the possibility of storage and of nonfood uses (Claassen 1991). A focus on the food value of the animal remains in middens continues (Erlandson 1988), but archaeologists are more cautious about generalizing from small samples and make more use of comparative regional information.

Sampling problems are considerable, and not easily resolved. A relatively small excavation may remove, from primary context, hundreds of thousands of pieces of shell, all essentially artifacts. The information context of the shell itself becomes redundant long before the information content of other aspects of the site, such as features, bird bones, tools, is adequately sampled. Because of the high cost of collecting and analyzing shell and other small midden constituents (Koloseike 1970), archaeologists commonly retain limited samples of the excavated materials. Results are frequently noncomparable between projects because of variation in sampling intensities and size fractions. Strategies for selecting the samples spatially are frequently non-probabilistic and not based on the growing understanding of the internal spatial structure of shell middens. For example, the use of a few column samples to characterize the midden, and especially to discuss changes over time, is questionable given research by Campbell (1981) which shows that the shell composition of middens may vary as much horizontally as vertically. In the analysis process, archaeologists variously choose weight, number of identified specimens, and minimum number of individuals to quantify the relative abundance of taxa, adding to the noncomparability of samples. Each measure provides different estimates of abundance, even changing rank order, because of differential fragmentation and recognizability of parts of different taxa.

Geoarchaeological Approaches to Shell Midden Formation

The use of geoarchaeological methods to understand the physical processes involved in the deposition and diagenesis of shell deposits is a much more recent endeavor. Examination of physical traits of midden sediments, such as sediment texture, clast orientation, soil chemistry, and surface weathering on particles is increasingly being used to interpret the physical process of transportation, bioturbation, and leaching (Stein 1992). Diagenetic changes, especially those related to saturation can create apparent stratification that has been misinterpreted by archaeologists interpreting culture history.

Environmental Reconstruction

The role of shell middens in reconstruction of coastal paleoenvironments ranges from simply using them as indicators of paleo-coastlines to inferring nearshore habitats from the molluscan species present, to using them as a means of dating geological deposits that reflect coastal accretion or neotectonic events (Grabert and Larsen 1973). The need for caution and independent evidence to sort out changes due to habitat versus those due to cultural effect is widely recognized. Molluscan assemblages in coastal sites are indicators of coastal ecology and habitat as determined by substrate, exposure, salinity, and water temperature (Rollins et al. 1990) but they also reflect human selection. The most effective work is interdisciplinary and uses multiple lines of independent evidence (e.g., Sanger and Kellogg 1989; Shackelton 1988).

Shell middens are playing a growing role in coastal geomorphological and neotectonic research because they are a specific type of deposit that can be assumed to represent a coastal position, although not causally related to a specific short position as a beach deposit might be. Shell middens contain abundant datable material, including historically diagnostic artifacts (McIntire 1971), radiocarbon datable organic material, and shells datable by isotope analysis (Thackeray 1988) and have been used to provide bracketing dates for subsidence events and tsunami deposits.

Summary

Despite the methodological caveats that plague research on shell middens, they have been shown to reveal a wealth of information on prehistoric adaptations and on coastal paleoenvironments. Interdisciplinary research efforts are increasing and new analytic techniques are being developed. Unfortunately the rate of destruction of these unique coastal

resources is likewise increasing due to development. Although shell middens, like other archaeological sites, are protected by cultural resource protection and management laws in most countries, enforcement is inconsistent or nonexistent. With the increased rate of development added to the natural erosion processes that have always affected coastal sites, we are in danger of losing much of this nonrenewable resource. It is not only a scientific data base for interpreting prehistory and paleoenvironments, but also a historic record and direct physical connection to traditional places and practices for native peoples in many coastal areas.

Cross-References

- [Archaeology and Sea-Level Change](#)
- [Geochronology](#)
- [Human Impact on Coasts](#)

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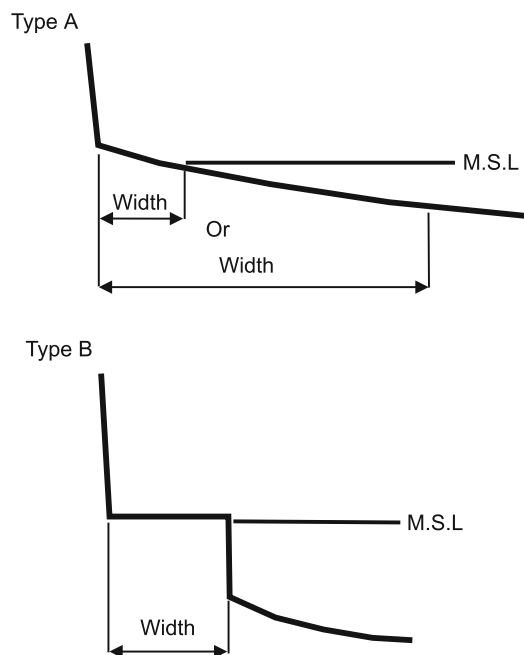
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Shore Platforms

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Shore platforms are horizontal or gently sloping surfaces backed by a cliff, eroded in bedrock at the shore. The erosional origin of these surfaces is evident because they cut across and expose geological structures. Shore platforms have long been classified using a tripartite scheme with respect to elevation in relation to the tide. Thus high-tide, intertidal, and sub-tidal platforms have been identified. Another classification uses the two most common profile forms, either the sloping platform (commonly 1–5°) or the horizontal platform. In recent times these have been referred to as Type A and Type B platforms, respectively (Fig. 1). Shore platforms are common along much of the world's coastline occurring in all but the very highest coastal latitudes, as well as lakes. However, the total percentage of coastline composed of shore platforms is unknown. Associated with shore platforms are a variety of coastal features such as sea caves, ramps, notches, potholes, ramparts, and low-tide cliffs.

Many different terms synonymous with shore platform have been used, often having quite different genetic and morphological meanings. Such terms include: rock bench, high-water rock platform, Old Hat Type platform, abrasion platform, shore bench, storm-wave platform, marine bench, sloping wave-bench, intertidal platform, sea-level shore platform, wave-cut terrace, surf-cut terrace, coastal platform, bench, abrasion bench, denuded bench, tidal bench, wave-cut bench, rock platforms, high-water rock ledges, wave ramp, wave-cut platform, wave-cut terrace, wave-cut shore platform and bedrock platform. Clearly, the wide variety of



Shore Platforms, Fig. 1 Two morphologies of shore platforms, Type A and Type B. (Adapted from Sunamura 1992)

terms used reflects different interpretations of morphology and processes by individual workers, and of the relationship between a surface and sea level. The term “shore platform” has been the most widely used, because it has no genetic connotations and is the most appropriate term since the development of shore platforms and the processes involved are still not fully understood.

The origin of shore platforms has been the subject of debate for 150 years. This debate is concerned with the relative roles of marine and subaerial processes in the development of shore platforms. Advocates of a marine origin have often referred to shore platforms as “wave-cut” platforms. The marine origin for shore platforms has dominated the literature but there is little quantitative evidence to support such a mode of development. The same is also true for subaerial weathering. Little quantitative research of either mechanism has occurred. Investigators have too often relied on interpretative approaches based on morphological features, despite numerous warnings of the ambiguity of morphology as an indicator of process. Subaerial weathering has often been ascribed a secondary role in platform genesis. It has been proposed that the remarkably horizontal nature of some shore platforms results from a secondary planation of the surface by subaerial processes following initial cutting by marine forces. A small group of researchers have suggested that weathering is a primary process. In this scenario, waves serve only to remove the debris formed by weathering, but without the weathering there would be no platform. A third view is that both marine and subaerial processes play strong

roles, but very few workers have suggested both processes are equally important.

Mechanical wave erosion includes impact forces, cavitation, hydrostatic pressure, and compression of air in joints spaces. Air compression results from the sudden inrush of water into a crack or space and causes an explosive shock upon the sudden release of the compressed air. Cavitation occurs when high water velocities at the bed cause a drop in pressure. If vapor pressure is attained then the rock surface can be damaged by the sudden formation and destruction of vapor pockets. Hydrostatic pressures increase with the depth of water. Since depth changes frequently under waves, variations in pressure occur at the bed, but for these variations to cause erosion the force exerted must exceed the strength of the rock. It remains to be determined whether or not these variations cause erosion. Impact forces include shock pressure and water hammer but these are restricted in extent because specific circumstances are required for them to operate effectively. Water hammer is the impact between a body of water and a solid and only operates if no air is trapped between the water and solid. The generation of shock pressures require the trapping of an air pocket by the front of a breaking wave as it impacts against a vertical structure. Shock pressure and water hammer are also dependent on the occurrence of breaking waves impacting directly on a platform or against the cliff. Since water depth determines the position of wave-breaking relative to the shore, true impact forces from breaking waves seldom operate on shore platforms. Larger waves break in deeper water further from the shore so that it is even less likely very large storm waves generate these forces. Only in situations where the depth of water in front of a platform is exactly suited to the breaking, water depth relationship, is there any likelihood of shock pressure and water hammer occurring and then only at the seaward edge of a platform. Water depths on platforms at high tide and even during storm and wave setup do not permit large breaking waves to impact directly against the cliff. More common is for the bore of a broken wave to shoal across a platform surface. While forces exerted are less than those generated by shock pressure and water hammer the greater frequency of air compression in joint spaces is likely to mean that it is the most important of marine processes. Associated with marine processes is abrasion, which is the wearing away of bedrock by the to and fro movement of sediments under wave action. Under more violent conditions sediments can be hurled against cliffs, acting like missiles. The effectiveness of abrasion relies on a suitable sediment and thickness of deposit. Harder sediment lying over softer bedrock will be more effective than sediments similar in nature to the platform rock. When sediments are too thickly deposited on a platform a protective role occurs. Wave tank experiments have provided some evidence that breaking and broken waves are capable of initiating shore platforms in blocks made of sand

and plaster. Ice grounding by wave action and wind has been identified as an important development process on shore platforms in high latitudes.

Studies utilizing the Schmidt Hammer test have confirmed weathering occurs on platforms. Subaerial weathering includes both mechanical and chemical processes, such as solution, salt weathering, and wetting and drying. Some evidence exists to link higher erosion rates with greater numbers of wetting and drying cycles. The number of cycles is determined largely by the tide and varies with change in elevation within the tidal range. Salt spray, wave splash and rainfall contribute to and prevent wetting and drying depending on the timing in relation to the tidal cycle. The number of wetting and drying cycles also show seasonal variations with more in summer months than winter since drying is facilitated by higher temperatures. The growth of algae on platform surfaces reduces the number of cycles by preventing drying, especially in winter months. A subaerial process often identified is water-layer weathering. This involves a combination of weathering processes but seems to be most reliant on wetting and drying and to a lesser degree salt weathering and solution. Water layering morphology occurs where pools of water accumulate and are afforded time to evaporate and the surface dry. Pools up to several meters in diameter are surrounded by raised rims 2–5 cm high that occur along fracture lines the rock. These ridges develop along fissures because they hold water and undergo fewer cycles of wetting and drying. Water-layer weathering operates wherever seawater can accumulate and evaporate. Thus, it can occur from the low-tide level and has been reported as high as 24 m above sea level where spray accumulates. In high latitudes frost shattering has been proposed as significant in platform development. Although frost shattering is not limited to higher latitudes, it has been reported as a significant process on chalk platforms on the southeast coast of England.

Biological activity on shore platforms causes erosion but also inhibits other erosion processes. Bio-erosion can be subdivided into mechanical and chemical. Solution processes operate where biota release CO₂ into seawater, lowering the pH and in turn causing calcium carbonate to convert to calcium bicarbonate which has the effect of weakening the rock. Algae probably contribute significantly to erosion relative to other organisms. Endolithic algae have a wide vertical distribution at the shoreline and weaken rock by boring. Both endolithic and epilithic algae cause variations in the chemistry of water through metabolic functions and probably cause chemical erosion. Algae also serve as food for grazing organisms such as chitons and gastropods. Molluscs that graze on endolithic algae can achieve considerable erosion on coastal bedrock. Some chitons and gastropods carve out home scars that appear as impressions similar in size to their own shell in rock surfaces. The protective role of biota on shore platforms has been recognized but it is poorly understood. Dense and

continuous mats of algae often cover significant areas of shore platforms during winter months in temperate environments. During warmer months desiccation removes these algae exposing the platform. Algae coverage protects the substrate from mechanical wave erosion and reduces some weathering processes by preventing drying. Thick growths of kelp such as *Durvillea antarctica* on the seaward edge of platforms dissipate wave energy through motion but also contribute to erosion when they are removed. Holdfasts thrown upon platforms often contain sizable pieces of rock.

Shore platform morphology can be characterized as being either near horizontal or gently sloping. Sloping platforms (Type A) grade into the sea whereas horizontal platforms (Type B) terminate at the seaward edge with a marked cliff that drops into the sea. This drop is sometimes referred to as the low-tide cliff. Platform morphology is largely controlled by geological factors. Evidence suggests that platform type is a function of the compressive strength of rock, with Type A platforms developed in softer rock than Type B. However, wave energy also appears to play a role with Type B platforms subjected to higher wave energy than Type A platforms. A demarcation between the two morphologies can be made based on the relative differences in the magnitude of the wave energy and rock strength. Variation in the gradient of platforms has been attributed to tidal range with steeper gradients generally occurring where the tide range is the greatest. This relationship holds best for platforms in macro-tidal environments and weakens as tidal range becomes smaller. So that platforms in micro-tidal areas have the lowest gradients although sloping Type A platforms are common. However, explanations for platform gradient employing tidal range ignore the role of geology.

Platform elevation is an important, if poorly understood, component of morphology. This is surprising given the wide use made of raised platforms for reconstructing sea level and tectonic histories in many coastal settings. Traditionally the view has been that platforms develop down to the low-tide level. To date it is unknown to what elevation platform surfaces develop in relation to the sea. This problem is more complicated because of the tendency of some platforms to be located at or above the high-tide level while others are intertidal. It becomes difficult to determine whether higher platforms have developed at that elevation in relation to the present sea level or if they have emerged or relate to a higher stand of sea level. Attempts have been made to link elevation with wave energy and geology. Interestingly the field evidence does not show a consistent relationship between elevation and wave energy. Some examples occur where higher platforms are found on headlands and lower ones in embayments, examples showing no such relationship have also been reported. Again this reflects the role of geology in determining platform morphology. Platform elevation has been found to increase as the compressive strength of the rock increases.

The depth of water in front of platforms has also been shown to be an important control on platform elevation. This is because water depth controls the type of wave arriving and therefore the amount of energy delivered to the platform. If both geology and wave energy are considered together better explanations of platform elevation can be made.

Width is taken as the horizontal distance from the edge of a platform exposed at low tide to the cliff platform junction, and has often been associated with exposure to wave energy. Wider platforms might be expected where exposure to wave energy is greatest. A number of examples exist to support this proposition, but examples can also be found that show an opposite relationship with wider platform in sheltered embayments. The lack of a clear relationship reflects the strong influence geology has in determining morphological characteristics. Narrower platforms have been found to occur in rock with higher compressive strengths. The expectation that there should be a relationship between width and wave energy is reliant on a marine origin for shore platforms. The absence of a clear association between width and the wave environment may also reflect the role of subaerial processes. The distance to the seaward edge of a platform is largely irrelevant to weathering processes. The wave energy needed to remove weathered debris is significantly less than that required to erode rock from in situ, so that platforms could be much wider than under a wave energy control model of platform width. Understanding platform width is also made difficult because there is uncertainty that the seaward edge of a shore platform retreats landward. This is important because if the edge does not retreat then it marks the original position of the coastline. It might be reasonable to expect that platform width is limited by wave attenuation across the widening platform. Widening ceases when waves are no longer capable of removing weathered debris or erosion of the cliff, the platform can be said to be in static equilibrium. If, however, the seaward edge does retreat then the width is determined by the relative rates of erosion of the cliff and the seaward edge. Initially the rate of retreat of the cliff should be greater than the seaward edge to allow a platform to develop. Increasing width would attenuate wave energy and cause cliff recession to slow and could not increase until the seaward edge retreat allowed wave energy to do more work by reducing the width of the platform. It has been suggested that once a platform has developed both the cliff and edge of a platform erode at the same rate so that the entire profile retreats landward in a type of dynamic equilibrium.

The rates at which shore platforms develop are of considerable interest since these may provide an indication of the ages of platforms. Determining the age is necessary since it is possible that some platforms are polygenetic, having developed previously when sea level was at a similar position as today, or a platform that initially developed at a lower sea level has emerged to be coincident with the present sea level

and reactivated. There are two components to be considered with respect to the rate of development. The retreat of the cliff backing the platform and rate of vertical lowering of the platform. It remains to be determined whether or not the seaward edge of a platform retreats landward. Cliff retreat rates vary considerably both spatially and temporally, but commonly occur at less than 1 m a^{-1} . Vertical lowering rates have only been measured at a few locations worldwide and with only a few exception over short time periods of about two years. Generally rates are in the range of $0.5\text{--}1.5 \text{ mm a}^{-1}$ with a mean of 0.95 mm a^{-1} . Although there is considerable variation in these rates depending on location on the platform, season, and rock type. On tropical limestone, mean rates of lowering of 1.97 and 1.25 mm a^{-1} have been reported. Mean erosion rates on temperate limestone and mudstone of 1.13 , 1.48 , and 1.53 mm a^{-1} were recorded. Erosion rates on schist and greywacke 0.625 and 0.37 mm a^{-1} , respectively, have been measured. The publishing of mean erosion rates belies the fact that there is a considerable range in the measured rates, in some instances rates as high as 10 mm a^{-1} have been reported. In settings where seasonally driven weathering processes such as wetting and drying are important higher erosion rates occur in summer compared with winter. In those setting where winter storm activity is dominant higher erosion rates have been measured in winter. Two studies have extend the length of vertical lowering records beyond 10 years. Both presented erosion rates that were the same order of magnitude as short-term studies. One based on an 11-year period was cautious about extrapolating short-term erosion rates while the other based on a 20-year period suggested that short-term data can be extrapolated at least to the decadal scale. Another difficulty is that all vertical lowering rates are only measured at a scale of millimeters. There are no reports of large block disintegration thus the contribution of erosion at larger scales is unknown.

Numerous attempts have been made to numerically model platform development. These fall into two categories. Those based on the Gilbertian notion that platform development only begins when the erosive force of waves (F_W) exceeds the resisting force of rock (F_R) so that development begins when $F_W \geq F_R$. The second approach is based on empirical field evidence. Models from both approaches have attempted to elucidate the types of equilibria, elevation, width, and gradient that platforms attain. Attempts have also been made to model the effect of sea-level fluctuations and wave forces on platform development. A fundamental problem is that all modeling attempts are based on the view that platforms are wave cut. Not surprisingly these models are then used to support the view that platforms have a wave-cut origin. As yet no model has been developed that is based on a subaerial origin. A number of models have also produced contradictory results. In some instances the equilibrium form predicted is

dynamic while other models suggest a static form of equilibrium will be reached. This difference in equilibrium type arises because different assumptions about erosion of the seaward edge of the shore platform were used. Platform gradient has been modeled as a function of tidal range in some examples while others have used wave height, water depth in front of a platform, and rock strength.

Suggested further readings may be found in the following bibliography.

Cross-References

- [Cliffed Coasts](#)
- [Cliffs, Erosion Rates](#)
- [Cliffs, Lithology Versus Erosion Rates](#)
- [Notches](#)
- [Thalassostatic Terraces](#)
- [Weathering in the Coastal Zone](#)

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Shore Protection Structures

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Shore protection in its widest usage refers to the reduction or elimination of damage to the shore and backland as might be caused by flooding, wave attack, and erosion. The shore may consist of cliffs, reefs, beaches, and artificial or engineered structures that form part of the water and land interface. Shore protection structures can be classified as hard, soft, or a combination. Soft structures or soft methods of shore protection usually involve placement of beach-quality sediment, typically sand, directly on the beach, a process called beach nourishment or beach fill. The beach fill may be placed across the upper beach profile and as a dune system. In such designs, the beach berm protects the dune against erosion, and the dune protects the backland

Nicholas C. Kraus: deceased.

from flooding and wave attack. Another type of soft shore protection structure is a “nearshore berm,” referring to placement of material in an approximate linear form along the shore to break storm waves and to feed material to the beach during times of accretionary wave conditions (Hands and Allison 1991; McLellan and Kraus 1991). Information about beach fill design for shore protection can be found in Dean (2003) and in Coastal Engineering Manual (2003).

This entry principally concerns hard shore protection structures designed to reduce erosion of sandy beaches. Hard structures should usually be placed together with beach fill, as emphasized below with respect to preservation of the neighboring beaches. The generic forms of hard shore protection structures are groins, detached breakwaters, and seawalls, and these have several variations.

Elements of Planning and Coastal Sediment Processes

In considering local shore protection, the neighboring shore is also a concern and must be included in considerations because waves and currents transport sediment along the coast. Shore protection actions taken at one section of coast will usually have consequences for neighboring sections, and the time-scales of the change induced may be short (months) or long (decades). Therefore, a regional approach is best considered that includes understanding of the sediment budget for the coast. Information about shore protection in the wider sense is available in Marine Board (1995), Komar (1998), Coastal Engineering Manual (2003), and Dean and Dalrymple (2002).

Shore protection is part of coastal zone management in which the uses and functions of the shore, as well as the benefits and costs of maintaining it, are evaluated. Such an evaluation is usually done by an associated group of stakeholders, including property owners, regulatory agencies, and other organizations with interest in preserving the property, environment, and functionality of the particular shore in question.

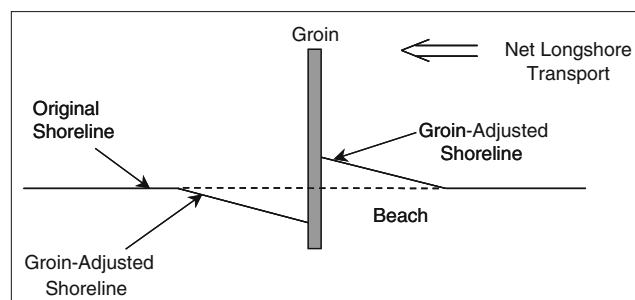
In approaching a shore protection design to mitigate beach erosion, the cause of the erosion should first be identified to determine the general approach and optimal design. For example, a rule of thumb is that if high-tide shoreline recession is greater than 1–2 m/year for a long stretch of shore, beach fill alone is probably not cost effective. Local areas of the shore experiencing significant beach erosion as compared with the adjacent shore are called erosional hot spots (Kraus and Galgano 2001). Structures may be an effective hot-spot countermeasure under certain situations. Both initial costs of construction and the maintenance costs should be considered. Shore protection is a form of infrastructure; therefore, maintenance of the structures and beach fill is part of the project life cycle.

It is convenient to classify sediment transport as being directed either alongshore or across shore. Longshore sediment transport occurs throughout the year, with storms often creating the most transport. For an observer standing on the coast, it is convenient to define left- and right-directed longshore sediment transport. The net transport is then defined as the difference of the right- and the left-directed transport, and the gross transport rate is defined as the sum of the left- and right-directed transport. Changes in the high-tide shoreline position, or erosion and accretion, are caused by differences in net transport from one location to another on the coast.

Cross-shore transport is directed either onshore or offshore. Offshore transport occurs during storms, when the beach is eroded and sediment transport is transported seaward to create a storm bar. Sediment may also be transported onto the land (overwash), and this material is called washover. After a storm and during summer or mild wave conditions, sand that is stored in storm bars is moved onshore to widen the beach. In the following, the predominant direction of transport enters in consideration of the action of a shore protection structure.

Groins

Groins are shore-normal structures (Fig. 1) emplaced for the purpose of either (1) maintaining the beach behind them, or (2) controlling the amount of sediment moving alongshore. Because modern coastal engineering practice includes a regional perspective that incorporates the stability of down-drift beaches, groin construction is normally done together with beach nourishment so that sediment impounded by the groin or groins is replaced in the total system. If beach fill is not included with groin construction and maintenance, downdrift beaches may suffer (Nersesian et al. 1992). Groins have varied shapes, such as the simple groin, spur or L-shaped groin, and T-head (Fig. 2). These shapes were developed to better retain sediment by reducing offshore transport by the rip current that tends to form at the stems of groins, and by providing a sheltered region from waves.



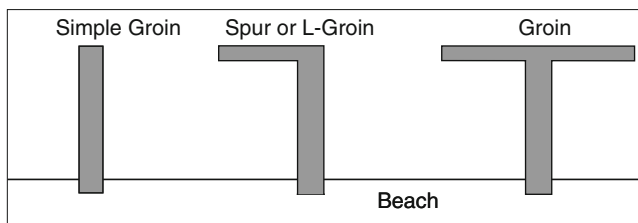
Shore Protection Structures, Fig. 1 Definition sketch for beach response to a groin

Groins are designed to operate where there is appreciable longshore sediment transport. Groins will not function well if cross-shore transport is strong, such as in the US Great Lakes, where steep waves tend to move sand offshore that can then bypass the groins by transport along longshore bars. Groins are often placed as a system or a field. Fig. 3 shows a schematic of a tapered groin field, with shortening of the groins in the direction of net transport to transition into the beach without structures.

Groin functioning and design can be parameterized in a numerical simulation model through three key processes (Kraus et al. 1994) as (1) sediment bypassing around the seaward end of the structure, (2) structure permeability in allowing sediment to go through or over the groin, and (3) ratio of net to gross longshore transport rate, which controls in great part high-tide shoreline recession and advance at a groin, as schematically shown in Fig. 3. Groin length determines the amount of bypassing. Typically, a groin will be shorter than the average width of the surf zone. A long groin will intercept too much sand moving alongshore and cause erosion of the downdrift beach by impounding or blocking sediment on the updrift side.

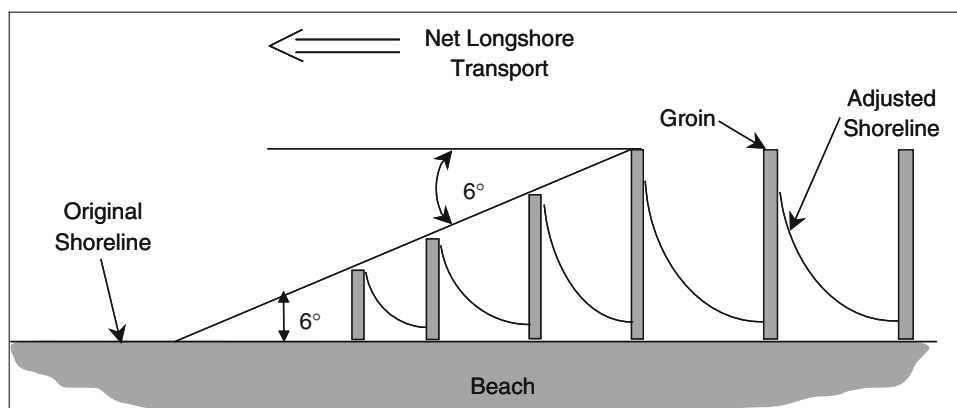
Groins may be a possible component of a shore-protection project and sand-management program under the following situations:

1. Where there is a divergent region of longshore transport, such as at the center of a crenulate-shaped pocket beach, or where the curvature of the coast changes greatly.



Shore Protection Structures, Fig. 2 Examples of types of groins

Shore Protection Structures,
Fig. 3 Tapered groins



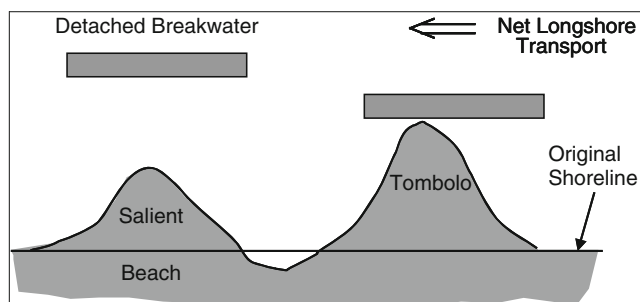
2. Where there is no source of sand, such as the downdrift side of a large harbor or inlet with jetties.
3. Where intruding sand is to be managed, such as to retain sand on the updrift side of a harbor to prevent shoaling of the channel and to stockpile the sediment for bypassing by land transport.
4. Where the sand transport rate is to be controlled or gated, such as to prevent undue loss of beach fill to unwanted areas.
5. Where an entire littoral reach is to be stabilized, such as on a spit, near a submarine canyon.
6. Where stabilization of the high-tide shoreline is required under extreme conditions.

Groins cannot prevent sand from moving offshore, such as during storms. For such situations, detached breakwaters serve better for shore protection.

Groins may be made porous or low to allow some amount of sediment to bypass them. Also, groins may be notched at the shoreward end to allow sand to pass through them that is being transported near the high-tide shoreline (Kraus 2000). Such designs are intended to balance the amount of material bypassed to the amount retained under certain wave conditions and water level. Also, permeable and notched groins tend to minimize abrupt change in high-tide shoreline position on the sides of the groins, promoting a straight shoreline.

Detached Breakwaters

Detached breakwaters are sometimes referred to as offshore breakwaters. They function to reduce wave energy on their landwards sides. Sediment will accumulate in the wave-sheltered region because the water is calmer and the longshore current behind the detached breakwater weaker than on adjacent shore that is open to full wave energy. Detached breakwaters are often constructed to be partially submerged, for example, at higher tide, to allow wave transmission over and, sometimes, through them. In this way, a certain amount of wave energy reaches the sheltered region and promotes



Shore Protection Structures, Fig. 4 Definition sketch for beach response to detached breakwaters

sediment transport alongshore. Detached breakwaters are often built in groups, and this configuration is referred to as a segmented detached breakwater system. Detached breakwaters may be an appropriate shore protection measure in areas where cross-shore transport is large, or where wave energy must be reduced, such as at a change in orientation of the coast.

The response of the beach to detached breakwaters is controlled by at least 14 variables (Hanson and Kraus 1990), making these structures more difficult to design than groins. The response of the beach to the presence of a detached breakwater (Fig. 4) may be as a tombolo, for which the high-tide shoreline reaches the structure, or as a salient that describes a cusped morphologic form that grows toward, but does not reach the structure. The type of beach response is determined by design requirements. Typically, it is desired to allow some amount of sediment to pass alongshore behind the structure. A potential design deficiency for detached breakwaters is to place them too far offshore. In such a case, the high-tide shoreline will show no response.

Main design parameters for detached breakwaters are distance offshore, length of structure, and transmission (depends on elevation of structure and its composition). The design takes into account the prevailing wave climate at the site and the desired beach response. For segmented detached breakwaters, the gap width between structures is also a key design parameter. Empirical design procedures have been developed that incorporate the geometric parameters of detached breakwaters (Pope and Dean 1986; Rosati 1990), whereas numerical simulation models can be applied to account for a wide range of conditions, including wave transmission (Hanson and Kraus 1990). As depicted in Fig. 4, a detached breakwater built closer to shore will tend to create a tombolo. Erosion or high-tide shoreline recession between structures can occur and must be accounted for in design.

Submerged detached breakwaters are sometimes called reefs, particularly if they have a broad crest. Reef breakwaters allow transmission, the wide crest reducing wave height behind them. Reef breakwaters often serve as an offshore retaining structure for sediment near the beach, acting to perch or elevate the beach.

Headland control is a variation of detached breakwaters, in which hard points or shore-connected detached breakwaters are constructed, often with orientation parallel to the crests of the predominant incident waves (Silvester and Hsu 1993; Hardaway and Gunn 1999). In this way, a curved headland beach, or a pocket beach if bounded laterally by two such structures, will form. The intent is to develop isolated beach compartments that will lose a minimum amount at their longshore ends, letting the high-tide shoreline achieve a form in equilibrium with the predominant direction of the incident waves.

Seawalls

Seawalls are constructed to prevent inland flooding from major storms accompanied by high water levels (storm surge) and large waves. Seawalls also fix the position of the land sea boundary if the sea reaches the structure. The main functional element of a seawall is the elevation to minimize overtopping from storm surge and wave runup. A seawall is typically a massive, stone and concrete structure with its weight providing stability against sliding forces and overturning moments. Whereas groins and detached breakwaters are constructed with preservation of the beach as a design goal, the purpose of seawalls is to protect the backland. Knowledge on seawalls and their interaction with beaches is compiled in a collection of papers found in Kraus and Pilkey (1988), and the literature on the subject has been brought up to date by Kraus and McDougal (1996).

Seawalls have many counterparts in nature, as do groins and detached breakwaters. Wiegel (2002a, b, c) gives an account of many types of seawalls and their performance.

Cross-References

- Beach Erosion
- Beach Processes
- Cross-Shore Sediment Transport
- Gross Transport
- History, Coastal Protection
- Longshore Sediment Transport
- Net Transport
- Washover Effects

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Shoreface

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Introduction and Origins of the Term

The shoreface, a relatively steep surface that slopes away from the low-tide shoreline and imperceptibly merges with

the flatter inner shelf or basin plain ramp, is an integral feature of nearly all clastic coasts (Fig. 1). The term has been applied in a facies context in many geological interpretations of ancient coastal deposits in the rock record. This usage commonly invokes additional attributes that are unrelated to the morphologic feature and commonly unfounded.

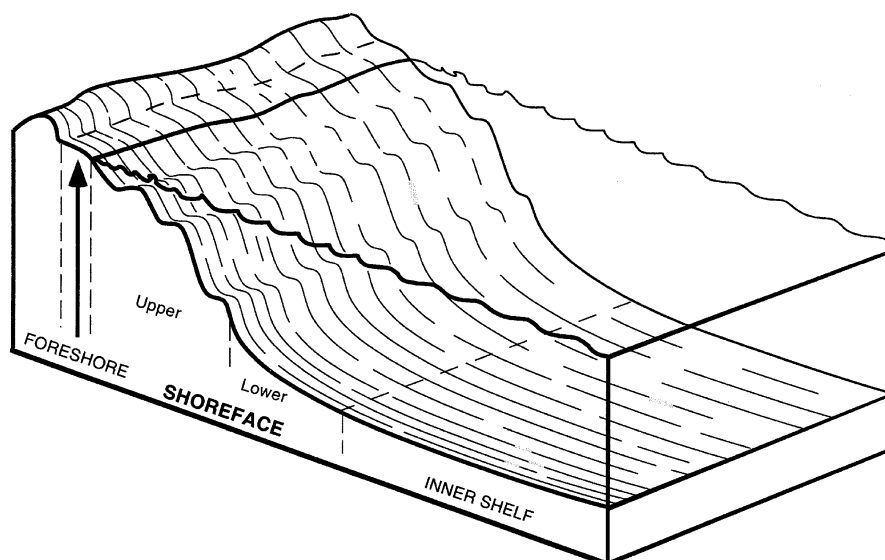
Barrell (1912) coined the term “shore face” to describe the transition between the subaerial and subaqueous plains of a deltaic topset system: “The shore face is the relatively steep slope developed by breaking waves, a slope which separates the subaerial plain above from the subaqueous below.” Johnson (1919) consolidated the term (“shoreface”) and redefined it as the zone between the low-tide shoreline and the more nearly horizontal surface of the offshore. He expanded its usage to other types of coasts, as “the steeper, landward portion of the shore profile of equilibrium, of which the profile of the gently sloping subaqueous plain is the seaward continuation.”

Development and Maintenance of the Shoreface

Early in the history of the concept, it was recognized that the shoreface is part of an equilibrium profile generated by shoaling waves. Many workers have assumed that the base of the shoreface is defined by fair weather wave base, but as Johnson noted in 1919, the two are unrelated. Wave base is commonly taken as a water depth equal to half of the length of the passing waves. As such it is primarily dependent on wave period. On the Pacific coast of the United States, long period (310 s) waves are common under fair-weather conditions, and their wave base lies in water depths of 78 m or more, well beyond the shoreface. On coasts receiving shorter period waves, the base of fair-weather waves lies in shallower water (e.g., about 20 m for 5 s waves), but coincidence of fair-weather wave base and the bottom of the shoreface is just that: a coincidence.

Although unrelated to wave base as such, the development and maintenance of a shoreface is at least partly controlled by aspects of shoaling waves. Niedoroda and Swift (1981, see also Niedoroda et al. 1984) provide a plausible explanation for the maintenance of the shoreface, that balances the onshore transport of sand under (mostly) fair-weather conditions with the offshore transport during storms. In shallow water, the oscillatory flow of water induced by passing waves becomes asymmetric, with short relatively strong landward flow under the crest of a wave and a more prolonged weaker, seaward flow under the trough. This orbital velocity asymmetry drives sediment in a landward direction, building the shoreface. Storms sporadically interrupt this process and seaward flowing rip and geostrophic currents erode the shoreface. Much of the eroded sediment is subsequently returned to the shoreface during fair weather, maintaining its

Shoreface, Fig. 1 Sketch showing the character of a simple (progradational) shoreface. Shore-parallel breaker-bars, which may not occur on all shorefaces, are here used to separate it into “upper” and “lower” parts, which may be the only components recognizable in ancient shoreface deposits



concave-up profile. Evidence for this combination of processes will be presented in a following section “Sedimentologic aspects of the shoreface.”

Shoreface Profiles

The shape of the shoreface profile on constructive (progradational) coasts differs from that of erosive (or transgressive) coasts. Progradation occurs where high rates of sediment supply (relative to base level change) cause a coastline to build laterally into the adjacent basin. Sediment is sufficient to provide a bottom profile in equilibrium with the shoaling waves. A simple concave-up profile typically results (Fig. 2) that resembles the theoretical curve for landward increase in bottom orbital velocity asymmetry generated by shoaling waves (Fig. 3). Because of the large volume of sediment available, the relief of progradational shorefaces reflects the energy of the coastal waves. On high-energy coasts like the southern coast of Washington State, the shoreface extends to a depth of 10 m (Fig. 2c). On coasts with lower wave energy (e.g., the Gulf coast at Galveston, Texas), the shoreface bottoms out in shallower depths (Fig. 2a).

Most coasts today are still in a transgressive or erosional phase, where the volume of sediment is insufficient to create a simple equilibrium profile. Commonly on such coasts, older deposits crop out on the shoreface, where they resist erosion by waves and coastal currents. Differential erosion creates irregularity in the shoreface profile (Fig. 4), although many such profiles contain a simple “wave-tuned” component in their upper part, either because that is where unconsolidated sediment is concentrated or simply owing to intensified wave erosion in the shallowest water (Fig. 5). Because transgressive coasts typically lack an abundance of sediment, the depth to which they extend reflects in part the topography crossed by

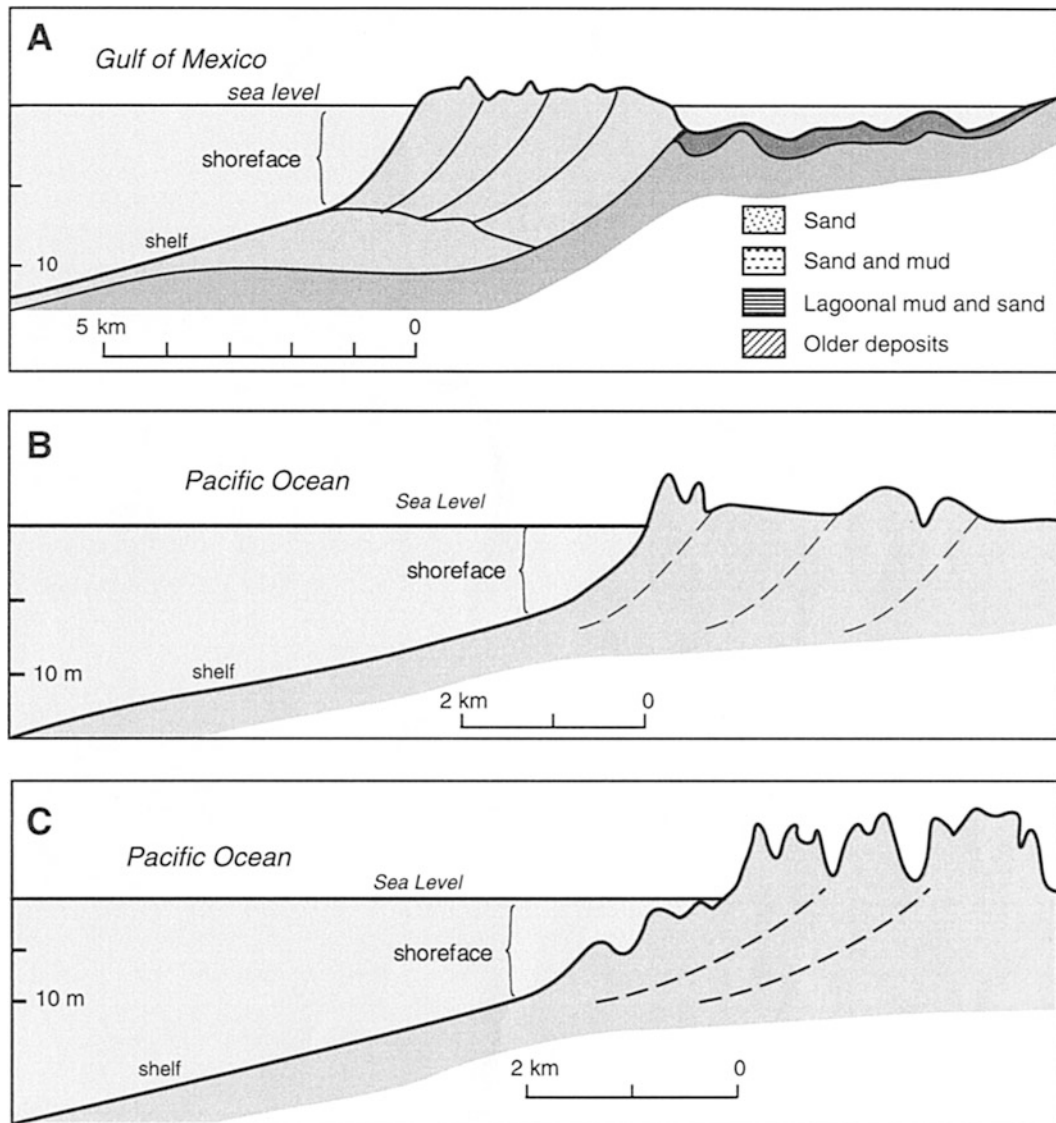
the transgression. The relief of transgressive shorefaces is highly variable (Figs. 4 and 5), but can exceed 20 m (Fig. 4c).

Sedimentologic Aspects of the Shoreface

The textural character and sedimentary structures of shorefaces differ with energy and provenance. Typically, wave energy reaches a maximum in the shallower part of a shoreface, and coarse sand and gravel, if available, are concentrated in the active breaker zone adjacent to the shoreline (the coarsest typically being at the very base of the foreshore). The grain size typically decreases to seaward across a shoreface to fine or very fine sand at its base.

In some areas, the base of the shoreface marks the transition between sand and mud (Fig. 2a), but in many places sand continues well out onto the inner shelf. On the coast of southwestern Washington State, for example, fine to very fine sand extends to water depths of 60 m or more. Although some geologists refer to the entire sandy part of a shoaling-upward coastal succession as shoreface deposits, there is no general basis for this interpretation. The proportion of sand, silt, and mud, depend on the sediment available and the wave-energy regime (Galloway and Hobday 1996) and are not specifically related to the shoreface morphology.

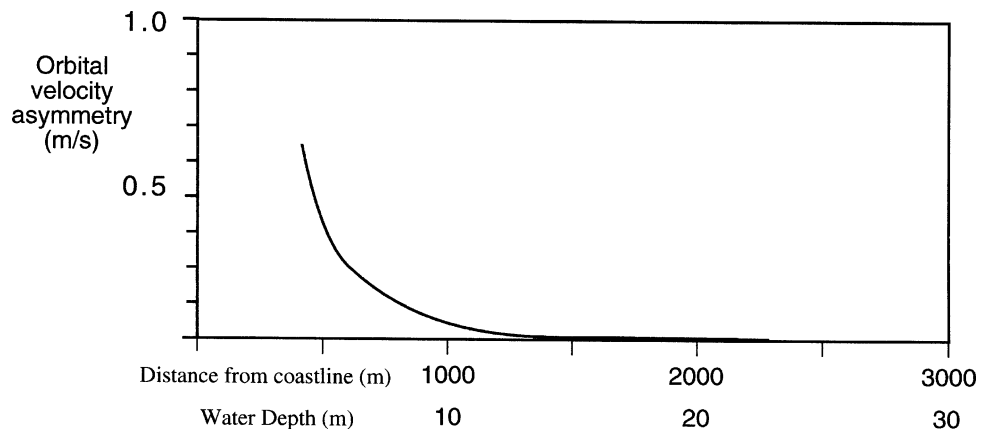
An exception to the generally seaward-fining on most shorefaces occurs where gravel is present. Studies of modern gravel-bearing shorefaces (Shipp 1984; Howard and Reineck 1981) show a scattering of gravel at the base of the shoreface and the immediately adjacent inner shelf. Vertical successions of ancient gravelly shoreface deposits typically have a distinctive facies composed of scattered small pebbles in a matrix of fine- to very fine-grained sand in the interval 2–3 m below cross-bedded nearshore gravelly sand. The gravel at the base of the shoreface probably results from a

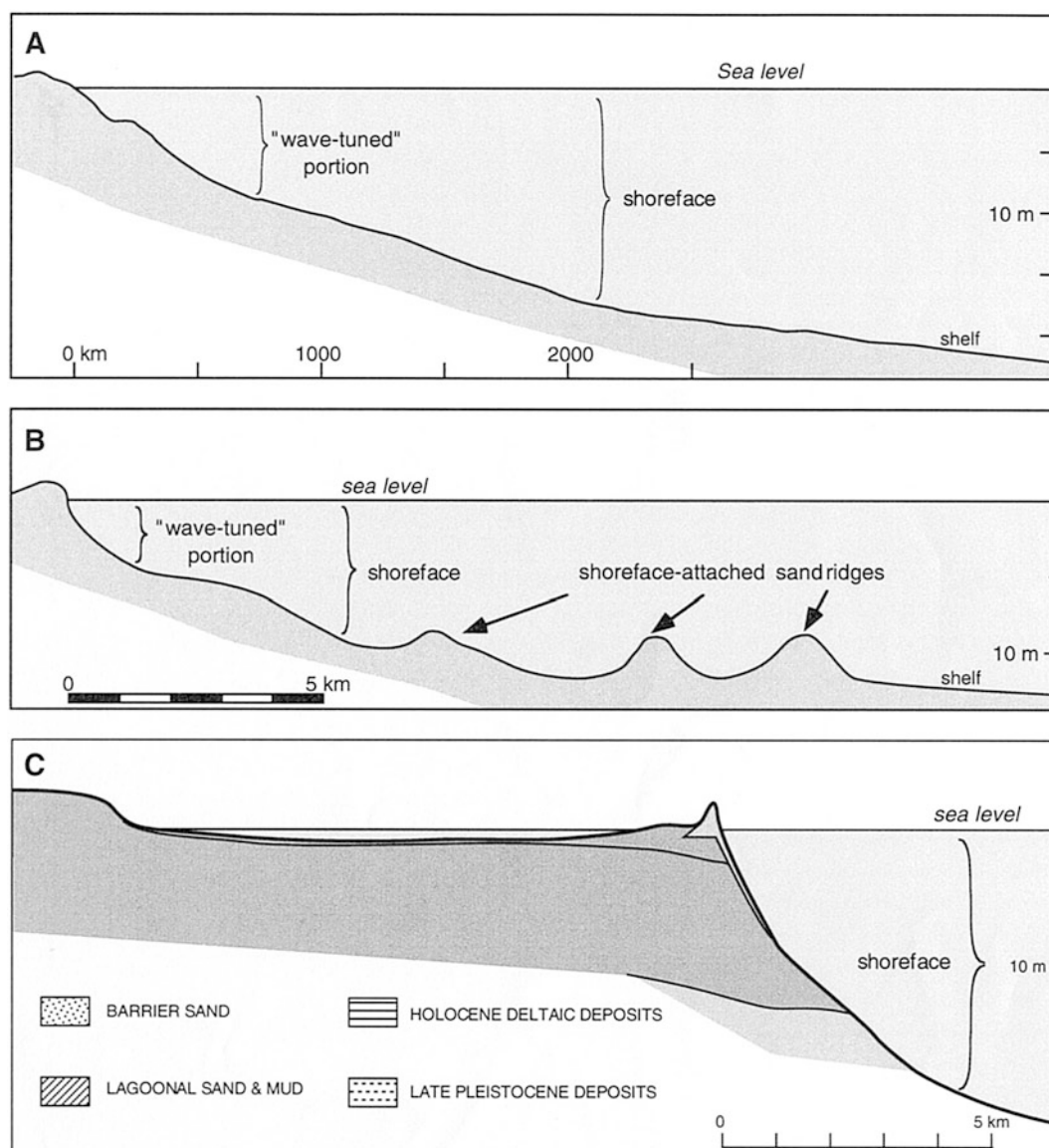


Shoreface, Fig. 2 Shoreface profiles on progradational coasts, showing simple concave-up character. (a) Galveston, Texas, Gulf of Mexico. (After Morton 1994). (b) Nayarit coast of Mexico, Pacific Ocean. (After

Curry et al. 1969). (c) Long Beach, Washington, Pacific Ocean. (After Dingler and Clifton 1994). Note differences in scale

Shoreface, Fig. 3 Predicted orbital velocity asymmetry (difference between the bottom velocities under the crest and trough of a shoaling wave, using Stokes second order wave theory) under 10-s waves with a deepwater wave height of 1.0 m over a bottom with a uniform slope of 1:100. Water depth at which velocity asymmetry becomes noticeable approaches the depth of the base of the shoreface off the Long Beach, Washington (Fig. 1c) where these wave conditions obtain during fair weather





Shoreface, Fig. 4 Shoreface profiles on transgressive coasts, showing variability of profile shape. (a) Long Island, New York. (After Niedoroda et al. 1984). (b) North of the Edisto River, North Carolina. (c) South Padre Island. (After Morton 1994). Topographic irregularity below

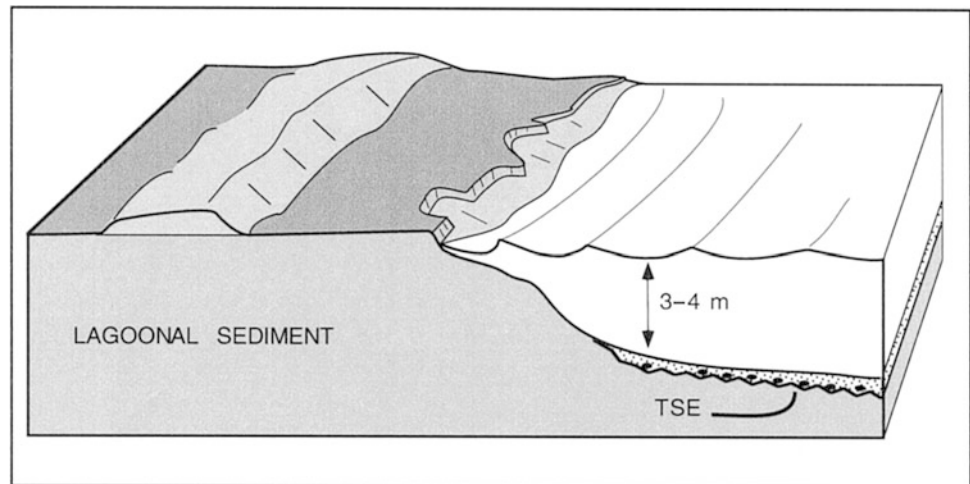
“wave-tuned” portion in a and b presumably due to differential erosion of older deposits exposed on the shoreface during transgression. Note differences in scale

combination of erosion of a shoreface during storms and its subsequent rebuilding in the storms aftermath as suggested by Niedoroda and Swift (1981). Gravelly sand eroded from the upper shoreface is carried seaward by rip currents or geostrophic flow and deposited at or near the base of the shoreface. Waves following the storm rework and winnow this material and transport the finer component landward. The coarsest part (typically small pebbles) is left behind as a “post storm lag.”

The sedimentary structures of a shoreface depend on both wave energy and sediment texture. The shallower part consists of the nearshore, in which bedforms are generated by

passing waves or by longshore and rip currents of the near-shore circulation cell. Bedform size, shape, and orientation differ as a function of wave size and direction of approach and the grain-size of the sea bed material (Clifton and Dingler 1984). Storms may generate hummocky or swaley crossstratification. Guttercasts are a characteristic shoreface feature. Shore-parallel or other breaker bars are common features on the upper part of many shorefaces (Fig. 1) and can serve to focus longshore and rip currents. Depending on infauna and sediment texture, bioturbation can obliterate much of the physical structures on the outer part of a shoreface.

Shoreface, Fig. 5 Erosional shoreface on a transgressing coast, Cape Romaine, South Carolina. (After Hayes 1994)



Shoreface Subdivisions

Many workers divide the shoreface into upper, middle, and lower components. Galloway and Hobday (1996) define the upper shoreface as the inner surf zone, the middle shoreface as the portion occupied by breaker bars systems and the lower shoreface as the area to seaward where the shoreface merges with the inner shelf. Howard and Reineck (1981) recognize three principal zones without relating them to shoreface morphology: a *nearshore* made up of clean sand in which physical sedimentary structures predominate, a *transition* in which fine and silty sand contain a mixture of biogenic and physical structures, and an *offshore* composed of sandy silt with only remnant stratification.

Shoreface zonation is less evenly applied to the stratigraphic record. Although geologists commonly subdivide ancient shoreface deposits into upper, middle, and lower facies (and some subdivide these further into proximal and distal components), they generally agree only on the character of the upper shoreface as an interval with abundant crossbedding and other physical structures. This would probably include the middle shoreface of Galloway and Hobday (1996) where currents focused by bar-trough systems produce structures that resemble those of the surf zone. Deposits on the lower part of a shoreface may be indistinguishable from those on the adjacent inner shelf, unless gravel is present.

Otherwise the most realistic facies subdivision is that of an upper shoreface overlying a lower shoreface/inner shelf transition.

Cross-References

- Bars
- Beach Sediment Characteristics
- Beach Stratigraphy

- Coastal Sedimentary Facies
- Cross-Shore Sediment Transport
- Dynamic Equilibrium of Beaches
- Shelf Processes
- Surf Zone Processes

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Shoreline and Coastal Terrain Mapping

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Concept

Mapmaking process of coastal environments depicts forms, processes, and evolution. This graphic representation of the coasts provides a necessary background in marine sciences, geosciences, and engineering for those studying coastal systems. The coastal environments are complex open systems with several types of sources of energy and dynamics. The common ground for shoreline and coastal terrain mapping is to shape the landforms and occurrences in terms of actions and developments, which draw on different types of coast over thousands of years. Maps of shores and coastal terrain could have different purposes but a shared meaning: to ensure a sound communication. In the form of hard copy or digital and no matter what the output is, maps establish a platform that allows handling data. In addition, it enables to display all the synthesized information, to interpret and analyze the past and present, and to predict the future of coastal or environmental conditions. Moreover, maps provide the conceptualization of inventories, documentation, and geo-databases to monitor and assess shoreline and coastal changes. They offer an approach through baseline conditions and the coastal resources implications for coastal management and territory planning assemblage.

The scientific interest in coasts started in the post-Renaissance period in the context of navigation beyond the Mediterranean. Global exploration such as geological or biological investigations was frequently extensions of the survey work carried out by naturalists and followed by the analysis of hydrographers (Woodroffe 2003). During the 1950s, the geologists and oceanic cartographers Marie Tharp (1920–2006) and Bruce C. Heezen (1924–1977) were challenged to make

maps and to provide the first bathymetric representation of the mid-ocean ridge in the Atlantic Ocean. This impressive map illustrates the seafloor topography and coastal mapping. Marie and Heezen's outstanding work shows essentially the importance of mapping. These maps allowed evaluating the processes that contribute to the evolution of coastal environments (more details in Hamylton 2017). During the twentieth century, maps developed exponentially. The more advanced the technologies, the more exceptional the maps that will be generated. From the several devices available, more accurate data and surveying, to the photogrammetry techniques, it is possible to perform more accurate mapping. In the last decades, it is possible to benefit from the rugged computers, high-resolution GPS, or even tablets or iPad's that make it possible to create maps in real time. On the other hand, remote sensing approaches are also available which allow an outstanding quality of satellite imagery, geospatial information, and geo-visualization (e.g., single or combined LiDAR, Landsat ETM+, SPOT, ...), sweeping almost the entire Earth with current maps. Therefore, whoever works with maps must be aware that he is living in a digital world exposed to constant evolution.

When talking about mapping, it is imperative to address cartography, either traditional or modern, which constitutes many theoretical and practical foundations to build maps. Many researchers have gained consciousness that is necessary that contemporary cartography matches today's challenges or perhaps a rigorous tailoring the maps to our requirements and characteristics (e.g., Ruas 2011a, b; Hennig 2013; Pires et al. 2016; Peterson 2017). The sources of data used to map shoreline and coastal terrain can differ widely depending on the applications and initial requirements to characterize the coastal systems and contiguous submerged areas. In general terms the spatial coverage could be provided by (i) imagery (aerial photographs or satellite imagery data), (ii) digital elevation models (DEM) or triangulated irregular network (TIN) derived from DEM, (iii) raster layers or features (resulting from images), and (iv) vectorial layers or features (points, lines, polygons, which are representations produced from digitization). In addition, historical data (e.g., old maps or charts, archives, occurrences, atlases, ...) may be sourced from digitization. All these types of data sources can involve georeferencing operations using ground control points to adjust the source accurately and properly. Moreover, thematic layer's creation or production has great importance, for example, physical, biological, geological, morphological, and socioeconomical parameters, for marine and terrestrial areas. Most of the recent advances in science and technology used for gathering data, geo-visualization and analysis, or even for mapping shores and coastal terrain are discussed in other sections of this encyclopedia.

Maps of shoreline and coastal terrain are presented in a wide range of formats, with several scales depending on the

different purposes. The key role of coastal maps in a dual perspective focused on the main purposes and on end users could be displayed as the following (Chaminé et al. 2016): (i) general shoreline and coastal maps proposing strategies and recommendations in terms of littoral management (generally, regional scale to continental and global scales), often simplified, are produced to communicate with politicians, the general public, and students; and (ii) thematic and applied maps or models, at several scales (mainly, large scale to local scales), are created by practitioners and or researchers for the exploration, characterization, description, and evaluation of coastal systems, resources, ecosystems, or environments. It is worth noting the importance of temporal and spatial coastal scales, mentioned by Woodroffe (2003), which represents examples of coastal systems or patterns of variation and their general position in terms of space-time scales. This is one of the many perspectives available for representation of space and time scales appropriate for the study of coastal morphodynamics (Hamylton 2017). It is fascinating how different background and scientific interests could influence the perspectives and consequently the directions and approaches to research and practice (e.g., Dykes et al. 2005; Pérez-Alberti et al. 2012, 2013; Trenhaile 2014; Daffor et al. 2015; Chaminé et al. 2016; Pires et al. 2015, 2016).

Presently, maps in digital format can be stored in servers, computers, DVDs, external disks, or even Internet websites and cloud services, which are available to users for fast dissemination. The use of geographic information systems (GIS) technology made available is one of the possible types of enhanced platforms that allows displaying multiple layers and performing spatial analysis or advanced calculations. Both GIS and cartography can influence each other, but the most important is that mapping could get the best of the relation. However, this physical visualization and manipulation of coastal maps can also be made by handling paper throughout large atlases comprised by several sheets and series or oversized charts. In Parry and Perkins (2002), it is possible to understand the dimension of mapping industry within each country by providing a unique summary of the wealth of maps and spatial data available.

Almost 20 years have passed, when emerge the vision of the so-called Digital Earth; “(…) a multi-resolution, three-dimensional representation of the planet that would make it possible to find, visualize and make sense of vast amounts of geo-referenced information on physical and social environments” (Gore 1999). In fact, digital mapping, Digital Earth, GIS applications, geo-visualization, GIScience, or even the traditional cartography and mapping techniques are all labels of a broad approach which should include all aspects of visualizing geospatial information (Dykes et al. 2005; Hennig 2013). Finally, it is also important to mention and highlight

the overall importance of maps as essential tool in climate variability framework and climate change assessment.

Background and Initial Context

It is important to understand what has changed in the field of shoreline and coastal terrain mapping. Therefore, it is necessary to analyze the past, to understand the present, and to predict the future. This section reveals the impressive vision by Morton (2005) of *Mapping Shores and Coastal Terrain* concept. Figure 1 and Table 1 show the structure adopted and to perceive a shared definition which emphasizes three main ideas: (1) maps of coastal features could have several attributes; (2) maps are applied to interpret the geological background of a given region, in order to plan and manage coastal resources; and (3) maps are used to communicate.

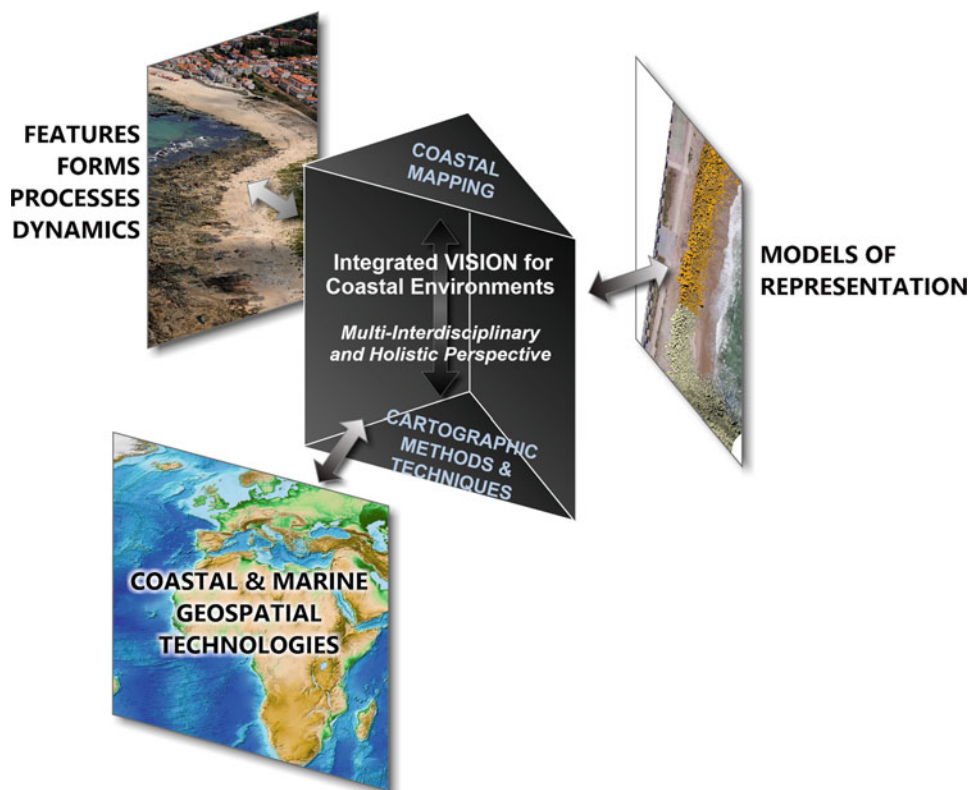
Morton (2005) insights of *Mapping Shores and Coastal Terrain* essentially is reflected in Fig. 1, which displays a tag cloud of the Morton's text, and the keywords that stand out are coastal, mapping, terrain, hazard, maps, parameters, scientific, risk, vulnerability, change, data, resources, environmental, etc. The text was organized in three main sections: (A) coastal mapping strategies, (B) coastal mapping applications, and (C) integrated maps and predictive models. These key concepts were developed in a comprehensive approach based on the core sections: strategies, applications, and models (Table 1).

Regarding the existing needs and future directions, Morton's point of view leads to a contemporary facts and insights. There is no doubt that the coast plays an important role in global transportation and is the most popular tourist destination around the world. During the years coastal scientists have tried to understand the shoreline in relation to the processes that shape it and its interrelationships with the contiguous superficial marine and terrestrial hinterland environments (e.g., Woodroffe 2003; Bird 2008). Those factors encourage the need for coastal maps, because of its possible use in identifying coastal management issues to consider in policy strategies, actions, and planning. As stated by Morton (2005), there is a demand for coastal maps to be more accurate. There has been a considerable development of maps that evolved from being quantitative and descriptive maps to being more quantitative and process-oriented. It has never been so crucial to improve these maps in terms of error and accuracy, as well as to design maps in terms of hazards and or risks. Moreover, it was also stated the concern about obsolete maps and the importance of making dynamic, interactive, and multimedia maps. This approach was set up to promote the importance of coastal mapping by suggesting methods and techniques, the latest state-of-the-art, in terms of terminologies, design, and concepts, as well as the more advanced technologies available. Nevertheless, over 12 years have

Shoreline and Coastal Terrain Mapping, Table 1 Main ideas based on the core sections by Morton (2005): strategies, applications, and models (see A, B, and C on Fig. 1)

 A Coastal mapping strategies	 B Coastal mapping applications	 C Integrated maps and predictive models
1. "...the purpose of morphostratigraphic mapping is to reconstruct the paleogeography and thus the geologic history of a region."	1. "Global coastal classifications typically involve aerial photograph and map compilations that compare and contrast various attributes of the shore and adjacent land (tectonic setting, morphology, biogenic characteristics, depositional origin, terrain composition), (...) combined with climatic overprints (...) and modifying processes (relative sea level, tide range, wave energy, wind regime, and volcanic activity) (...)"	1. "(...) assume that coastal change in the future will be similar to that recorded in the past. (...) the historical record of observed changes is the best predictor of future changes."
2. "Genetic stratigraphy incorporates lithofacies and biofacies attributes, three-dimensional geometries, vertical sedimentary successions, and lateral facies relationships to interpret the origins of sedimentary deposits and to reconstruct the paleogeography at the time of deposition (shore position, depositional strike, and dip)."	2. "A modern trend in computer-assisted mapping of coastal terrain integrates sophisticated computer applications and a GIS to generate indices or some other quantitative attribute that can be used to classify coastal terrain with respect to some anticipated hazard."	2. "(...) are based on the premise that coastal change is caused primarily by a relative rise in sea level. (...) employ several simplistic assumptions such as a smooth, curved equilibrium profile (no offshore bars) that does not change shape as the beach retreats. (...) allow only for onshore and offshore movement of sediment (no net alongshore movement) and they presume a water depth on the profile beyond which no sediment is eroded or deposited."
3. "Maps of biological resources include the distribution and composition of diverse wetlands, reefs (coral and oyster), faunal assemblages, fishery habitats, and areas of environmental protection, or concern such as breeding or nesting grounds."	3. "(...) Environmental Sensitivity Index (ESI) maps (...) ESI mapping represents a conceptual advancement that recognizes different susceptibilities to environmental damage from oil spills depending on characteristics."	3. "Some methods of predicting coastal change combine long-term rates of change, determined by air photo methods (taken as background change), with shoreline retreat predicted by the Bruun Rule."
4. "...maps depicting areas threatened by natural coastal hazards such as beach erosion, storm flooding, storm-surge washover, earthquakes and liquefaction, tsunamis flooding, landslides, subsidence, and active faulting."		4. "Most numerical models, also known as deterministic models, presume that coastal changes are mainly caused by wave energy. Like the geometric models, most of the deterministic models also assume a smooth, curved offshore profile that does not change shape as the beach retreats, only onshore and offshore movement of sediment, and a water depth on the profile beyond which no sediment is eroded or deposited. Most of these models also assume that sea level remains constant for the period of time that coastal change is being predicted."
5. "Maps of coastal terrain can serve either as catalogs of natural resources or as historical snapshots of dynamic conditions."		

Shoreline and Coastal Terrain Mapping, Fig. 2 Prismatic figure sketches of the modern framework for shoreline and coastal terrain mapping: multidisciplinary and holistic outlook



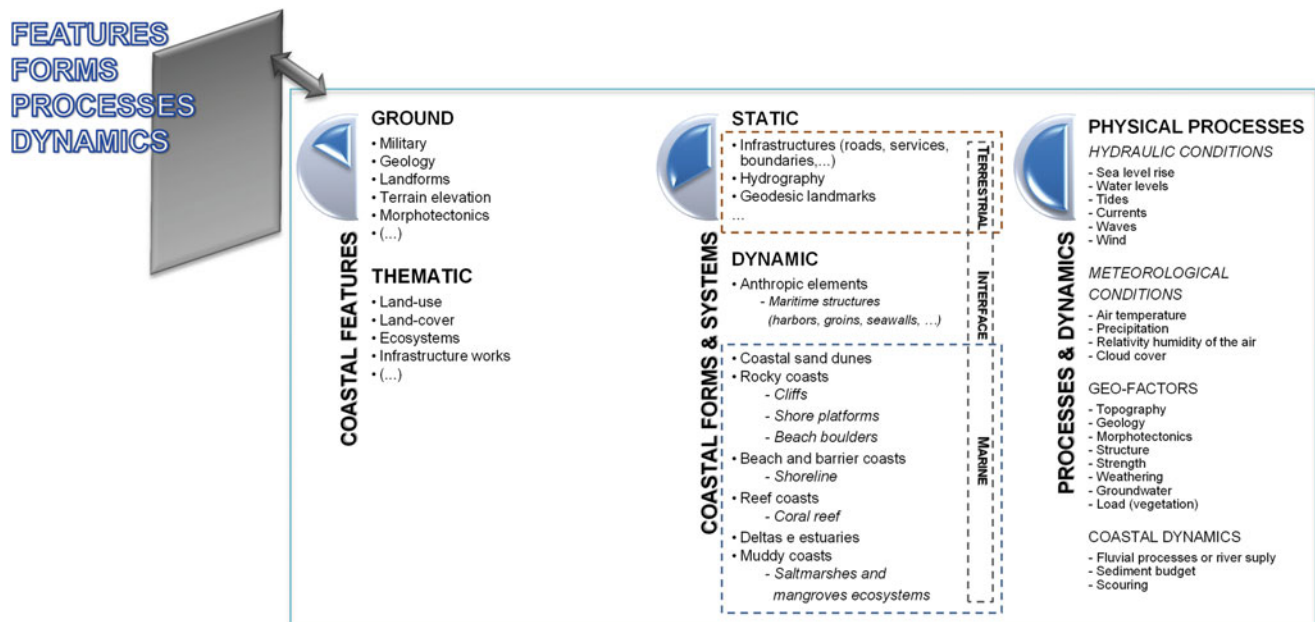
study areas. This integrated vision aims to be an “umbrella structure” (prism) bringing together different types of coastal and shoreline applications and mapping.

Cartographic Strategies and Analysis

Coastal Systems

The coastal environment is one of the most dynamic and energetic interfaces between human society, ecosystems, and sustainable environment (Bird 2008). Furthermore, the coast comprises a complex and energetic transition between the marine environment and the terrestrial environment. The front view of the prism shows a broad outlook of all the elements to consider in shoreline and coastal mapping (Fig. 3): (i) features, (ii) forms and systems, and (iii) processes and dynamics. Coasts are defined by their singular geological features or by the geological event that formed them. It also includes a wide range of areas, topographies, forms, and geologic settings (Davis 2002). Coastal features category relates to the source of layers available and applied as baseline or primary overlay (Fig. 3). Ground features are used as basis to support coastal studies, providing the first indicators and data to assist fieldwork or even to support a detailed analysis on the area. Thematic features are derived from ground features and related with or specific

type of maps which can also help the coastal research and assessment. Coastal forms and systems category relates to the coastal environments and the complex interaction processes involved which are continually modifying rocks and sediments (USACE 2017a). As stated by Dean and Dalrymple (2004), to understand coastal processes (including near-shore flows and the resulting sediment transport) and to have the skill to change this understanding into operational engineering measures, the following factors are required: (i) a blend of analytical capability, (ii) an interest in the workings of nature, (iii) the ability to interpret many complex and sometimes apparently conflicting pieces of evidence, and (iv) experience gained from studying a variety of shorelines and working with many coastal projects. Due to these facts, processes and dynamics category is intrinsically linked with the previous one. The nature and form of the coastal system at any part of the world is influenced by several factors and those usually associated with coastal morphodynamics (e. g., Woodroffe 2003; Davis 2002). Within the coast occur beaches, barrier islands, tidal inlets, coastal estuaries and lagoons, rocky coasts, and reefal and glaciated coasts (Woodroffe 2003). Figure 3 proposes a layered structure of the coastal forms and systems organization in terms of movement and position (static and dynamic elements, terrestrial, marine, and interface features). The relationship between all the physical processes, hydraulic and



Shoreline and Coastal Terrain Mapping, Fig. 3 Flowchart and structure outline of the cartographic elements to be considered in shoreline and coastal mapping

meteorological conditions, geo-factors, and coastal dynamics, with all the elements, forms, and systems can lead to the production of several thematic maps of coastal geo-engineering and geosciences, as well as a better understanding of the coastal morphodynamics.

Morton (2005) justification regarding coastal mapping strategies is equivalent to this section about cartographic strategies. As stated by the author, maps of shores and coastal terrain have different applications and purposes. The current overview presented herein gains new importance by showing several dimensions of information, where it is also possible to visualize multilevels, multilayers of data in order to support coastal management decisions and strategies, as well as to ensure the role of stakeholders' actions and/or public policies. Accordingly, considering that there is a long history of coastal processes and landforms, Davidson-Arnott (2010) and references herein sustain the importance of coastal geomorphology and how this research is significantly influenced by other fields of science. The overlapping of coastal features, different fields of actions and sciences, and the multi-interdisciplinary approaches are the focus and the core of the current vision that also couples two directions: past knowledge and know-how.

This approach of the architectural or framework prism for shoreline and coastal terrain mapping is of highest importance (Fig. 3). The cartographic elements comprising features, forms, processes, and dynamics are the starting point for shoreline and coastal mapping.

Coastal Mapping Applications

General Remarks

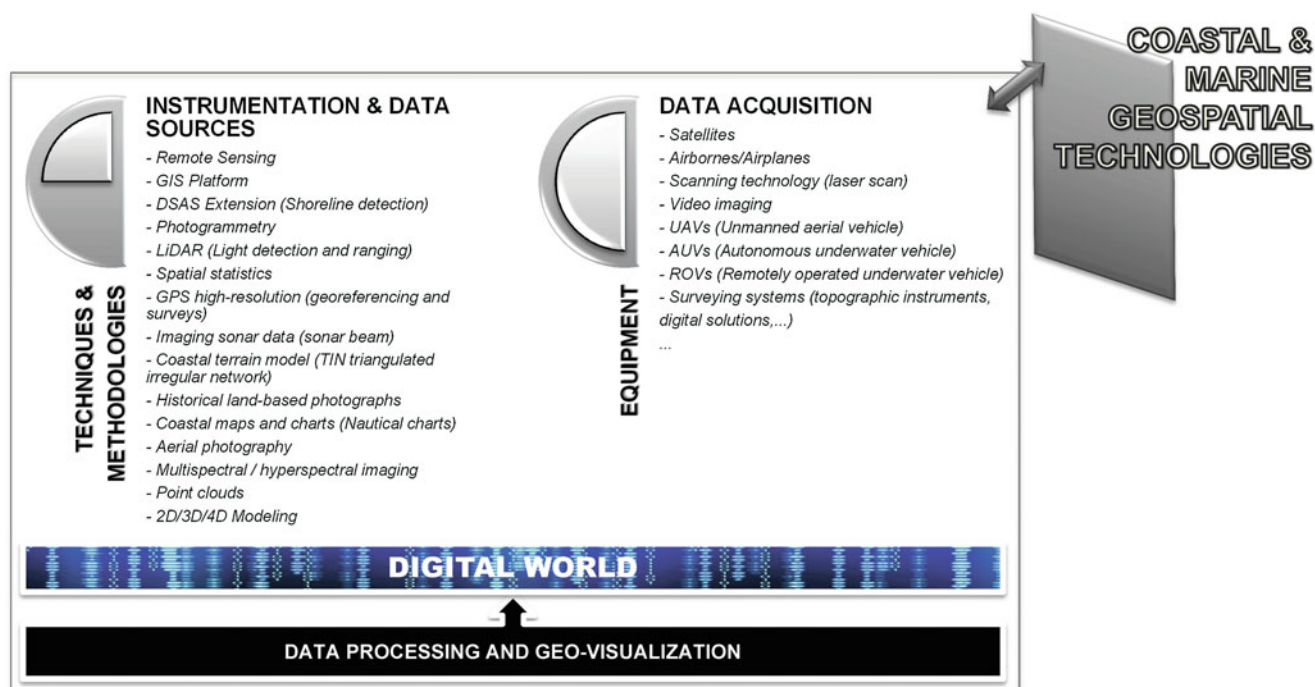
The development of cartography and maps is related to the progress made in military operations, geography, geology, topography, and engineering, in broad terms. In military operations and war history, there are important scientific basis, methodologies, and techniques in cartography that have developed over the years. The use of road or strategic maps in battles, from Ancient Rome to the World Wars and up to today, meets the demands of military or administrative organizations. Throughout time, these maps or charts were all created by skilled surveyors, military geologists, and military engineers using up-to-date instruments and surveying principles, to improve military efficiency and to support military strategies. It is important to emphasize that the use of geologic and morphological information became common during World War I and World War II, and, for instance, a military geology branch in the United States Geological Survey was created (Haneberg 2004). The continuous growth of these research areas (topography, geography, geology, hydrogeology, engineering, GIScience, remote sensing, photogrammetry, etc.) over the years has allowed cartographic methods and mapmaking to expand to other fields of action and to higher levels of design, visualization, outputs, and complexities.

It was also during and after World War II that global mapping of shore types presented high relevance because of essential military applications (Morton 2005). There are many types of classification which have been applied to coasts in

attempts to characterize dominant features in terms of modes of evolution, geographic occurrence, and physical or biological properties (Finkl 2004 and references herein). According to Finkl (2004), the characterization or classification of natural coastal environments was wide-ranging in scope but needed specificity, while other specialized systems were narrowly focused, providing irregular coverage of genesis units for coastlines of the world. As a result, Finkl's research integrates a systematic approach to consider in the development of a comprehensive scheme for coastal classification. The increasing availability of GIS information and frameworks, particularly digital formats, endorses integrated and systematic approaches to coastal classification. It is required to apply sophisticated solutions to correspond and interconnect problems in the littoral zone. Climatic factors can determine the nature of some coasts, but there are many other parameters necessary in order to categorize the so-called geomorphic units (Trenhaile 2014). The proposed approach by Finkl (2004) employs discriminating criteria for hard rock and soft rock coasts that are related to the antiquity of littoral landforms and divided by chronometric parameters. Finkl's work is an excellent study in terms of its literature review. Also, it is important to draw the attention to the conception of systematic approaches, which are one of the principles of the current perspective concerning shoreline and coastal terrain mapping.

Coastal and Marine Geospatial Technologies

The second view of the prism shows an extensive outlook going through several techniques, methodologies, and equipment which integrate the geospatial technologies for coastal and marine applications (Fig. 4). All the views of the prism complement each other, but in this specific angle, it is important to mention the fact that is established in the top and in the base of this geometric framework, particularly the cartographic methods/techniques and the coastal mapping. There is an innovative domain arising, even more integrative which takes into consideration the strong influence and dependence of digital world. As defined by Cartwright et al. (2007), multimedia cartography is best defined as a world atlas, which gathers in book format a collection of maps and is a reference to people. *"The atlas has been a window to the world for millions of people. It is consulted when one needs to know where something is located or something about a region of the world. The atlas forms the basis for how people conceive the world in which they live"* (Cartwright et al. 2007). The atlas is the computer, image, and maps. The World Wide Web (WWW) broadcasts geospatial data, and according to only few, the geosciences field has not (yet) found a use for the WWW. In this process a map plays an important and emerging role and has multiple functions. Moreover, to search and publish geospatial data, it is required to have an acceptable geospatial infrastructure. Currently,



Shoreline and Coastal Terrain Mapping, Fig. 4 Diagram of the covering technologies applied in coastal data acquisition, processing, and geo-visualization

maps are also applied to navigate, to map cyberspace, and to map websites.

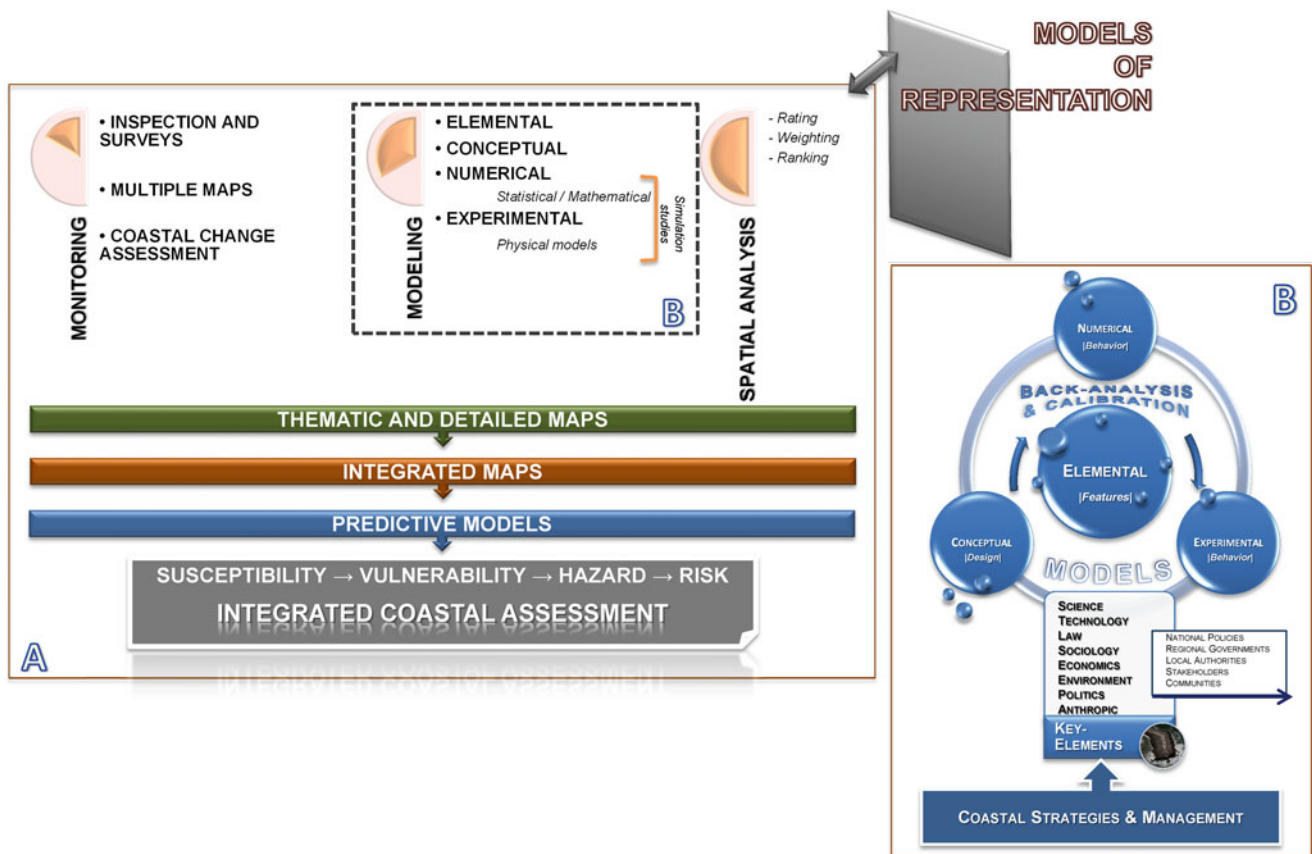
The perspective displayed in Fig. 4 establishes the relationship of the necessary procedures to optimize the applications of geospatial technologies: (i) data acquisition using different techniques or methods and equipment, (ii) processing all the information, and (iii) geodata visualization. In the end, it will be possible to enable different types of angles for this side of the prism and to see the whole picture toward a digital world, the ultimate driver of innovation.

Integrated Coastal Assessment

Models of Representation

Finally, the framework prism for shoreline and coastal terrain mapping is closed with the third view concerning models of representation (Fig. 5a). The scheme shows three categories that work together in order to produce thematic and detailed maps: monitoring, modeling, and spatial analysis. These integrated maps allow creating predictive models to support coastal assessment. The stratified framework adopted for the integrated coastal assessment shows the first category regarding monitoring process that comprises inspection and survey

operations and creation of several maps to estimate the shoreline and coastal change. The second category is related with modeling and its development. Figure 5b reveals how these types of models are interrelated and how the following four types of development models are applied in research studies with general approaches and methods. From the *elemental ground model* (e.g., fieldwork and desk studies), *conceptual site model* (e.g., geologic and/or geomorphological models with data parameters, coastal evolution and morphodynamics models), *numerical model* (e.g., based on mathematical modeling, i.e., probabilistic, deterministic, or stochastic approaches) to the *experimental modeling*, the research methods along the investigation development are shaped (e.g., Griffiths and Stokes 2008, and references therein). All the models must be robust, calibrated, and supported on a permanent back-analysis scale based on a logical understanding of the real ground behavior (Chaminé et al. 2016). The modeling types are displayed as a cyclical process of four types of models which are characterized by their role to support coastal strategies and management. These roles represent what the models visually indicate in terms of research; elemental ground models express the features, conceptual site models express the design, and numerical and experimental models express the behavior. Ultimately, when developing a



Shoreline and Coastal Terrain Mapping, Fig. 5 Methodological approach and models of representation for the integrated coastal assessment

project of a given study area, the balanced approach is an integrative and combined modeling. It is important to reach the experimental modeling and then return to the field to verify the assumptions and corroborate the results. And then... it starts all over again. It is important to highlight the key-element inclusion on this strategic system (Fig. 5b) that influences several types of societal contributors in shoreline and coastal management (details in Pires et al. 2016).

The last category displayed in Fig. 5a concerns the spatial analysis necessary to improve the models of representation rating, weighting, and ranking the available data. According to Hamylton (2017), the concept of spatial analysis has geographical foundations. Furthermore, quantitative geography can also be defined through analysis of numerical spatial data, conceptualization of mathematical models of spatial processes, geostatistics, numerical data analysis, and map algebra, including error and uncertainty factors in data collection and model design (Hamyton 2017 and references herein). Spatial analysis has evolved over time and has become an extension of several types of technologies to improve the knowledge of coastal systems. Moreover, with this technique is possible to gather higher resolution data from remote sensing and global positioning system (GPS) information, to process, integrate, and visualize coastal data (Hamyton 2017). The use of GIS application and platform has been applied to support the studies for assessing the environmental processes operating in coastal environments, acting like a data handling tool very powerful for modeling and spatial analysis.

Global Synthesis

Models of representation perspective together with the rest of the edges of the framework prism for shoreline and coastal terrain mapping allow the “edification” of the integrated vision for coastal environments. This vision also includes the basis and top edges, cartographic methods/techniques, and coastal mapping, respectively, as substratum to different applications in shoreline and coastal research.

Therefore, morphostratigraphic models and genetic stratigraphy referenced by Morton (2005) could really act as geo-history of the coastal region. The morphostratigraphic mapping outlines the geologic history, the correlation between the surface features and landform, slope, soil composition, or other factors, as well as recognized the genetic implications of sedimentary deposits including marine, fluvial, glacial, or basinal (Morton 2005 and references herein). These types of models could have a wide range of coastal applications and approaches, proposing streamlined methodologies for the construction of stratigraphic frameworks best suitable to the available data for coastal evolution assessment (e.g., Carter and Woodroffe 1997; Goudie 2004; Burt et al. 2008).

Maps of coastal terrain are also used to assess the environmental resources and managing impacts in a sustainable approach. These maps concern with the biological resources, ecosystems, natural resources, or even with specific or thematic maps, like bird colonies spatial analysis (Otero et al. 2018). Morton (2005) also refers to maps of typical geological resources that can include sand, gravel, beach nourishment, or mineral deposits that are exploited. Mining of coastal materials is discussed in another section of this encyclopedia. However, the development of mapping technologies to gather information about the environments of where these minerals occur is important to emphasize in this section. Especially because this sort of coastal models can help to support managing impacts assessment and analysis in deep-sea mining projects which is becoming gradually an important expansion field (e.g., MIDAS 2017; Sharma 2017) and related to quarry exploitation to use rock boulders in hydraulic defense structures or integrate ecological principles with the design of marine infrastructure (e.g., Pires et al. 2009, 2014; Daffor et al. 2015).

Maps of shoreline and coastal terrain can serve as valuable geo-databases framework and support coastal monitoring. This type includes large digital databases where it is possible to analyze and process historical data in terms of shoreline change and coastal evolution (e.g., Pires et al. 2009, 2016). One of the most important parameter or geo-indicator to document coastal changes is the shoreline position. Its definition and recognition are an important requirement to produce coastal maps to investigate environmental changes. According to Boak and Turner (2005), shoreline delineation comprises two steps: (1) selection and definition of a shoreline indicator feature and (2) detection of the chosen shoreline feature within the available data source. It is possible to monitor shoreline changes from high-resolution aerial imagery, but, not only that, potential data sources can include historical photographs or charts, coastal maps, beach surveys, in situ GPS measurements, and a variety of digital elevation or image data resulting from remote sensing technologies, particularly terrestrial laser scanning (TLS) (more details in Boak and Turner 2005). But coming back to the horizontal shoreline change and how to analyze it, the Digital Shoreline Analysis System (DSAS) is included from the range of software developed and available for this purpose. In accordance with several authors (e.g., Pires et al. 2009; Pérez-Alberti et al. 2013; Hamylton 2017), DSAS extension is a very user-friendly and powerful tool currently available for measuring the shoreline displacement. It was originally developed by Thieler and Danforth, and the latest version of DSAS could be freely accessed in US Geological Survey website (Thieler et al. 2005). DSAS is also useful for data sets that use polylines as a representation of a feature's position at a specific point in time (e.g., limit of a glacier, river channel boundaries, land use, and land cover maps). The extension has included

statistical methods, namely, end point (EP), linear regression, time-series analysis, and geostatistics, which allow the assessment of annual erosion rates and the prediction of shoreline positions.

Finally, coastal zone management and resource evaluation outlooks are related with specific mapping applications that include a much wider approach. These mapping applications can be defined as results or outputs of the concept described as integrated coastal zone management (ICZM). Therefore, it is important to show some examples of practical and real projects that present a global vision in terms of coastal management and based on shoreline and coastal mapping.

During the years coastal scientists have tried to understand the shoreline in relation to the processes that shape it and its interrelationships with the contiguous superficial marine and terrestrial hinterland environments (e.g., Woodroffe 2003; Bird 2008). Those factors encourage the need for integrated coastal zone management (ICZM), because of its possible use in identifying coastal management issues to consider in policy strategies, measures, and planning.

Many definitions or explanations of what ICZM (or ICM) is can be found in publications, but a balanced definition was stated by Cicin-Sain and Knecht (1998): *“Integrated coastal management (ICM) can be defined as a continuous and dynamic process by which decisions are taken for the sustainable use, development, and protection of coastal and marine areas and resources. ICM acknowledges the interrelationships that exist among coastal and ocean uses and the environments they potentially affect and is designed to overcome the fragmentation inherent in the sectoral management approach. ICM is multi-purpose oriented, it analyzes and addresses implications of development, conflicting uses, and interrelationships between physical processes and human activities, and it promotes linkages and harmonization among sectoral coastal and ocean activities.”*

According to Post and Lundin (1996) theoretically, a wide number of sciences are involved in integrated coastal zone management, including law, oceanography, sociology, economics, planning, geology, geography, physics, ecology, chemistry, and engineering. Hence it is possible to say that ICZM is a holistic approach to researching and studying coastal areas in every scientific branch, including the participation of different entities and to ensure public awareness.

Consequently, taking into account all these considerations, the following examples display three types of shoreline and coastal data: (1) geodatabases, archives of mapping or charts information, modeling, and spatial analysis; (2) susceptibility of the coast to sea-level rise, coastal vulnerability maps, and coastal and shore classification; and (3) maps related with climate change risk assessment, hazards, resilience, and littoral management. The implementation of an integrated approach requires planning decisions. In addition, all the projects exemplify cases where it is necessary to establish

“rules,” policies, and strategies which will handle problems or future interventions in coastal areas through basic concepts, guidelines, and theoretical procedures by putting them into practice.

The Coastal and Marine Geology Program (CMGP) of the US Geological Survey displays regional interactive maps but also specific research project maps comprising a large set of information available and enabling digital data (more information in USGS 2017a). Similarly NOAA Raster Chart Products is a global map viewer and database catalogue of all the charts made available by the National Oceanic and Atmospheric Administration (more details in NOAA 2017). The USGS coastal vulnerability index provides a preliminary outline of the relative susceptibility of the US coast to sea-level rise using a coastal vulnerability index (CVI), which ranges from low to very high (USGS 2017b). In addition, susceptibility, vulnerability, hazard, and risk maps and models are important quantitative attributes or indexes that are generated (with the support of GIS platform, numerical analysis or other coastal mapping techniques and applications, as well as fieldwork campaigns) in order to classify coastal terrain (e.g., Morton 2005 and references therein; Pérez-Alberti et al. 2013; Pires et al. 2016). But always keeps in mind the accuracy and reliability problems. The SWOT satellite mission is related with the Surface Water Ocean Topography (SWOT) research (for further details see Grosbois 2017 and NASA 2018). SWOT will bring innovative technologies and approaches with an extraordinary resolution that will enable researchers to measure how water bodies change over time (Grosbois 2017). According to NASA (2018), the gathering data will have a wide range of applications, namely, flood modeling and mapping, hydrodynamic models, water management, water security, spills and pollution management, marine safety, marine transport, sea ice forecast modeling, ocean circulation models, heat content studies, ocean acoustics, and bathymetry studies.

Finally, the two European examples are related with maps related with climate change risk assessment and littoral management, supporting local authorities, policies, and stakeholders. The coastal data, results, and outputs of the POLGalicía (2010) project support and contributed to the master coastal management plan carried out by the local authorities (Xunta de Galicia, Consellería de Medio Ambiente, Territorio e Infraestructuras). Such a monitoring plan suggests some changes in coastal policy and encourages future trends in land use and coastal engineering issues. This necessarily leads to resilience processes that must be part of the coastal and maritime environment and dependent on sustainable coastal management (more details in Pérez-Alberti et al. 2013). In comparison, the coastal change assessment made in Dynamic Coast (2015) aims to create a solid database, to improve sustainable coastal and terrestrial planning

strategies, to address climate change, as well as to model coastal erosion susceptibility maps (more details in Fitton et al. 2016). Furthermore, Dynamic Coast (2015) generated shoreline management plans (SMPs) in short studied sections along the coast of Scotland. The project is still ongoing with the next phase starting in 2018, comprising several actions: categorize the resilience of natural coastal defenses (dunes), evaluate studies to understand how future climate change may intensify erosion on soft coast, and apply innovative monitoring techniques to assess natural resilience and shoreline vulnerability (Dynamic Coast 2015).

Future Trends and Coastal Perspectives

There is always the feeling in relation to coastal studies that it is an unfinished work and that it is always possible to improve some aspects of the research. Consequently, it is crucial to understand the importance of coastal systems and processes involved. Therefore, the coast is a dynamic environment with a history of change and which will continue to change in the future. In fact, the relationship between all the processes, elements, features, and forcing conditions is essential in order to produce several ground or thematic coastal maps, as well as a better understanding of the coastal evolution and morphodynamics. The current vision of shoreline and coastal terrain mapping has presented some considerations regarding the past definition (Morton 2005) in order to reflect the advance of technologies and methodological approaches in the field of marine sciences, coastal engineering, and terrain mapping. It began with an integrated vision for coastal environments, with multi-interdisciplinary and holistic perspective. As stated in Fig. 2, the different edges of the prism displayed previously are all inherently related: (i) cartographic elements (features, forms, processes, and dynamics), (ii) coastal and marine geospatial technologies, and (iii) models of representation. The geometric figure is also sustained in a basis of cartographic methods and techniques but covered by coastal mapping applications.

Modern cartography points out that the concern with design begins to be significant, and even though GIS dominion has ignored design in the past, there is an awareness that a map must communicate. The integrated coastal assessment section, earlier described, lists the models of representation as well as the possible outputs: thematic and detailed maps, integrated maps, and predictive models. The integrated coastal assessment comprises incorporating and communicating the results. In recent years, great progresses have been made in cartography and mapmaking, but the most amazing step is the level of world projection of maps. Coupling media with an interactive linking structure is feasible to achieve multimedia cartography, enhancing the user experience.

The EU societal challenges have identified as priority targets the climate action, environment, and resources efficiency. Biodiversity and ecosystems are also under stress, and the investment in research and innovation can have a real influence in society, supporting decision-making and defining coastal strategies. Marine technologies innovations, digital maps, and the outputs are promising tools available to coastal scientists, which can generate and produce several shore and coastal models to evaluate the changes and assess evolution under the framework of climate variability.

The current definition follows the trends of coastal studies, spatial analysis, and modeling to embrace an interest make-over in maps and considered cartographic representation along with the mapping process. Figure 6 shows the image of the “word world map” with current vision presented herein. It is interesting to observe that in several ideas related to shoreline and coastal terrain mapping, the words that stand the most are maps, coastal, models, data, and processes but also, environment, technologies, analysis, strategies, management, modeling, integrated, techniques, digital, terrain, and vision.

The digital mapping, geo-visualization, and GIScience, which should include all aspects of visualizing geospatial information, work together in a continuous evolution system (Hennig 2013). Geographic information systems (GIS) had an important role in coastal research and represents an outstanding technology that is in constant development and is basically established as a mapping science. According to Goodchild (2010), GIScience is also an emerging concept with a wide field of applications. Therefore, the definition presented here assumed a multidisciplinary and holistic context (e.g., Pérez-Alberti et al. 2012, Pires et al. 2014, 2015, 2016) that can be described as an integrated coastal and geosciences approach applied to dynamic maritime environments.

The influence of geospatial technologies in cartography is important to mention: the progress of information, communication, techniques, equipment, as well as the geographical understanding. It is becoming increasingly necessary to adopt a “customized” cartography and mapping. This includes multi-interdisciplinary maps with several scientific purposes to be used by different specialists. Consequently, there was a need to shape and develop new cartographic elements, which inevitably involves interaction with the real world. The link between the current maps and the final users or the practical applications of these maps is important to discuss, because basically maps should have a dynamic context with concrete purposes. This could be the future of terrain mapping: real-time cartography, automatic processes, mobile maps, and working apps. Static maps are not enough in a demanding digital world, continuously evolving in a smart technology approach. This continuous growth momentum of marine technologies enables the expansion and

- Finally, multidisciplinary and interdisciplinary studies have many advantages, especially in developing holistic methodologies in broad coastal projects under climate variability framework.

The next generation of coastal researchers, engineers, and practitioners will have many challenges ahead. There are still many undiscovered and incredible technologies related with shoreline and coastal terrain mapping to be unlocked, developed, and improved. The coast and the sea are aimlessly so blue.

Cross-References

- Beach Erosion
- Beach Features
- Changing Sea Levels
- Coastal Changes, Gradual
- Coastal Changes, Rapid
- Coastal Management Practices
- Coastal Subsidence
- Coasts, Coastlines, Shores, and Shorelines
- Erosion: Historical Analysis and Forecasting
- Geographic Information Systems (GIS)
- Global Positioning Systems (GPS)
- Holocene Coastal Geomorphology
- Monitoring Coastal Geomorphology
- Natural Hazards
- Nearshore Geomorphological Mapping
- Remote Sensing of Coastal Environments
- Remote Sensing: Wetlands Classification
- Sea-Level and Climate Change
- Wetlands

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Shoreline Response to Littoral Drift Barriers

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Definition

We consider in turn, firstly, examples of shorelines where only natural processes are operating; secondly, shorelines with artificial barriers and whose influence extends both up- and downdrift; and thirdly, shorelines where there is a complex interplay over time between natural processes and artificial interventions, particularly associated with tidal inlets. Barriers making significant reductions in the littoral drift of beach material will cause downdrift erosion, and where there is little, or no prior, consultation with property owners in the downdrift area, this can lead to claims for damages. This is a serious and worldwide problem which we need to include as a major consequence of the potential shoreline response to a barrier. Finally, we will look into the future and consider how sea level rise (*see* “article title”) and the coastal management responses to it may impact upon the problems arising with littoral drift barriers.

Shorelines with Only Natural Processes Operating

Bays kept isolated by purely natural headlands become a littoral drift cell within which the beach profile will adapt to the general form of a logarithmic spiral in response to a prevailing wind and wave direction (Komar 1998). In some circumstances, the cell can be a closed system preventing the escape of sediment, but much recent work with littoral drift cells in many regions has shown that sand by-passing of headlands is common, although the movements may be complex and require careful instrumental fieldwork to record it (da Silva et al. 2016). An example given by Ferrari et al. (2014) on the Italian Adriatic coast is particularly interesting in that sand moving around the downdrift headland is prevented from moving offshore by meadows of the sea grass *Posidonia oceanica* and thus diverted to the downdrift bay. A series of littoral cells on the southern Californian coast receive a supply of beach material from rivers and cliff erosion which then moves Southeast as littoral drift but eventually becomes trapped within a submarine canyon and lost offshore: the series of cells thus becoming a closed system (Komar 1998).

Artificial Barriers

Groynes are a traditional method of retaining a beach and combating coastal erosion. It is originally built of wood, later combined with steel, then made of concrete, and now more regularly of rock fill with the more massive designs being referred to as a strongpoint. Assessments of groyne performance have shown that generally wooden groynes perform less satisfactorily than other types, but the use of rock-fill strongpoints has to be balanced against the expense that will be involved (Pietrafesa 2012). The spacing and length of the groynes are designed to suit the beach and hydrodynamic characteristics of the locality. As well as protecting the shore against wave erosion, they may now have an additional function of maintaining a volume of beach nourishment (*see article title*). The terminal groyne of a groyne field is under threat from being undermined or outflanked landward; protection against this can be obtained by installing a short length of shore-parallel structure known as a “spur” (Anderson et al. 1983).

A wide variety of coastal structures will interfere with, and indicate the direction of, littoral drift (Fig. 1). Large structures such as jetties, breakwaters, and harbors which can have a substantial influence on littoral drift are best treated as a single entity when considering their performance as a barrier (Komar 1998). Seawalls are normally considered in respect of their response to wave action and their ability to offer passive support to a cliff face. However, seawalls act to reflect wave energy with the effects influencing wave action and beach material movements beyond the downdrift end of the wall (Griggs and Tait 1989; Bernatchez and Fraser 2012).

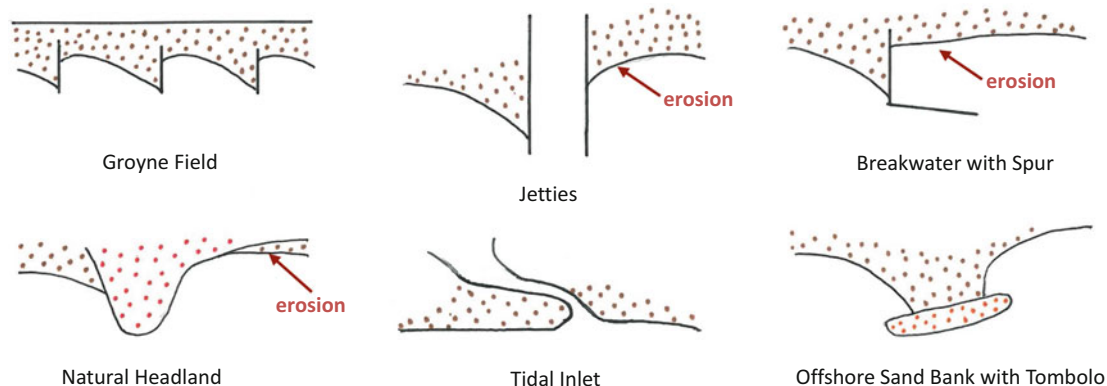
Downdrift Erosion

Where an undefended coast had a wide and protective beach consistently replenished by a large littoral drift volume of beach material, the introduction of a littoral drift barrier,

such as a groyne field or breakwater, will expose the downdrift shoreline to wave attack. Over time, the increased wave attack produces a significant recession of the coastline and will show up on maps and aerial photographs in a marked contrast to the defended coast still well-protected against erosion. Bruun (1995) has shown that as the development of the setback proceeds progressively downdrift; it releases additional beach material which can form a “bump” in the shoreline (Fig. 2) but typically with the bump proceeding faster than the setback.

An excellent example of the effectiveness of a groyne to act as a barrier to littoral drift is the long groyne constructed at Hengistbury Head, (Dorset, UK) shown in Fig. 3. In the nineteenth century, ironstone nodules, which had formerly protected the cliffs from erosion, had been quarried leaving the Headland seriously exposed to erosion. With the prospect of the sea breaking through into Christchurch Harbour (part of which can be seen in the top right hand corner), the quarrying was stopped. In 1937/1938 a long (100 m length) groyne was installed, and its value can be seen from the large accumulation of sand and gravel retained on its updrift side; however the downdrift side of the headland was then exposed to erosion, as shown in the photograph, so subsequently this part of the Headland also had to be supplied with a groyne field.

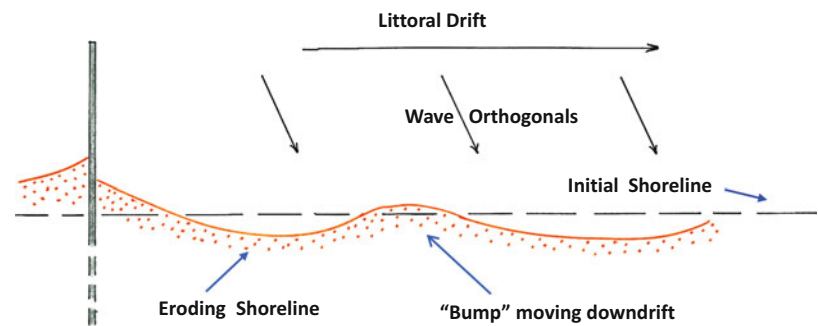
A set-back in the line of an undefended cliffed coast with respect to an updrift defended coast is a very common world-wide coastal feature with the set-back ranging from tens to thousands of meters in length. Examples from Canada are given by Bernatchez and Fraser (2012), South Korea by Kim et al. (2013), and Mexico by Escudero et al. (2014). A particularly clear example occurred when groynes constructed to augment the piers protecting the entrance to a port on the east Indian coast resulted in significant downdrift erosion and updrift accumulation (Mohanty et al. 2012). The prospect of downdrift erosion demands that the defenses are designed as part of a plan covering the adjacent areas. It becomes absurd when the occurrence of downdrift



Shoreline Response to Littoral Drift Barriers, Fig. 1 Varieties of barriers: to littoral drift – artificial at the top, natural below. The littoral drift direction is proceeding from left to right

Shoreline Response to Littoral Drift Barriers, Fig. 2

Beach profile responding to littoral drift barrier. A slight surplus of material from the eroding shore is moving downdrift. (Based on Bruun 1995)



Shoreline Response to Littoral Drift Barriers, Fig. 3

Aerial photograph of the long groyne at Hengistbury Head (Dorset, UK) showing the extensive sand accumulation updrift and reduced beach with erosion of the cliffs downdrift. (Copied from Stanley & Banks, Britain's Coastline from the Air (1986) and reproduced with kind permission of B.T. Batsford, part of Pavilion Books Company Limited)

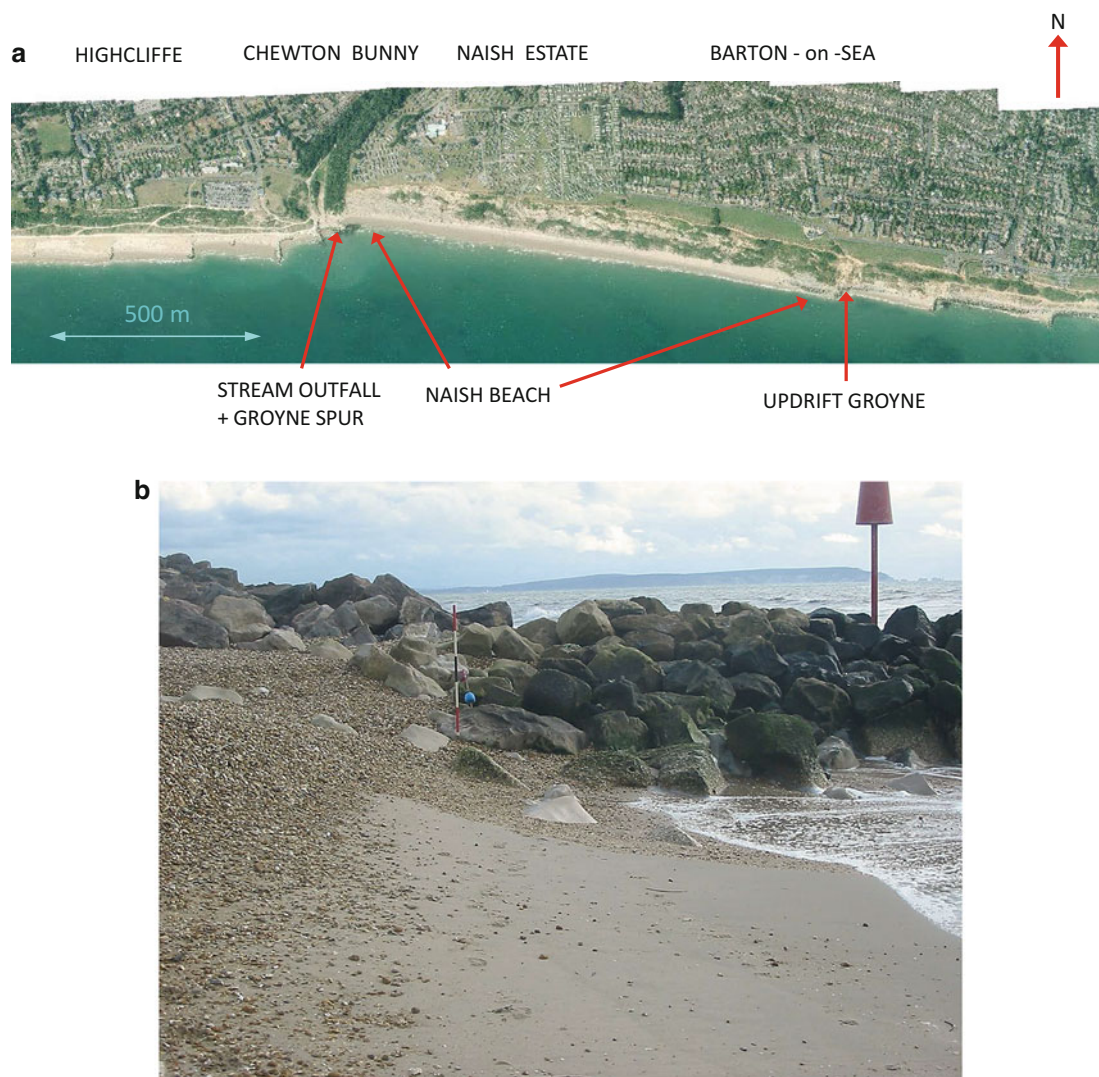


erosion is promptly followed by further defense downdrift in a piecemeal sequence as noted by Granja and de Carvalho (1995) in respect of a rapidly developing tourist area on the Portuguese coast.

A locality with excellent examples to study shoreline responses to the construction of littoral drift barriers is Christchurch Bay on the coastline of Southern England (Brown et al. 2012). It is about 14 km wide, with cliffs ranging up to just over 30 m high and composed of mainly soft, Tertiary-age clays and sands overlain by Pleistocene terrace gravels. Less than 200 years ago, much of the area was heathland with scattered farms, but two coastal towns, namely, Highcliffe and Barton on Sea developed rapidly from the late nineteenth century and onward. For protection against coastal erosion plus the retention of a good beach for their attraction as coastal resorts, groyne fields were constructed initially with

timber piles but now replaced with rock-fill strongpoints (Fig. 4a). The coast between these two towns (approximately 1.22 km) is mainly a geological conservation area where coast protection and cliff defense works are not allowed, so it is now developing as a pocket beach, with an increased rate of cliff toe erosion on the downdrift side of the Highcliffe groyne field and accumulation of beach material on the updrift side of the Barton Sea groynes. Beach material produced by erosion of the cliffs at this locality mainly comes from the sand and gravel of the overlying Pleistocene terrace, the gravel notably enriching the updrift accumulation, but the sand is subject to removal by rip currents (*see* "article title") adjacent to the rock-armored groyne, leaving a notable berm between the storm beach and the foreshore, as shown in Fig. 4b.

East of the Barton on Sea groyne field, considerable downdrift erosion and recession of the cliffs has taken



Shoreline Response to Littoral Drift Barriers, Fig. 4 (a) Christchurch Bay showing the groyne fields at Highcliffe and Barton-on-Sea, enclosing the unprotected shoreline fronting the Naish estate. Littoral drift is from west to east. (Imagery courtesy of Channel Coastal Observatory and based on photography from New Forest District Council).

(b) Most westerly rock armored groyne of the Barton-on-Sea groyne field acting as a barrier to shingle transported along the Naish Beach. Rip current action as seen in the photograph transports sand around the groyne. The ranging rod on the groyne is marked in feet. In the distance is the outline of the Isle of Wight

place (Fig. 5). At high tide, the sea is washing the base of the cliff, here composed mainly of a locked and weakly cemented Eocene sand, which forms a very steep and clearly unstable cliff face. Cliff falls of the granular material contribute to the relatively narrow storm beach. Maps showing the progressive recession of the cliff line in this part of Christchurch Bay are given in Brown et al. 2012.

Proof that a Littoral Drift Barrier Has Been Responsible for Accelerated Recession

It is clear that any plan to introduce a littoral drift barrier must take into account the effects on the downdrift coast and that

any such plan must have an agreement between all the parties that will be involved with the consequences. Where such agreement has not been established, the result can become an issue to be resolved in a court of law. The problem then becomes an issue of whether or not the constructed barrier has been responsible for the accelerated recession and whether this can be proved beyond reasonable doubt. It is necessary to allow for the all the factors which may be responsible for variations in the rate of coastal erosion over different time periods. A major one of these is the wave climate, particularly the incidence of storm waves superimposed on sea level raised by storm surge. Thus the erosion rate is subject to temporal variations which will be superimposed on the longer-term effect of a reduced beach width. It is necessary,

Shoreline Response to Littoral Drift Barriers, Fig. 5 Beach and cliff erosion downdrift of the terminal groyne (plus a spur) at Barton-on-Sea, Christchurch Bay. The sand and sandy clay strata in the Eocene Barton sand formation are rapidly receding with small cliff falls contributing to the narrow storm beach at the cliff foot



therefore, to examine the recession that has taken place over a sufficiently long time interval to reflect the influence of the barrier beyond that of the short-term effects due to storm waves.

A method to resolve this problem has been outlined by Barton and Brown (2014). This relies on establishing the recession that would have occurred had the barrier not been constructed, called the “maintained” rate of recession ($R_{\text{maintained}}$). We compare this with the observed recession (R_{observed}) but allow in both cases for the errors (E) which are inevitably involved with determining the recession, such as from map or aerial photographic evidence. The difference is the “excess” recession which is calculated as follows:

$$R_{\text{excess}} = [R_{\text{observed}} (\pm E)] > [R_{\text{maintained}} (\pm E)]$$

The observed rate of recession must be sufficiently long enough to eliminate the short-term storm effects from the calculation. An example of this approach to the problem of assessing the influence of a barrier has been made for the village of Mableton on the Holderness coast of Yorkshire (Fig. 6a).

The Holderness coast is backed by low cliffs of boulder clay which are subject to toe erosion by high spring tides. Littoral drift proceeds from north to south, but there is no significant supply of beach material from the Chalk cliffs to the north, and the boulder clay is not rich enough in sand or gravel to develop a protective beach for the whole Holderness coast. Groyne fields have been constructed to protect the

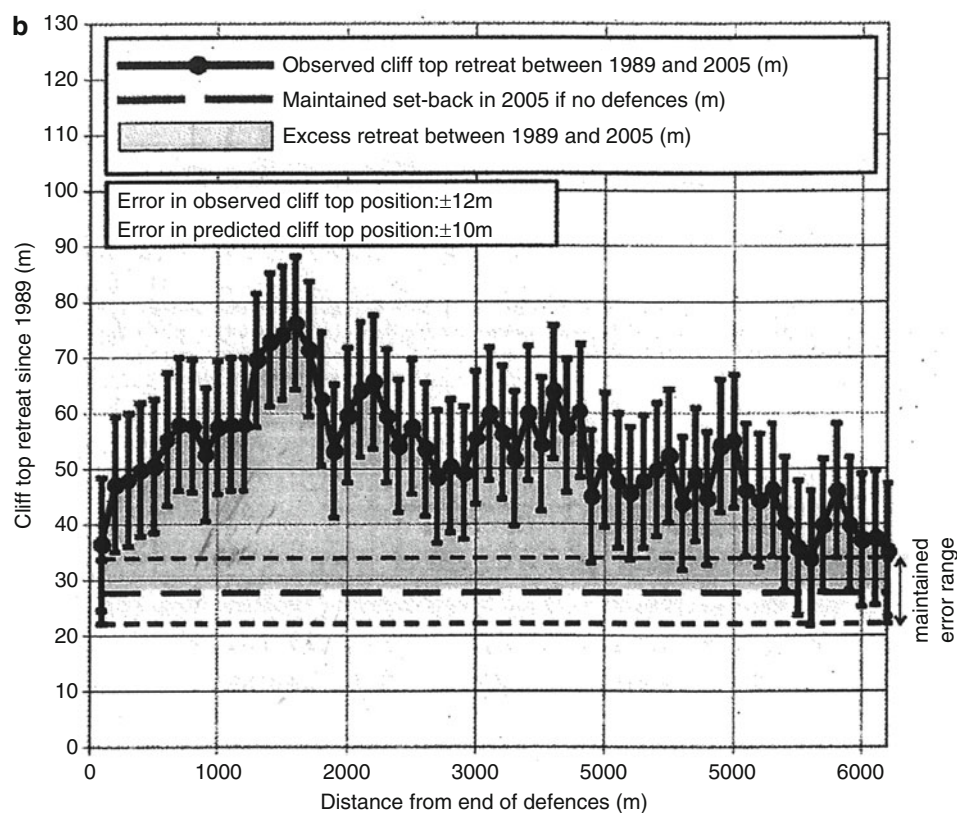
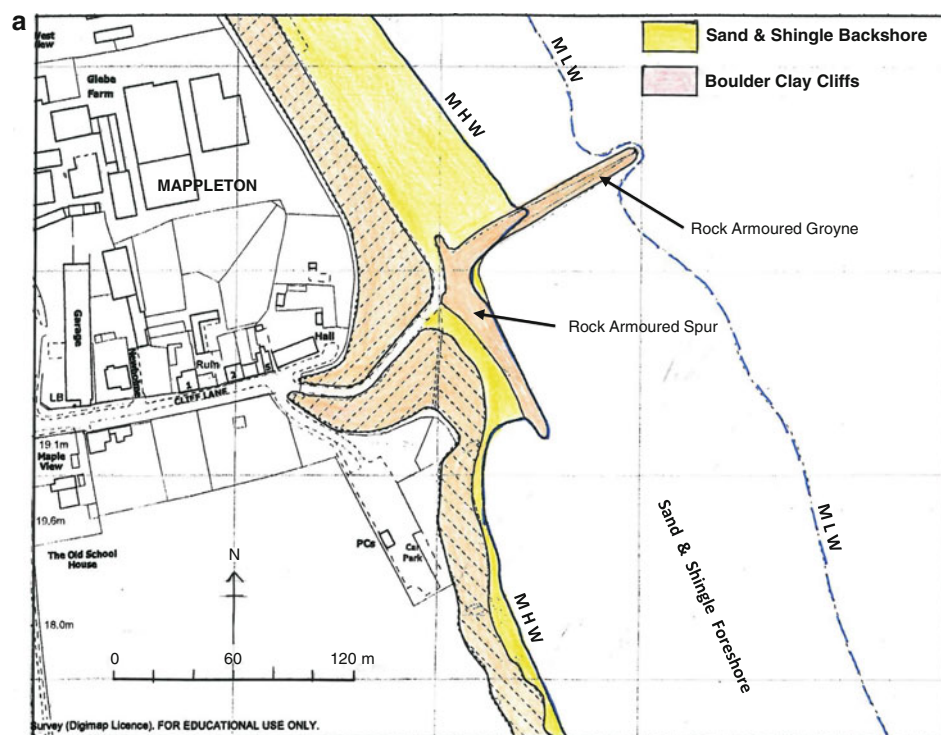
principal towns on this coast such as Hornsea at 3.1 km north of Mableton. In order to protect Mableton, 450 m of rock armoring and two rock groynes were constructed in 1991 leading to a setback downdrift. Examining the coast 4 km downdrift of the groynes, we find that over the period from 1952 to 1989, the recession rate was $1.7 (\pm 0.6)$ m/year, but from 1989 to 2005, the rate increased to $3.3 (\pm 0.8)$ m/year. We can take the former figure as the maintained rate of recession without the groynes, then the average excess retreat over the 16-year period is $25 (\pm 12)$ m (Fig. 6b). Landowners who suffered from recession 7 years after the groynes were installed requested compensation, but a tribunal argued that the area frequently suffered heightened erosion due to variable natural factors and dismissed the claim (Maddrell and Gowan 2001). Taking a longer-term view as shown by Fig. 6b, this verdict is unreasonable: one could also argue that if someone steals the beach from another landowner, then that person’s land no longer has the defense which could mitigate the worst effects of marine erosion.

A second example of the propensity of downdrift erosion to generate the level of acrimony that leads to the law court is the problem that arose as a result of coastal erosion on Fire Island, a long, thin barrier island that parallels the south shore of Long Island, New York. A tidal inlet on the south side of the island (the Moriches inlet) was opened by the Corps of Engineers and permanently stabilized with groynes in 1953. The groynes immediately caused downdrift erosion on the southwestern side, the littoral drift direction (Galgano and Leatherman 1999). An action against the Corps of Engineers

Shoreline Response to Littoral Drift Barriers, Fig. 6 (a)

Mappleton map showing beach, cliffs, and sea defenses. (Adapted from O.S., 1:2,500 Scale Digimap with NGR (copyright reserved)).

(b) Recession shown by the observed and maintained values from 1989 to 2005. Allowance for the possible errors in measurement still indicates that excess erosion has occurred subsequent to the construction of the defense works



was brought by the homeowners of the well-developed real estate on the southwest side of the island, after having experienced increased coastal erosion brought on by storms during the 1990s. However, careful study of the distribution of the

downdrift erosion showed that the increased recession decreased asymptotically away from the inlet such that there was no influence of the inlet groynes beyond a distance of 18.5 km from the inlet. Since the nearest part of the estate was

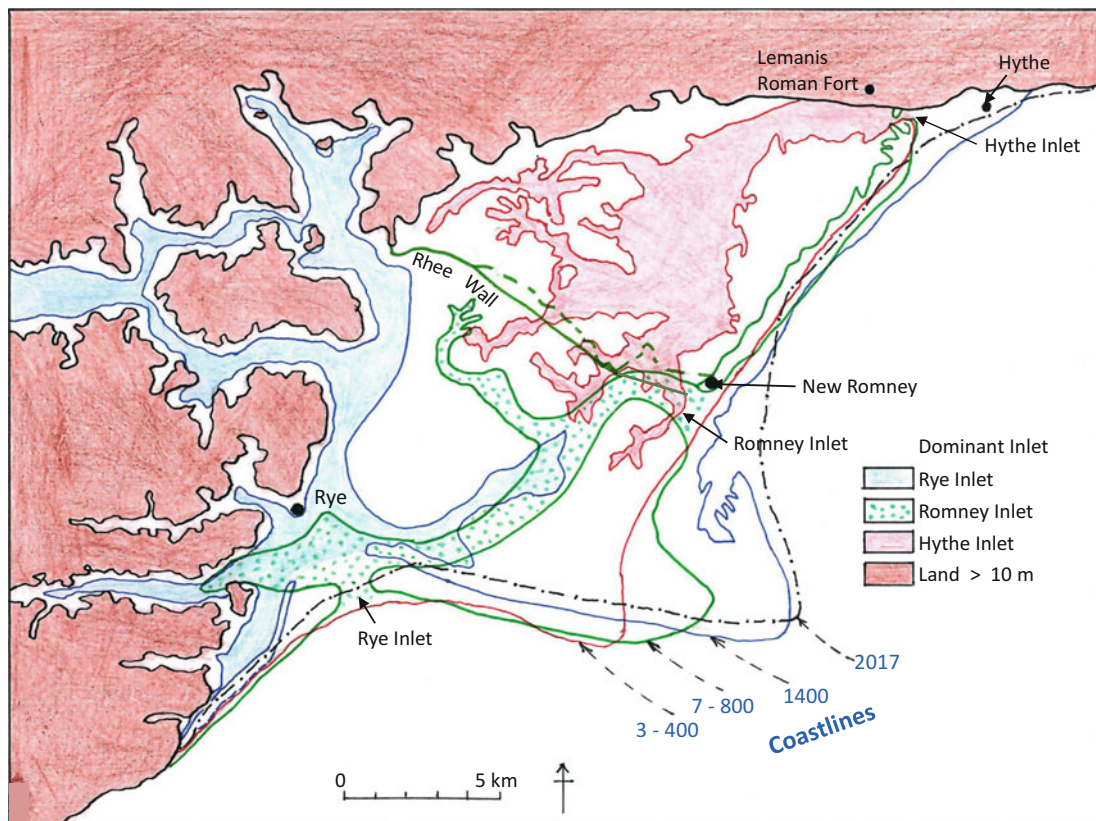
approximately 25 km from the Moriches inlet, it could not be shown that blame for the coastal erosion could be laid on the construction of the stabilized inlet (Galgano and Leatherman 1999). While it is clear that downdrift erosion is an inevitable result of cutting off the supply of beach material, it is clearly necessary to carefully examine the coastal morphology, rates of recession, and historical factors which could have influenced the result before embarking on an action for damages.

Tidal Inlets, Littoral Drift, and Shoreline Response: As Exemplified by Dungeness

The extent to which tidal inlets (see “article title”) function as barriers to littoral drift and their consequent influence on the adjacent shoreline can be considered in respect of one very well-known, large cusped foreland, namely, Dungeness. This region had three tidal inlets, at separate times during the last two millennia, in each case arising from a breach of the surrounding shingle barrier and giving rise to large floods. The inhabitants responded by controlling the floods, as far as possible, and making use of the inlet for navigation, sometimes very successfully. The long history of

shoreline response to both natural and artificial influences can be studied using historical and archaeological evidence to augment the geological over the last two millennia (Fig. 7). Dungeness is renowned as being a feature containing possibly the largest accumulation of shingle in the world (Eddison 1983).

The cusped foreland occupies a large embayment fronting a curved interglacial age cliff line, the embayment sloping down to a shore platform (see “article title”) at about 32 m below current sea level. During the Holocene rise of sea level, a sequence of shoreline materials (comprising a thin lag gravel; a series of sands, silts, and clays; and, at the top, shoreface and storm beach gravels) were deposited on this platform. The upper layer of shingle forms a series of “ribs” aligned with the present eastern face of the foreland, indicative of the littoral drift currents which eventually formed the present outline of Dungeness. The outer rib of shingle forms a storm bank which protects the mudflats and salt marshes (see “article title”) which stretch back to the abandoned cliff line. The surface elevation of the salt marshes is below high-tide level, so a breach in the shingle bank causes a large area to be flooded. In the past, there were three breaches of the shingle bank giving rise to tidal inlets. The locations of the inlets and maximum extent of the flooded area at times when each of the



Shoreline Response to Littoral Drift Barriers, Fig. 7 Map of Dungeness cusped foreland showing the three inlets and their associated flooded areas when each of the inlets was dominant. The areas shown are only approximate (based on separate drawings given in Long et al. 2006)

inlets were dominant have been estimated by Long et al. (2006) and shown in Fig. 7.

Hythe was the earliest tidal inlet, probably of late Holocene origin, and used as a port by the Romans. The extent of the flooded area subsequently increased with much of Romney Marsh inundated at its maximum extent (3–400 AD) is shown in Fig. 7. Increased occupation of Romney Marsh indicate that tidal waters were retreating in Saxon times. The closure of the Hythe inlet appears to have occurred during the development of the Romney inlet suggesting that northward littoral drift was responsible, with the Hythe inlet finally being sealed by blown sand. The Romney inlet provided good maritime access during Saxon times allowing Romney to become one of the *Cinque Ports*, while reclamation and training walls kept the inland flooded area under control as shown (for 7–800 AD) in Fig. 7. However, siltation of the inlet was threatening the success of New Romney, so a new 12 km watercourse (subsequently known as the *Rhee Wall* owing to corruption of the early English word for water) was built in three stages and combined with a series of sluices to flush silt and sand from the channel at low tide. Records show that it was in use from the mid-thirteenth century to the early fifteenth century before nature finally won and the inlet closed.

The third tidal inlet (the Rye inlet) became very large after a series of storms made a substantial breach in the shingle barrier, with the flooded area turning Rye into a major port as shown for 1400 AD in Fig. 7. However, siltation upstream and gravel accumulation, probably as a result of littoral drift and storm surges, over the period from 1400 to 1600, reduced the navigable area, leaving Rye as it is today: a relatively minor fishing and sailing center. Access to Rye is maintained by means of an embanked estuary of the river Rother. Diversion of the shingle from the barrier beaches and into the estuaries made by the inlets left fundamental weaknesses to the shingle barrier, so 45% of the Dungeness Foreland is now protected by seawalls (Eddison 1983). Elsewhere, the shingle bank is frequently controlled by earth moving to maintain suitable slopes and height, especially in the vicinity of the nuclear power station at the southeast end of the foreland. The final conclusion concerning the benefits of the inlets must be that for lengthy periods; they provided good navigable facilities but ultimately were doomed by the lack of control of littoral drift and by the inadequate outflow from the marshlands during the ebb tide.

The Influence of Littoral Drift Barriers on the Shoreline Response to Sea Level Rise

Allowing for an average global rise no greater than say 0.5 m by the end of the current century (accepting the more modest IPCC forecasts), this still indicates more wave energy being

expended onto shorelines and the prospect of an increase in the rate of marine erosion. Coasts bereft of beaches will recede faster, while protected coasts, given that the protection is modified to cope with the additional wave height and energy, will form new (artificial) headlands. The new headlands will require extra protection along their sides to prevent outflanking. The implementation of *managed retreat* (see “article title”) as used in coastal management strategies will need to allow for the increased costs of maintaining new headlands. The long-term impact of sea level rise and the need for more studies replete with computer modelling are emphasized by Vitousek et al. (2017).

In designing a strategy for the coast which will include hard protection where needed and soft engineering measures elsewhere to allow for natural geomorphic processes to operate, it is necessary to make the choices clear to the public at large. Cooper and McKenna (2008) point to the public misunderstanding that arises when coastal work is being carried out, and an assumption is made that the result will be for complete protection against marine erosion, whereas the objective was to produce a coast in harmony with the natural processes: one where nature is doing the work and producing a coast in equilibrium with the natural forces.

Cross-References

- [Beach Nourishment](#)
- [Changing Sea Levels](#)
- [Managed Retreat](#)
- [Rip Currents](#)
- [Salt Marsh](#)
- [Sea-Level and Climate Change](#)
- [Shore Platforms](#)
- [Tidal Inlets](#)

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Simple Beach and Surf Zone Models

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Definition

Beach and surf zone models refers to conceptual models which provide some basic insight on the dynamic interaction between waves and beaches. In this entry, only simple aspects of these models will be discussed.

Background

The physical processes that mold beaches and that are at work in the surf zone are anything but simple. The best scientific minds that study these processes often disagree on what are the important cause and effect relationships at work in this complex and dynamic area. However, there are some relatively simple models which are useful for understanding important aspects of how beaches respond to waves, such as erosion and accretion. Beach erosion falls into two general categories: (1) When the littoral transport of sand, or possibly other sediment, to the beach is less than the littoral transport of sand from the beach, that is, a net loss in the longshore transport of sediment. (2) When sand moves from the subaerial (above water) beach to the submerged beach or the movement from the submerged beach to the subaerial beach. It is this latter category of offshore and onshore sand movement, referred to as cross-shore sediment movement, which will be regarded as erosion or accretion, respectively in this entry.

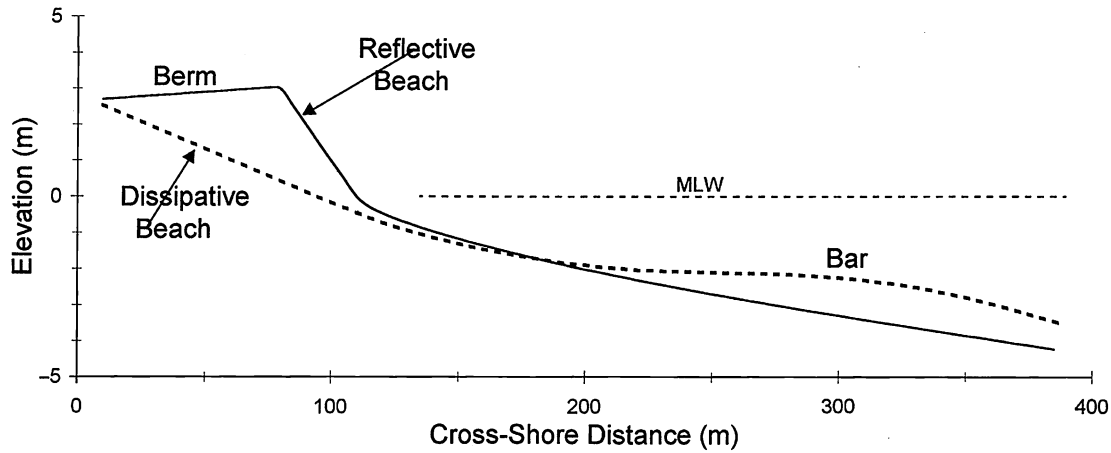
Summer Beach–Winter Beach Model

The simplest model for onshore and offshore sediment movement is commonly referred to as the summer beach–winter beach concept. By the 1950s, coastal scientists became aware of a strong seasonality of beaches in California. Generally, during winter storms waves would remove sand from the subaerial beach and deposit it offshore typically in bar formations. Milder wave conditions, during summer, would move the bars onshore and ultimately they would attach to the shore and rebuild the wider berm associated with the “summer beach.” Figure 1 shows the profiles of a reflective and a dissipative beach. Using the terminology of Wright and Short (1984) reflective and dissipative beaches correspond roughly to summer and winter beaches, respectively. In interpreting Fig. 1, note that there is very high vertical exaggeration.

Weather patterns in many parts of the world undermine the simple concept of a summer/winter beach pattern, and even in California, the concept is not too reliable. The important insight of the concept, that wider beaches are associated with a prolonged period of mild wave conditions and narrower beaches are associated with storm waves, is true.

U_t Beach Index Model

The U_t parameter was developed by Ahrens and Hands (2000) and it can be used like an index to classify wave events on beaches as either erosion or accretion. U_t can be calculated in



Simple Beach and Surf Zone Models, Fig. 1 Typical profiles of the highest beach state, dissipative, and the lowest beach state, reflective

terms of a sediment mobility number, N_{so} , and the deepwater wave steepness, H_{so}/L_o . The mobility number is defined: $N_{so} = H_o/\Delta d_{50}$, where H_o is the deepwater wave height offshore of the beach in question, d_{50} is the median grain diameter of sediment in the surf zone, and Δ is the relative density of the sediment given by, $\Delta = (\rho_s/\rho)$, where ρ_s and ρ are the density of the sediment and water, respectively. Deepwater wave steepness is the ratio of the deepwater wave height to deepwater wave length, L_o , where $L_o = gT^2/2\pi$, T is the wave period and g is the acceleration of gravity. U_t can be defined:

$$U_t \equiv \left[(N_{so}/33)(H_{so}/L_o)^{0.57} / \exp(8.3(H_{so}/L_o)) \right]^{1/2.2} \quad (1)$$

The U_t parameter was calibrated using an extensive field data set of primarily qualitative observations of beach erosion or accretion, Kraus and Mason (1991). Data are from beaches all over the world, collected by many researchers, and include observations of 99 distinct wave events. The range of the parameter for this data set was, $0.42 \leq U_t \leq 6.56$, values of $U_t > 2.22$ were always observed to be erosion events and values of $U_t < 1.45$ were always observed to be accretion events. There was a transition region, $1.45 < U_t < 2.22$, where a mixture of erosion and accretion events occurred, but generally U_t did a good job of classifying events in the Kraus and Mason data set. Figure 2 shows the N_{so} versus H_{so}/L_o plane using the U_t parameter to determine various regions.

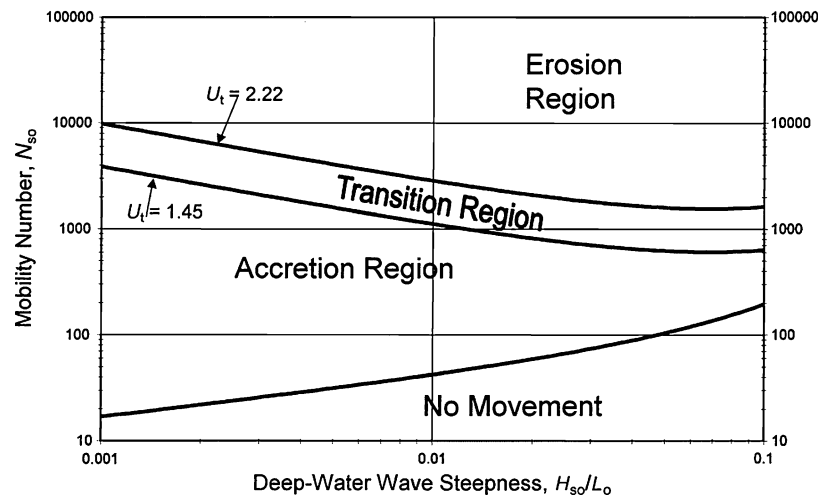
Certain features of Fig. 2 should be noted because they provide insight into the erosion or accretion processes. Highly mobile sediment, that is, high mobility numbers, is associated with erosion and low sediment mobility is associated with accretion. The accretion region shrinks and the erosion region expands in going from low- to high-wave steepness. These are trends which are consistent with current understanding of erosion and accretion processes.

Antecedent Beach Conditions (ABC) Model

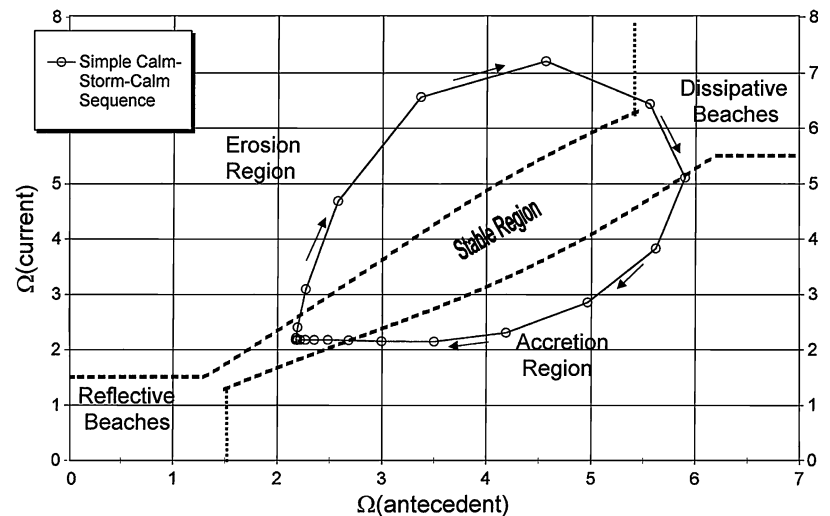
There have been a number of attempts to develop a single parameter to quantify the tendency of sand to move onshore or offshore. The most useful simple parameter is the fall-velocity parameter given by $\Omega = H/wT$, where H is the wave height and is the fall velocity of sediment in the surf zone. Some scientists have used the breaker height, H_b , on the beach to calculate Ω others have used the deepwater wave height, H_o , offshore of the beach to calculate Ω . Regardless of whether H_b or H_o is used, the principle is, if Ω is small, it indicates that sediment movement is predominately by bedload and there would be a net onshore movement of sediment. If Ω is large, it would favor offshore movement of sediment because there would be large quantities of suspended sediment which would be carried offshore by the bed return flow or, possibly, by rip-currents.

An interesting perspective on beach erosion and accretion can be gained by comparing the current value of the fall-velocity parameter, $\Omega(\text{current})$, with the antecedent value, $\Omega(\text{antecedent})$. $\Omega(\text{current})$ is the average value of Ω for the day in question and $\Omega(\text{antecedent})$ is a weighed average of $\Omega(\text{current})$ starting yesterday and including the previous 30 days, Wright and Short (1984). The function is heavily weighted toward the recent past with the previous days contributing 90% of the total weight. Wright and Short found that $\Omega(\text{antecedent})$ correlated very well with the six fundamental beach states that they had identified. The six beach states can be ranked hierarchically from the highest state, dissipative, to the lowest state, reflective. Wave conditions which move a beach toward a higher state cause erosion and wave conditions which move a beach downstate cause accretion. These beach states have some very interesting quantum aspects which suggest analogies to quantum physics. Typical profiles for the end beach states are shown in Fig. 1. A dissipative beach is the product of storm waves and has a flat beach face

Simple Beach and Surf Zone Models, Fig. 2 The U_t plane, showing erosion and accretion regions as defined by the U_t parameter



Simple Beach and Surf Zone Models, Fig. 3 Current versus antecedent beach condition plane showing a “simple” calm-storm-calm sequence



and a rather subtle bar or possibly subtle bars. A reflective beach is the product of a period of mild wave conditions and has a berm, steep beach face and no bars. Both dissipative and reflective beaches are relatively two-dimensional, although reflective beaches usually have pronounced cusps. The four intermediate beach states all have complex three-dimensional, dynamic topography, so, it is difficult to associate a single profile that is typical of intermediate states. Figure 3 shows the $\Omega(\text{current})$ versus $\Omega(\text{antecedent})$ plane adapted from the model of Wright et al. (1985). The figure identifies the highest beach state, dissipative, and the lowest beach state, reflective, connected by a stable region. The stable region trends along the diagonal where $\Omega(\text{current}) = \Omega(\text{antecedent})$; that is, when the current conditions are approximately equal to the antecedent conditions, the beach state is stable. The stable connecting region contains the four intermediate beach states defined by Wright and Short (1984). Boundaries for the stable region shown in Fig. 3 are approximate based on research by Wright et al.

(1985). Also shown in Fig. 3 are an erosion region and an accretion region. The erosion region is characterized by, $\Omega(\text{current}) > \Omega(\text{antecedent})$, and the accretion region by $\Omega(\text{current}) < \Omega(\text{antecedent})$.

To help illustrate the ABC model, a simple calm-storm-calm sequence is shown in Fig. 3. Simple indicates that an extended calm period is interrupted as breaker heights increase monotonically from small values to large values during the height of the storm and decrease monotonically to small values as the storm wanes and another extended calm period begins. The sequence shown is not specific, but is consistent with breaker heights over an inshore bar, wave periods, and sand size that occur on the east coast of the United States or the coast of New South Wales, for example, $0.25 \leq H_b \leq 1.75$ m. Circles are shown along the path of the sequence to bracket 1 day intervals and arrows are used to indicate the time order of the sequence. The sequence starts at the lower left when a calm period is interrupted by the initial arrival of waves from a storm. With increasing breaker

heights, the sequence quickly leaves the stable region and moves into the erosion region since $\Omega(\text{current})$ quickly exceeds $\Omega(\text{antecedent})$. As long as the storm intensity increases, $\Omega(\text{current})$ increases causing $\Omega(\text{antecedent})$ to increase too, this is the main period of erosion. However, at some point the storm intensity will start to wane and $\Omega(\text{current})$ will start to decrease. During this initial waning period, $\Omega(\text{antecedent})$ will continue to increase but the storm sequence will quickly leave the erosion region and drop into the stable region, and then into the accretion region. It is this initial waning period of the ABC model, when both $\Omega(\text{current})$ and $\Omega(\text{antecedent})$ are large, that is anomalous from the perspective of the U_t Beach Index model and also other beach index models. The U_t model predicts that when wave conditions are energetic, $U_t > 2.22$, erosion will occur regardless of the antecedent beach conditions; the ABC model predicts that accretion will occur when $\Omega(\text{current})$ is roughly 20% less than $\Omega(\text{antecedent})$ regardless of how large $\Omega(\text{current})$ is. These criteria can easily be in conflict during the early portion of a waning storm and current research indicates that the ABC model is the more realistic. As the storm continues to wane, the beach continues to rebuild, albeit slowly, until it returns to a stable condition similar to the condition at the beginning of the cycle.

Summary

Three models of beach and surf zone behavior are discussed. In order of increasing complexity they are: (1) summer–winter beach profiles, (2) the U_t Beach Index model, and (3) the antecedent beach condition (ABC) model. From a historical perspective, the summer–winter beach profile concept is interesting, but is certainly not a reliable model. The U_t model is a simple model to determine if beaches will erode or accrete for a given sediment size and wave conditions, but gives no information about beach profiles. The ABC is the most complex of the three models and is difficult to apply considering the level of information required. However, the ABC model is conceptually the most realistic and provides a great deal of information about beach and surf zone processes beyond erosion, accretion, and beach profiles (Wright and Short 1984). The ABC model has also been widely accepted and generalized by other coastal scientists.

One of the most difficult aspects of the ABC model is to classify beaches and surf zones in the proper state. To help classify ABC beach states correctly, Lippmann and Holman (1990) developed computer enhanced video images of the surf zone to show bar and trough locations and shapes, and proposed some additional beach states. All of the models discussed are for microtidal beaches, that is, tide ranges less than 2 m. The ABC model has been generalized by Masselink

and Short (1993) to include the influence of tides on beaches and surf zones and, therefore, to extend the usefulness of the ABC model to beaches with tide ranges greater than 2 m.

Cross-References

- [Beach Erosion](#)
- [Beach Features](#)
- [Cross-Shore Sediment Transport](#)
- [Dissipative Beaches](#)
- [Erosion Processes](#)
- [Longshore Sediment Transport](#)
- [Reflective Beaches](#)
- [Rhythmic Patterns](#)
- [Sandy Coasts](#)
- [Surf Modeling](#)
- [Surf Zone Processes](#)

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Small Islands

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Definition

What is a “small island?”

The definition of a *small island* depends upon one’s perspective. Any country in the family of nations can declare themselves a member of the Alliance of Small Island States (AOSIS). AOSIS nations include Antigua and Barbuda, Bahamas, Barbados, Belize, Cape Verde, Comoros, Cook

Islands, Cuba, Cyprus, Dominica, Federated States of Micronesia, Fiji, Grenada, Guinea-Bissau, Guyana, Haiti, Jamaica, Kiribati, Maldives, Malta, Marshall Islands, Mauritius, Nauru, Niue, Palau, Papua New Guinea, Samoa, Singapore, Seychelles, Sao Tome and Principe, Solomon Islands, St. Kitts and Nevis, St. Lucia, St. Vincent and the Grenadines, Suriname, Tonga, Trinidad and Tobago, Tuvalu, and Vanuatu; observer nations include American Samoa, Guam, the Netherlands Antilles, and the US Virgin Islands. AOSIS is a political organization of the United Nations, stemming from the 1990 Second World Climate Conference, which recognized Small Island Developing States (SIDS) as having a unique need in a changing environment.

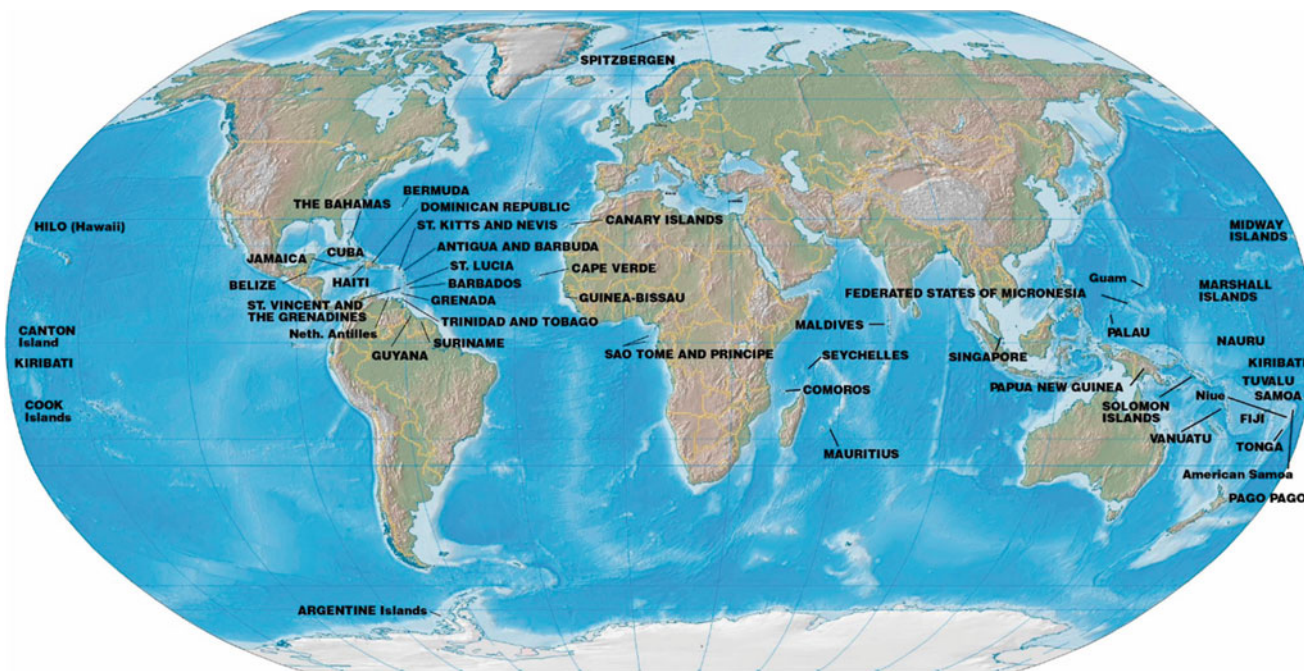
Introduction

From a geographic perspective, it is clear that AOSIS member states, and SIDS in general, are found in each major ocean, semi-enclosed, or marginal sea (see Fig. 1). Some “SIDS” are low-lying non-island coastal countries with a decidedly marine climate. They all share a number of common development issues including energy independence, coastal and marine resources, waste management, biodiversity, sustainable tourism, population growth, and climate change, among others. Their socioeconomic circumstance commonly includes a rapidly growing population and small gross domestic products with which to mitigate the effects of natural disasters and anthropogenic hazards. Most islands in

AOSIS, and SIDS in general, are positioned in low latitudes and thus are susceptible to common meteorological hazards such as tropical storms and flooding.

Geologically islands can be separated into seven island types: arc system islands, barrier system islands, coral reef islands, diapiric islands, estuarine and deltaic islands, intra-plate islands, precontinent islands, and rift islands. Many island types are associated with plate boundaries and fault zones (e.g., rift and diapiric islands); others are located at passive margins or in the coastal zone, while yet others are well within the boundaries of plates and may or may not have associated coral reefs. Their petrographic characteristics range from basalt for rift and intraplate islands to metamorphic and clastic sand and clay sedimentation for precontinent and estuarine islands, respectively, to the classic carbonates of coral reefs and the andesite of island arc system islands. Thus the cause of each island type varies considerably too, including subduction, volcanism at boundaries and within plates, coastal hydrodynamics, riverine terminus, and biological colonization upon a substrate. Other classification schemes exist, but the above serves to illustrate the diversity of geophysical settings.

Small islands also tend to have distinctly different meteorological environments. While tropical islands all tend to have sea level air temperatures in the range of 20–30 °C, precipitation patterns differ widely. The lush vegetation and high annual rainfall (>200 cm yr.⁻¹) of the islands of the western Caribbean Sea are in stark contrast to the near-desert conditions of Aruba, Bonaire, and Curaçao, barely 1000 km due



Small Islands, Fig. 1 Location of islands or island systems mentioned in the text. Member nations of the AOSIS include several non-island countries. Small Island Developing States (SIDS) and AOSIS do not define “small” in terms of area or population (See IPCC 1997)

east. Much of the interannual variability in low-latitude rainfall is related to the annual cycle of the intertropical convergence zone (the ITCZ), and this in many places is punctuated by ocean-basin scale events such as the El Niño-Southern Oscillation (ENSO). Extratropical islands have equally complex climatological variability, including extreme winter precipitation at high latitudes and the formation of sea ice. Again, no commonality exists in the biogeography of complex island ecosystems, because their environmental settings vary widely.

The physical notion of a small island then is well beyond the geopolitical perspective of AOSIS or SIDS. Humankind tends to categorize natural systems, but in the case of small islands, this is quite problematical. Setting limits on the area of an island as a definition of “small” is as arbitrary as specifying latitude or population density or biodiversity. Yet as a working definition, it might be most productive to consider their shared range of natural hazards and the common socioeconomic consequences of limited infrastructure and response strategies to events both anthropogenic and geophysical. In addition, the notion of sustainable development needs to be included in the definition as a focus for well-being. These concepts are addressed in the next sections.

Natural Hazards

Tropical and extratropical storms tend to be one of the most severe natural hazards to small islands, and yet such events are a major source of freshwater, especially in low latitudes. The West Indian hurricane, the western Pacific typhoon, and the Indian Ocean cyclone are similar tropical storms with varying degrees of severity. All coastal areas are at risk from the high surface winds, which can exceed 100 m s^{-1} with attendant rainfall in excess of 30 cm in 24 h. Extreme wind damage and flooding are common aftermaths of these storms, and it is not uncommon for years to pass before the socioeconomic infrastructure is restored. Tropical ecosystems have evolved to include the tropical storm as part of the natural cycle, and indeed it can be argued that storm events are needed from time to time to exercise the coupled land-air-sea system that defines the biological health of a small island.

Most of the toll in human terms from tropical storms is associated with storm surge and flooding. Many oceanic small islands are essentially mountains protruding above the sea surface and have very steep offshore submarine topography and narrow shelves. For storm surge height to build to the 5 m level of Hurricane Andrew in south Florida in 1992, a long and continuous coastline is necessary. Thus AOSIS countries such as Belize and Cuba are highly prone to extensive coastal flooding and loss of human life through drowning, which is the primary cause of death from tropical storms. The catastrophic loss of life, probably exceeding 250,000 persons, in

the estuarine and deltaic islands of Bangladesh in 1971, is not experienced on intraplate and island arc small islands due to the hydrodynamics of wind-stress induced storm surges. Modeling and observations suggest storm surges of 1–2 m, even in extremely powerful tropical storms, for small islands that are categorized as other than barrier islands, deltaic islands, or precontinent islands.

Many small islands are located in tectonically active regions and thus are subject to numerous earthquakes. Many are volcanically active or were in their early history. The Hawaiian Islands are a classic example of intraplate islands that owe their existence to volcanism, and it is the basaltic base of these islands that offers corals the necessary substrate to colonize and grow. The low-lying atolls of the Midway Group represent the ultimate fate of Hawaii as it drifts to the northwest over the geothermal hotspot in the central Pacific Ocean that gives it genesis and slowly subsides back into the deep. From a socioeconomic perspective, these and other tectonically active small islands require the necessary infrastructure to recover from a seismic event. It is an often-overlooked problem, especially in SIDS, to be able to recover from the aftermath of a major event, especially the control of fires, loss of electrical power, roads and bridges, and access to public health facilities for the rapidly growing human populations.

Seismic sea waves (tsunami) are another class of coastal hazard from a marine perspective. Although commonly thought of as a natural hazard of the Pacific Ocean, tsunamis are known to occur in all oceans, with historical references in the Mediterranean Sea as early as the sixteenth century BC. Most tsunami events are associated with strong, shallow focus, dip-slip submarine earthquakes, submarine or subaerial landslides or caldera collapse, submarine volcanoes, or asteroid impact. Not all of these geophysical events create a tsunami, it being especially difficult to forecast a tsunami from an earthquake alone. Small islands are particularly vulnerable to damage from tsunami events, and this has led to establishing warning systems facilitated by the Intergovernmental Oceanographic Commission of UNESCO and national agencies. In the Pacific, where major tsunami events seem to occur about every decade, an extensive array of seismometers, tide gauges, buoys, and communications systems can provide hours of warning; in the Atlantic where the frequency of occurrence is perhaps twice a century, a rudimentary warning system now exists.

In the deep sea, a tsunami travels as a small-amplitude shallow-water wave with celerity $C = \sqrt{gH}$ where g is gravity and H is water depth. A typical celerity in the deep sea is 200 ms^{-1} , and it can thus take 24 h for a tsunami to cross the Pacific Ocean, as did the wave from the 1960 Chilean earthquake. In smaller basins such as the Caribbean Sea, the total travel time will be an hour or much less if the wave is locally generated, as was the case with the Virgin Islands tsunami of 1867. As the tsunami enters shallow water, conservation of

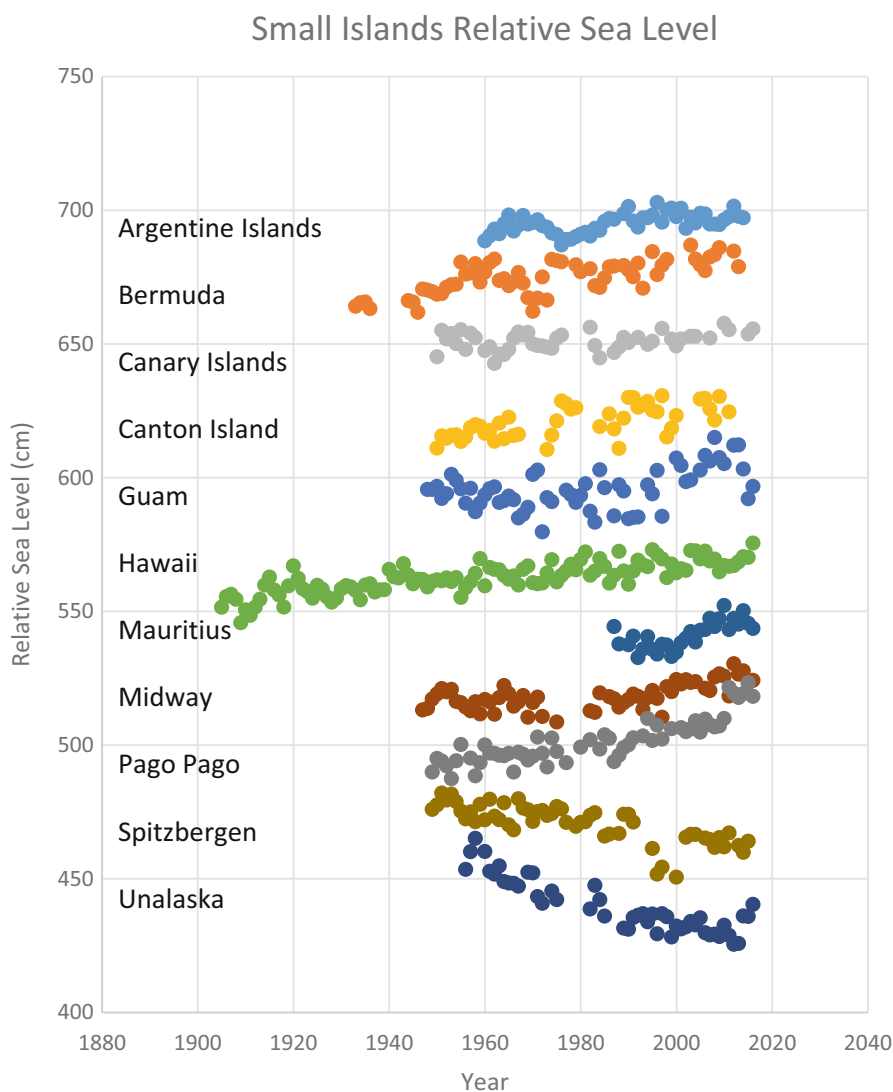
energy requires that the amplitude of the wave must increase, and this has led to the great waves, some exceeding 30 m in height, rushing in to destroy coastal communities, ecosystems, and infrastructure. While wave height is a factor in the devastation, wave run-up is often the larger issue. Much like breaking surf on a beach rushes up the beach face, tsunami run-up has been recorded to reach 490 m. Immediate evacuation of the coastal zone is the only viable course of action if lives are to be spared.

Sea level change on small islands varies as widely as the underlying geology. During the last 100 years or so, tide gauges have been operating to measure water level changes as is needed in tsunami warning systems, ENSO studies, and for determining the tides and chart datums. When the tide gauges are operated using accepted engineering practices including annual differential leveling surveys to juxtaposed benchmarks and regular visits by a trained observer, the long-term trend of relative sea level (RSL) can be discerned. It is

the change of sea level relative to the land that is of socioeconomic consequence, and the local effects of vertical land motion, steric seawater changes, oceanic currents, atmospheric barometric pressure, wind stress, and anthropogenic activities such as dredging, water and/or oil extraction, and coastal construction complicate it. Many SIDS and member states of AOSIS have expressed great concern about global sea level rise. For more information, see Pugh (2004).

Unfortunately, few small islands have information from quality-controlled water-leveling measuring systems to quantify relative sea level. Typically, a record length exceeding three nodal tidal periods (18.61 years each) is needed to statistically discern change in relative sea level above the natural interannual variability associated with long-period tides and coupled atmosphere-ocean events such as ENSO, the Pacific Decadal Oscillation, or the North Atlantic Oscillation. A survey of RSL at selected small islands shows a wide variety of trends (see Fig. 2), some showing sea level rise,

Small Islands, Fig. 2 Annual relative sea level at selected small islands. The mean of each curve is offset by 10 cm or so from PSMSL RLR data for clarity. See Table 1 for statistical summary and geographic coordinates (Updated and redrawn from Maul 1996)



Small Islands, Table 1 Summary of least squares linear trend sea level change statistics from selected small islands shown in Fig. 2. Database is the revised local reference (RLR) file of the Permanent

Service for Mean Sea Level (PSMSL), Bidston Observatory, Birkenhead L43 7RA, United Kingdom

Station	Years	Latitude	Longitude	N	Trend (cm/century)	r ²
Argentine Islands	1950–2014	65°15'S	64°16'W	54	12 ± 3	0.30
Bermuda	1933–2013	32°22'N	64°42'W	63	20 ± 3	0.51
Canary Islands	1950–2016	28°29'N	16°14'W	47	6 ± 2	0.11
Canton Island	1950–2011	02°48'S	171°43'W	49	21 ± 4	0.43
Guam	1948–2016	13°26'N	144°39'E	61	17 ± 4	0.20
Hawaii	1905–2016	19°44'N	155°04'W	112	14 ± 1	0.64
Mauritius	1987–2016	20°09'S	57°30'E	29	41 ± 8	0.47
Midway	1947–2016	28°13'N	177°22'W	60	13 ± 2	0.35
Pago Pago	1949–2016	14°17'S	170°41'W	62	33 ± 4	0.57
Spitzbergen	1949–2015	78°04'N	14°15'E	56	−27 ± 3	0.60
Unalaska	1956–2016	53°53'N	166°32'W	49	−47 ± 4	0.78

some sea level fall, and some with no statistically significant change at all (see Table 1). Complicating the RSL issue is that even close-by islands can have very different trends, either upward or downward, especially intraplate small islands, deltaic islands, and those within island arc systems. It is simply not possible with current RSL measurements to state unequivocally that small islands, or any coastal system, have an issue with relative sea level that can be attributed solely to global change, whether anthropogenic or geophysical.

Anthropogenic Hazards

The discussion of anthropogenic hazards on small islands and especially to SIDS is meant to be no more exhaustive than the above material on natural hazards. For example, the socio-economic issue of RSL to a small island such as Key West, Florida, which has experienced a well-documented 30 cm rise in sea level in the last 150 years (23 ± 1 cm / century), is quite different to that of an equally sized island in say Micronesia. Key West is a highly developed community with significant financial resources to mitigate RSL, and in fact due to extensive coastal engineering, the area of Key West island per se has *grown* by some 25% in the century and a half since RSL records have been maintained. Most of all developing AOSIS nations would not have the same experience.

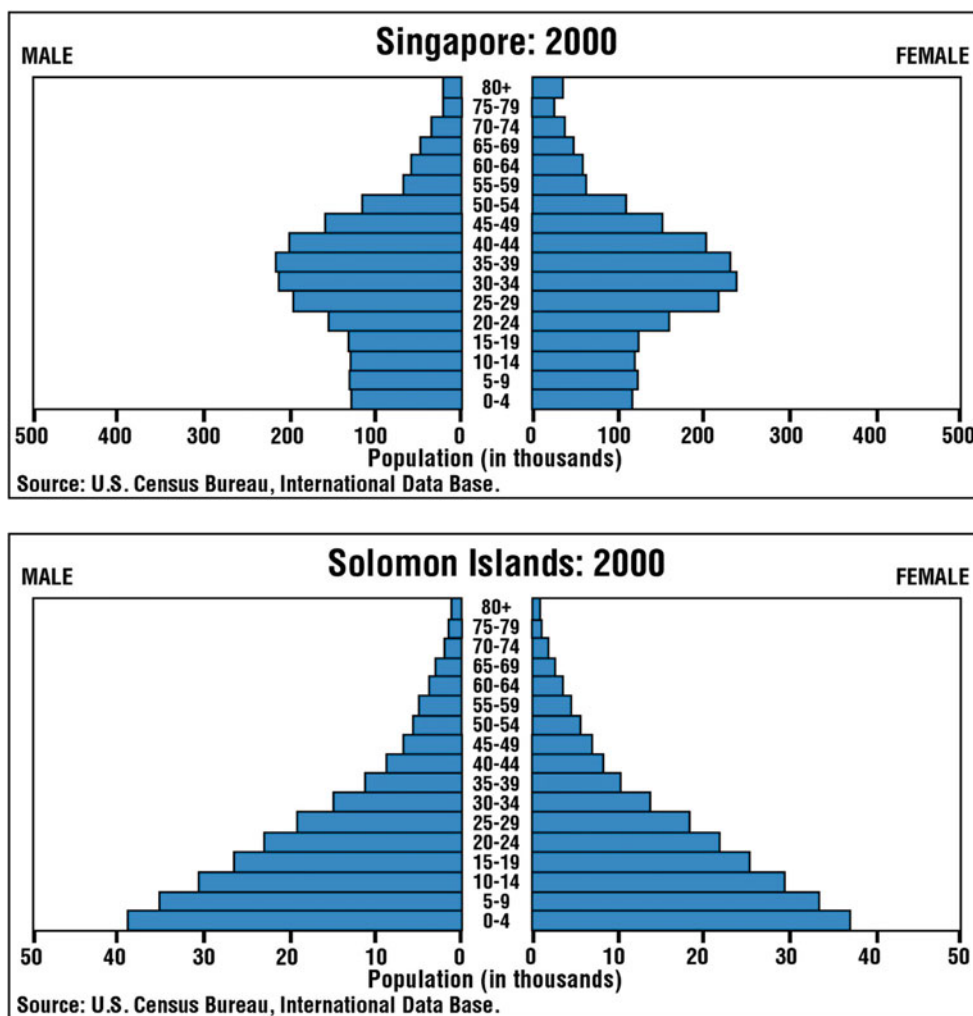
Of the anthropogenic hazards identified in many studies, growth of the human population, deforestation, waste management, and marine exploitation are often mentioned. Again, no single small island will have the identical human-caused “hazards” as neighboring islands, but for many SIDS especially in tropical climes, similar patterns seem to arise. While there is cause for concern regarding global change as an instrument of stress to the small island community, it appears that it is local populations causing local problems that will exacerbate environmental degradation for the foreseeable future.

Demographics of many SIDS are characterized by a population pyramid that is wide at the base (i.e., many younger persons) and narrow at the apex (i.e., few older persons). In comparison, a society characterized by a population pyramid that is essentially cylindrical (i.e., the number of births, deaths, and distribution by age class are quasi-equal) is one of numerical stability (see Fig. 3). Rapidly growing populations require matched growth in services such as education, medicine, food, housing, freshwater, and infrastructure. Small islands are by definition limited in areal extent, and the rapid conversion of agricultural lands to housing, particularly in mountainous islands where cultivatable land is limited to the coastal zone, leads to marginalizing the ability of internal nutritional sustainability.

Deforestation is another consequence of anthropogenic pressure on small islands and other developing areas. Much of the deforestation is past history, such as the conversion of island forestlands to agriculture, as typified by the sugarcane industries of former European colonies in the Intra-Americas Sea. In tropical small islands, it has been the mangroves that have been affected by coastal forestry practices more than any other species. While in many cases mangrove trees have been used for construction, often they are simply removed for ascetic reasons associated with tourism. Not only has this severely impacted the role of these trees as the basis of the tropical marine food web, it has weakened the ability of the coast to resist erosion from tropical storms, tsunamis, and sea level rise.

Waste management in small islands is a critical issue. While in most cases the atmosphere will remove airborne pollutants, the oceanic circulation can cause waste to concentrate on the leeward side. Ship-generated waste particularly that associated with the lucrative cruise ship trade is a common problem in many small islands regardless of their latitude. Governmental agencies such as the International Maritime Organization and the United Nations Environment Programme have invested heavily in creating the facilities to

Small Islands, Fig. 3 Population pyramid from two small island developing states. Upper panel shows a stabilizing population, and the lower panel shows a rapidly growing population (Data from IDP 2000)



manage such wastes, and yet with the growing pressure of other population-related activities on the environment, the coastal zones of small islands are increasingly stressing the marine ecosystem. Waste management of course is of a multiple nature, and while the focus herein is on the coast, the cause is both point source and distributed.

Finally, considering anthropogenic hazards, the actions of human populations toward marine exploitation must be included. In the Caribbean Sea, for example, the actions of one nation upon another lead to political as well as environmental stress. Overfishing by one small island state upon another as a concerted practice has caused entire reef systems to become incapable of sustainable production. Scuba divers have been known to sweep past an unprotected coral reef and essentially remove the entire breeding population of numerous marine species. But the practice of using modern technology to efficiently scour a fishery is not limited to SIDS, as is quite evident in the regions of George's Bank and the Grand Banks. While overfishing has been used to illustrate the issue, souvenir hunting and jewelry harvesting of corals and other

marine gems have equally led to exploitation of the marine environment well beyond a sustainable level.

Toward a Sustainable Environment

Small islands, in many cases, are tuned to the wealth associated with tourism. All the anthropogenic and natural hazards summarized above seem to converge on this central issue. The rugged beauty of isolated small islands, which were created by natural forces that would be considered catastrophic in a contemporary socioeconomic context, is just the scenery that lures visitors. James Michener's *Tales of the South Pacific* excites the imagination of fellow wanderers among the family of man. And it is that very attraction that stresses the ability of developing nations to sustain the economic growth necessary to support burgeoning human populations. Not achieving a stable population will lead to the failure of all services and sources of income. Ecotourism, while perhaps less lucrative than currently practiced tourism, is inherently sustainable and

should be considered in the mix of solutions leading to sustainable economic development.

Two other critical issues seem to have arisen that cause and limit sustainability: sand mining and water availability. Tourism and other growth industries require raw materials for construction and infrastructure development. All too often the most cost effective, at least at first, is inexpensive concrete products based on abundant beach sands. Healthy coral reefs in tropical small islands produce beach sands that balance the natural sediment transport to deep waters as integrated over time. Parallel processes provide materials in other climes for development. Indigenous peoples used these supplies of nature without regard to overexploitation because of low population densities in the coastal communities. This supply-and-demand balance is no longer viable within the context of exponential human population growth. Thus a paradigm shift is necessary for sustainability, not only regarding beach sand mining but also for perhaps the most central human issue: freshwater supply.

Peoples of small islands, whether off the desert coast of Bahrain or the rain forest coast of Singapore, or the frigid Argentine Islands, are focused on the supply of potable water. Sustainability requires that supply exceed demand. No human need except for air, at least on the diurnal timescale, exceeds the requirement for safe and abundant drinking water. Earth does not have a water shortage problem; it has a *cheap* water shortage problem! Small islands are perhaps the quintessential expression of this issue. OTEC (ocean thermal energy conversion) and other ocean and coastal engineering solutions to the general issue of supply and demand of abundant freshwater through technology, have as yet, realized their theoretical potential. While freshwater is the immediate issue, it is energy-inexpensive-energy ratio that is the central issue, and it is not limited to SIDS.

Conclusions

Small islands are an enigma in many respects within the context of coastal science. They are subjected to storms of tropical and extratropical origin that seem to exacerbate the impact by anthropogenic coastal development. Relative sea level change is seen to be extremely site specific, voiding the common perception that all such geographical entities are subject to inundation by external global forces beyond local control. Tsunamis, the devastating seismic sea waves, are recognized as not just a Pacific Ocean hazard, highly populated coastal zones of the Atlantic Ocean being perhaps most vulnerable from a socioeconomic context (Proenza and Maul 2010). France and Portugal have recognized the danger of the next repeat of the 1755 Lisbon earthquake and tsunami by establishing a system to prevent a European catastrophe on the scale of the 1998 Papua New Guinea tsunami.

Population density and pressure upon the coastal marine ecosystem by humankind is clearly a major issue of this quandary. Deforestation, mangrove forests in particular in tropical climates, has repercussions well beyond the erosion issues of sediment systems that provide the nourishment for tourism and other economic stimuli of developing island states. Development brings the inevitable issues associated with waste – what to do with all that stuff? Exploitation of the marine environment extends beyond coral reefs in the tropics, to include fisheries at all latitudes, tourism in pristine locales such as Antarctica, and turtles in the midlatitude hatcheries of the beaches of east central Florida.

Next, there is the quandary of intellectual sustainability. Human population is, and has been for millennia, characterized by migration. The drain of educated young people not returning to their island homes all too often plagues SIDS. How is the international community of scholars going to create the social environment that encourages young professionals to share their knowledge in a developing nation? Educators in developed nations must encourage the return of the best and brightest scholars to homelands desperately in need of their skills, insight, and worldview perspective; and developing states must accept responsibility by creating policies for students sent abroad to want to return.

Finally there is the issue of “Who do you say is my neighbor?” How is the issue of life, liberty, and the pursuit of happiness, however it is expressed in the social context of a particular small island society, executed? Clearly the “business as usual scenario” won’t create a sustainable society. And in that context, small islands are not separable from the larger global continuum of humankind. Certainly AOSIS and SIDS provide a venue for discussing the special problems of societies with radically limited geographic boundaries. Geophysicists in the widest context can isolate the science peculiar to small islands. Solution of their peculiar problems truly requires a paradigm shift from the confrontational litigious nature of geopolitics to the notion of a shared destiny that does not extort funds and resources from one segment of humankind at the expense of another.

Cross-References

- ▶ [Barrier Island Landforms](#)
- ▶ [Changing Sea Levels](#)
- ▶ [Coral Reef Islands](#)
- ▶ [Demography of Coastal Populations](#)
- ▶ [Mangroves, Ecology](#)
- ▶ [Mining of Coastal Materials](#)
- ▶ [Natural Hazards](#)
- ▶ [Sea-Level and Climate Change](#)
- ▶ [Storm Surge](#)
- ▶ [Tsunami Deposits](#)

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South America, Coastal Ecology

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Geomorphologic and Oceanographic Characteristics of South America

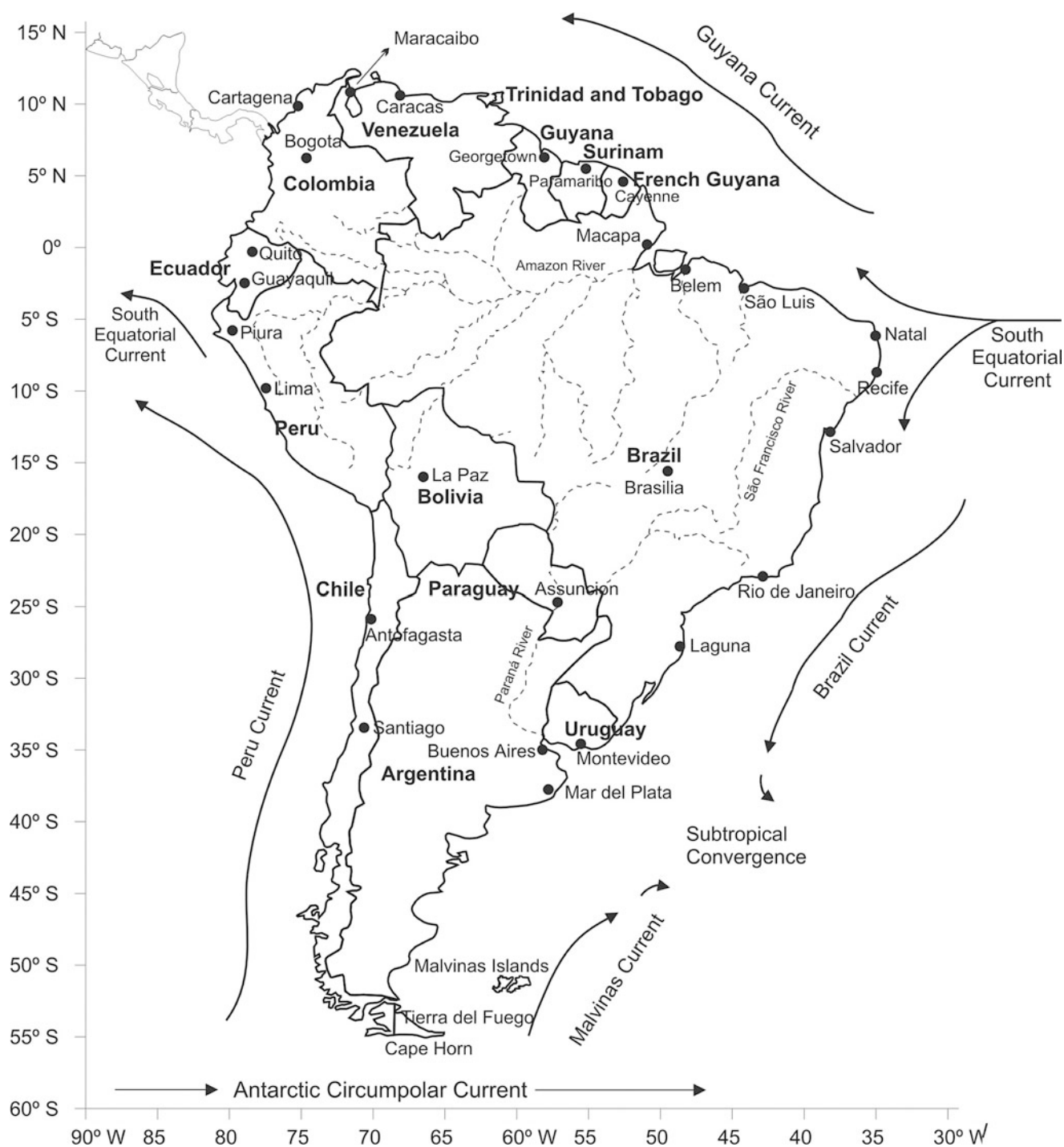
Simone Rabelo da Cunha

South America (Fig. 1) extends from tropical climatic zones (12°30'N) to cold polar zones (about 55°S), encompassing a great diversity of coastal and marine ecosystems. The tectonic history and geological factors, such as the present-day geomorphology and vertical motions of the coastline, influence the coastal and marine ecosystems of South America. Tectonically, South America is divided into two parts, the Andean chains to the west and a vast stable platform to the east, consisting of exposed Precambrian rocks and shallow sedimentary cover rocks (Kellogg and Mohriak 2001). The Pacific Andean coastline is characterized by high relief, a relatively narrow shelf bordering a deep trench, small drainage basins, and rapid vertical motions of the coast. Low relief, broad shelf, and extremely large drainage basins and alluvial fans characterize the Atlantic coastline. Approximately 93% of South America's drainage is to the Caribbean and the Atlantic, away from the Andes, and provides the world's best example of continent-scale drainage control by plate tectonics (Kellogg and Mohriak 2001).

The upper water circulation of Atlantic Ocean as a whole consists in its gross features of two great anticyclonic

circulations or “gyres” separated over part of equatorial zone by the eastward flowing Equatorial Counter Current. In the South Atlantic, the upper-water gyre extends from the surface to a depth of about 200 m near the Equator and to about 800 m at southern limits of the gyre at the Subtropical Convergence. The different portions of this gyre have different water properties. The South Equatorial Current flows west toward the American side of the South Atlantic. Part of the current crosses the equator into the North Atlantic, and is named the Guyana Current. The remainder turns south along the South American continent as the Brazil Current (Fig. 1), which then turns east at about 30°S, at the Subtropical Convergence, and continues across the Atlantic as part of the Antarctic Circumpolar Current (Pickard and Emery 1990). A contribution to the water in the South Atlantic comes from the Malvinas Current (or Falkland Current) flowing north from Drake Passage, up the coast of South America and reaches the Brazil Current at the Subtropical Convergence. The Brazil Current and Guyana Current are warm and saline, with a low concentration of nutrients, having come from the tropic region. The Malvinas Current is cold, with lower salinity and a higher concentration of nutrients, transporting subantarctic waters.

The circulation of the upper water of the Pacific is very similar in its main features to that of the Atlantic. In the South Pacific the superficial currents flow northward, from the Antarctic Circumpolar Current, forming the Peru (or Humboldt) Current (Fig. 1). The Peru Coastal Current is closest to the coast and confined to the uppermost 200 m depth, transporting cold water (14–16°C) in summer (Pickard and Emery 1990). The Peru Oceanic Current (down to 700 m depth) reaches higher velocities than Peru Coastal Current. Between these currents, the weak and irregular southward flow of the Peru (Humboldt) Subsurface Countercurrent is usually sub-superficial but occasionally reaches the surface (Pickard and Emery 1990). Along the Pacific coast of South America, an upwelling area extends from 4°S to 42°S. Peruvian coastal upwelling is peculiar because winds sustain the upwelling process throughout the year. The coastal upwelling system, which hardly comprises 0.02% of the total ocean surface, is of great significance because it determines the enormous productivity of Peruvian and Chilean coastal waters, representing almost 20% of the world's landings of industrial fish (Tarazona and Arntz 2001). Wind intensity and persistence are highest in winter and lowest in summer. Wind stress, and consequently upwelling strength, has intensified over the last decades (Pickard and Emery 1990). The oceanographic conditions along the Pacific coast are influenced by an interannual variability known as El Niño–Southern Oscillation (ENSO), which consists of a cold period (La Niña) and a warm period (El Niño), with anomalous warming of the eastern Pacific during about three months. El Niño occurs every 2–7 years, causing a deepening of the pycnocline and



South America, Coastal Ecology, Fig. 1 South America countries, showing larger rivers and main oceanographic currents

thus rendering the upwelling process inefficient (Tarazona and Arntz 2001).

These complex assortments of geomorphologic and oceanographic characteristics result in very diversified environmental settings, allowing the occurrence of many different coastal ecosystems in South America. The occupation of coastal environments is expanding very fast, and in most cases,

without adequate planning. The rising rates of coastal resources exploitation, the threat of global changes and sea-level rise and the economic losses due to environmental degradation, on the one hand making the coastal management a hard task, and on the other hand, these events evidence the urgency of planning the uses and conservation of these environments.

Sandy Beaches Ecosystems in South America

Omar Defeo and Anita de Alava

Sandy beaches constitute a dynamic interface between the land and the sea that stretch from the base of the dunes to the low tide mark or lower limit of the swash zone. Dunes, beaches, and surf zones are closely linked by the interchange of sand. The morphology of these dynamic and harsh environments is defined by the interaction of waves, tides, sediment characteristics, and topographic features, which combined characterize the morphodynamics of a beach (Short 1996) (see “► Beach Processes” and “► Sandy Coasts”). Sandy beaches dominate ocean and estuarine shores of the South American coast. Pacific sandy beaches display a variety of morphodynamic environments. Venezuela, Colombia, and Ecuador present a narrow discontinuous stretch of pocket sand beaches, alternating with rocky shores and mangrove swamps. In Peru, which constitutes the southern extreme of mangrove swamps in the beginning of the warm-temperate Pacific coast, alternating exposed and sheltered beaches, together with active sand dunes, become prevalent. Chilean beaches are mainly found from 19°S to 42°S. Exposed sandy beaches with different morphodynamics alternate with rocky shores in the north (19–30°S) and intertidal sand flats at the mouth of rivers in south-central regions (38–42°S) (Jaramillo 2001). Atlantic sandy beaches also display wide variations in morphodynamics. Sheltered, pocket beaches are found between Venezuela and northern Brazil, up to Cabo Frio (23°S). Pebble beaches, with a variable amount of rocky fragments, are also found. Southwards, within the warmtemperate biogeographic province, sandy beaches become more exposed and reflective, with coarse sands and steep slopes. A continuous (640 km) exposed, dissipative and microtidal (astronomic tides 0.5 m) sandy beach is found from Rio Grande do Sul (southern Brazil: ca. 29°S) to northeastern Uruguay (Barra del Chuy: 33°45'S), with fine to very fine, well-sorted sands, gentle slopes, well-developed frontal dunes, heavy wave action, a wide surf zone and large barometric tide ranges (Calliari et al. 1996). The physiognomy of this coastline is mainly determined by the prevailing winds and wind-driven surf off the South Atlantic. Southwards, the exposure of beaches is ameliorated by the estuarine effect of the Río de la Plata, between Punta del Este and Montevideo (Uruguay) and also in the northeast of Buenos Aires Province (Argentina). Beyond the estuarine influence of the Río de la Plata River and up to the gulfs of northern Patagonia (around San Matías Gulf/Valdés Península: ca. 41°S), beaches are mainly characterized by fine sands and gentle slopes. These gulfs define the southern limit of the warm-temperate southwestern Atlantic province, which also demarcates a biotic transition in faunal composition. The

cold-temperate coastlines south to 43°S, Patagonia, are mainly sand flats, pebble beaches, and long, high cliffs of Cenozoic marine sediments.

Pacific beaches of northern Chile and Peru present surf zones characterized by upwelling, high detrital inputs from kelp beds and the absence of surf diatoms (Arntz et al. 1987). Contrastingly, exposed dissipative beaches of the Atlantic coast of southern Brazil, Uruguay, and Buenos Aires Province are characterized by non-periodic events of high concentrations (7×10^8 cells L^{-1}) of brown colored patches produced by surf zone diatoms such as *Asterionellopsis glacialis* (Odebrecht et al., 1995). Favorable surf circulation patterns promote greater retention of particulate primary production, thus allowing the beach/surf zone ecosystem of dissipative coastlines to function as semi-closed ecosystems (McLachlan 1980). Long dissipative beaches of moderate to high energy, rip current activity and an associated dune system providing nutrients by groundwater flow, constitute suitable conditions primarily met in southern Brazil and Uruguay (Defeo and Scarabino 1990). Abundant suspension feeders play an important role as efficient mineralizers, providing ammonium sources for surf diatoms. Phytoplankton at exposed reflective and sheltered shores is much less abundant, and food for macrofauna mainly comes from the sea as living and non-living particulate organic matter, while debris and carrion are transported shoreward through the surf zones (McLachlan 1980). Warm-temperate Atlantic beaches are also subject to seasonal blooms of toxic dinoflagellates. The strong dynamics of oceanic waters and the influence of the Río de la Plata also produce several toxic outbreaks (Méndez et al. 1996).

Research on sandy beaches of South America has been mainly focused on macrofauna (community and population levels), particularly on exposed coastlines. Concerning communities, investigations deal with structure, seasonal dynamics and distribution patterns at spatial scales ranging from macro (km) to meso (individual beaches). Crustaceans, mollusks, and polychaetes are the most diverse groups. A review of existing information does not provide support to the well-known worldwide pattern of increasing number of species from temperate to tropical latitudes. The concurrent effects of morphodynamics, variations in tide amplitude and beach length, as well as the presence of localized areas with extremely high productivity due to upwelling events, mask the recognition of latitudinal trends. Indeed, the reverse pattern was found by Jaramillo (2001) in 3,000 km of Chilean coastlines (20–42°S): species richness increases southwards, from 15 (20–23°S) to 28 (40–42°S). Similar trends were reported by Dexter (1992) for 284 beaches around the world, who found an increase in species richness from tropical to cold-temperate beaches. Moreover, the number of species is not directly related to beach length or tidal amplitude: the greatest species richness in the Atlantic coast was recorded in pocket (short) sheltered and very sheltered

beaches with sand sediments mixed with rock fragments (pebbles and cobbles), along São Sebastião Channel, south-eastern Brazil (23°43'S), where 31 species of crustaceans, 35 of mollusks, and 24 of polychaetes were documented for only one beach (Engenho d'Água; e.g., see Nucci et al. 2001). Species richness, abundance, and biomass vary with beach morphodynamics: flat dissipative beaches have more diversity, abundance and biomass than reflective ones (Defeo et al. 1992; Jaramillo 1994). Intertidal suspension feeders are typical inhabitants in these high-energy dissipative beaches. On the contrary, only supralittoral species (insects, ocyropodid crabs, some talitrid amphipods, and cirolanids) remain in truly reflective beaches.

Sandy beaches of South America display across-shore zonation of their macrofauna. In general, the upper shore or supralittoral of tropical and subtropical beaches is characterized by crabs (Ocyropodidae), whereas on temperate beaches talitrid amphipods and cirolanid isopods prevail. The upper littoral or mid-shore fringe is dominated by cirolanid isopods and polychaetes, whereas the lower littoral and shallow sublittoral are mainly occupied by cirolanids, amphipods, bivalves, and hippid crabs of the Genus *Emerita*. Concerning tropical coastlines of Venezuela (ca. 10°N), the supralittoral tends to be dominated by the brachyuran *Ocypode quadrata*, talitrid amphipods and the cirolanid isopod *Excirolana brasiliensis* (also on the middle shore), whereas the mid-littoral is mainly occupied by anomurans of the genus *Emerita* and *Lepidopa*, the mollusks *Donax denticulatus*, *Tivela* sp., *Terebra cinerea*, and *Olivella verreauxi*, and the polychaetes *Glycera* sp. and *Sthenelais limicola*. The shallow sublittoral is dominated by the gastropod *Mazatlanina aciculata*, the echinoid *Mellita quinquesforata* and the starfish *Astropecten marginatus* (e.g., see Penchaszadeh et al. 1983). In Colombia, Pacific beaches with fine sands have higher species richness and abundance than the Atlantic counterparts with coarser grains (from 3°N to 11°N). Both coasts are dominated by the isopods *E. brasiliensis* and *Exosphaeremonia diminutum*, as well as the polychaetes *Scolecopsis agilis* and *Hemipodus armatus* (Dexter 1974). Sandy shores of Peru (about 8–12°S) are dominated by a low number of species with high biomass, notably the suspension feeders *Mesodesma donacium*, *Donax peruvianus* and *Emerita analoga* in the shallow intertidal (Arntz et al. 1987). Other members are (Penchaszadeh 1971): the sand crab *Ocypode gaudichaudii* and the cirolanid *E. brasiliensis* (supralittoral and upper/mid-littoral), the polychaetes *Hemipodus triannulatus*, *Nephtys monilibranchiata*, and *Nephtys multicirrata* (mid-littoral) and the anomurans *Blepharipoda occidentalis* and *Lepidopa chilensis* (lower littoral and shallow sublittoral). In Chile, the most common species are the insect *Phalerisida maculata* (supralittoral) the cirolanid isopod *E. brasiliensis* (supralittoral and upper littoral), the anomuran crab *E. analoga*, the bivalve *M. donacium*, and the polychaete *Nephtys impressa* (swash zone and

shallow sublittoral), all of which occur between 20°S and 42°S (Jaramillo 2001). The supralittoral ocyropodid *O. gaudichaudii* found in northern Chile (20–23°S) is replaced by the talitrid amphipod *Orchestoidea tuberculata* in south-central Chile (32–42°S), as a result of latitudinal gradients in rainfall and sediment temperature.

On Atlantic coastlines, coarse sandy beaches in Surinam (ca. 6°N) show an impoverished fauna characterized by the presence of *O. quadrata* in the supralittoral (Swennen and Duiven 1982). Sheltered sandy beaches of northern Brazil are dominated by crustaceans, mollusks, and polychaetes, commonly found on channels from Bahia to Rio Grande do Norte. The supralittoral is dominated by *O. quadrata* and the peracarids *Orchestia* spp. and *E. brasiliensis*. Lower fringes present a low number of crustaceans and mollusks, depending on tidal exposure and the existence of reefs or freshwater runoff. Southwards (Rio de Janeiro State: ca. 23°S), exposed reflective/intermediate beaches harbor an average of 7 species, notably the talitrid amphipod *Pseudorchestoidea brasiliensis* (supralittoral), the isopod *E. brasiliensis* (supralittoral and upper littoral), and the anomuran *Emerita brasiliensis* (lower littoral and shallow sublittoral), which together represent 95% of the macro-fauna (Veloso and Cardoso 2001). For roughly the same latitude, some 90 species of mollusks, crustaceans, and polychaetes appear in Engenho d'Água beach, along São Sebastião Channel (Brazil: ca. 23°42'S). The high species richness in these sheltered unconsolidated marine beaches, with rocky shores inundated by sand, is associated with higher habitat heterogeneity: rocky fragments form small tide pools and moist and shady microhabitats, which may widen the presence of some intertidal species (Denadai et al. 2001). In one of the few comprehensive studies that included sublittoral fringes, Borzone et al. (1996) showed a marked increase in the number of species from the intertidal to sublittoral of sandy beaches of Parana, Brazil (25°30'S), particularly in reflective beaches.

Sandy beaches between Southern Brazil (29°S) and Argentina down to Golfo Nuevo (ca. 43°S) show a gradient from warm-temperate to cold-temperate faunistic composition, where the Río de la Plata acts as an effective ecological barrier (Escofet et al. 1979). The wide chain of wave-exposed beaches along southern Brazil and Uruguay show maximum species richness (at least 23), density and biomass at the long chain (640 km) of dissipative beaches of fine sands and flat slopes, while the lowest values (5 species) occur in reflective beaches of Uruguay, with coarse sands and steep slopes (Defeo et al. 1992). Crustaceans are the most represented (nine species), followed by mollusks (six), insects (five) and polychaetes (three). The cirolanid isopod *Excirolana armata* dominates in numbers, whereas the suspension feeders *Mesodesma mactroides*, *Donax hanleyanus*, and *E. brasiliensis* dominate in terms of biomass. The highest beach levels are occupied by *O. quadrata* (only juveniles are sporadically

found in Uruguay), the amphipod *P. brasiliensis*, insects and the cirrolanid isopod *E. braziliensis*. The polychaetes *Euzonus furciferus* and *Scolecopsis gaucha* are found at mid-beach levels, together with the isopod *E. armata*. The higher intertidal levels are dominated by the amphipods *Bathyporeia ruffoi*, *Phoxocephalopsis zimmeri*, and *Metarpinia* sp., and the suspension feeders *M. mactroides*, *D. hanleyanus*, and *E. brasiliensis*, whereas the lower ones are occupied by the isopods *Macrochiridotea giambiagiae*, *Macrochiridotea lilianae*, and *Macrochiridotea robusta*, the polychaetes *Hemipodus olivieri* and *Sigalion cirriferum*, and the gastropods *Olivella formicacorsii*, *Buccinanops duartei*, *Olivancillaria vesica auricularia* and *Olivancillaria teaguei*. On Brazilian coastlines, *Callianassa mirim* (Callianassidae) is also found in this fringe (Gianuca 1983).

Species richness and abundance decrease towards Río de la Plata (ca. 35–36°S), as a result of high salinity variability and decreasing exposure. Low abundances of *P. brasiliensis* and *E. armata* co-occur with some brackish water species. In Buenos Aires Province (ca. 37°S), the supralittoral species *O. quadrata* and *E. braziliensis* are not found, and *Orchestia platensis* and *P. brasiliensis* are rarely present. The intertidal is inhabited by the isopod *Cirolana argentina*, the clams *M. mactroides* and *D. hanleyanus*, and gastropods of the genus *Buccinanops* and *Olivancillaria* (Escofet et al. 1979). The supralittoral of beaches of north Patagonian gulfs is inhabited by *O. platensis*; the upper littoral by *C. argentina*, the bivalve *Darina solenoides*, polychaetes *Scoloplos* sp., *Travisia* sp., and *Onuphis dorsalis*, and amphipods Haustoriidae and Phoxocephaliidae. The lower littoral and shallow sublittoral are represented by the bivalves *Tellina petitiata* and *Bushia rushi*, the gastropods *Olivella* sp., *Buccinanops globulosum*, *Olivancillaria arcelsis*, *Olivancillaria uretai* and *Olivancillaria urceus*, the isopod *Serolis gaudichaudii*, several amphipods (*Monoculopsis valentini*, *Stephensia haematopus*, Haustoriidae and Phoxocephaliidae) and polychaetes. *Callianassa* sp. and the polychaete *Arenicola brasiliensis* appear in sheltered beaches (Escofet et al. 1979).

Faunistical zones across the beach also showed important variability, with aperiodic (even daily) and seasonal components (Brazeiro and Defeo 1996). The across-shore position of patches varies according to the different susceptibility of each species to environmental variations. Generally, species richness increases from upper to lower beach levels in dissipative beaches, whereas in very reflective beaches the lower fringes tend to disappear. One or two faunistical belts are found in reflective beaches, and three or four in dissipative ones (Defeo et al. 1992).

Life history traits, dynamics, and structure of sandy beach populations show clear geographical patterns. In one of the few large-scale studies conducted in South America, Defeo and Cardoso (2002) analyzed macroecological issues of the

intertidal mole crab *E. brasiliensis* along some 2,700 km between Urca (Rio de Janeiro, Brazil: 22°57'S) and Arachania (Uruguay: 34°36'S). Most (11 from 14) life history traits of the crustacean *E. brasiliensis* were significantly correlated with mean water temperature of the surf zone, which could be summarized as follows: (1) a shift from continuous to seasonal reproductive and recruitment events from subtropical to temperate beaches; (2) an increase in individual sizes of the smallest ovigerous female, fecundity at size, predominance of females and individual weights from subtropical to temperate beaches; and (3) a decrease in male sizes, growth, and mortality rates towards temperate beaches. The concurrent effect of morphodynamics at a regional scale was also detected and in some cases masked clear latitudinal trends.

Lifespans appear to be also related to temperature, ranging from 1 to 3 years for the small, warm water species to ≥10 years in some cold-temperate clams (e.g., *M. donacium* in Chile), but most species live for 1–4 years and have relatively rapid growth to maturity. Wide seasonal growth oscillations occur in southern populations, with lowest rates during autumn and winter (Gómez and Defeo 1999 and references therein); alternatively, tropical populations show continuous growth throughout the year.

Sandy beach populations are labile to climatic variability, which generates resurgences and mass mortalities. Arntz et al. (1987) showed dramatic fluctuations of the suspension feeders *D. peruvianus*, *M. donacium*, and *E. analoga* in Peru, as a response of the strong (ENSO) event of 1982–83. Following the mass mortalities of the dominant *M. donacium* as a response of a strong increase in sea surface temperatures during ENSO, *D. peruvianus* increased from 5% to 60–100%, and *E. analoga* increased from <1% to 29%. This suggests differential responses to climatic events and also potential interspecific interactions because of competitive release of resources by dominant macrofauna members. Variations in species composition along the Chilean coast have also been related to latitudinal gradients in rainfall and sediment temperature (Jaramillo 2001).

Beach morphodynamics affect biodiversity on sandy coastlines. An increase in species richness, abundance, and biomass from reflective to dissipative beaches has been reported (Defeo et al. 1992). However, a large-scale study conducted in Chile (Jaramillo 2001 and references therein) showed that species richness, abundance, and biomass are higher at intermediate beaches, some of them located near areas of persistent upwelling. Populations that co-occur in contrasting environments are less sensitive to variations in beach morphodynamics, as revealed by comparisons of abundance, reproduction, recruitment, fecundity, growth, mortality, and burrowing time between reflective and dissipative beaches (Gómez and Defeo 1999; Defeo et al. 2001).

Most sandy beach populations have strong and persistent distribution patterns in response to an environment that is

spatially and temporally structured by sharp, small-scale gradients (Defeo and Rueda 2002). Specific habitat preferences across the beach generally determine significant correlations between abundance and mean grain size, beach face slope, sediment moisture and penetrability. Aggregations persist in time, but, in contrast to sessile species, the position of the patches varies according to the different susceptibility of each species to variations in physical (e.g., sediment moisture and temperature), and biological (swimming ability and burying) factors (Giménez and Yannicelli 2000). Unpredictable and strong short-term increase in tide ranges (i.e., up to sand dunes) generated by storm surges, wind-driven surf off the ocean and barometric tides could be a source of mortality. This mortality in many cases is size-dependent, as smaller intertidal organisms are more susceptible to being stranded in the upper littoral. Mass mortalities at dissipative beaches probably occur at a higher frequency than on reflective ones, where wave intrusion is mitigated by the steep slope on the lower shore.

Intra-and interspecific interactions are important in structuring sandy beach populations and communities. Results from long-term and large-scale field experiments, together with laboratory observations and field monitoring, suggest that population fluctuations in dissipative sandy shores are produced by the intertwined forces of environmental, density-dependent, and human-induced factors operating together at different spatial and temporal scales (Defeo 1996a). Concurrent field sampling and laboratory experiments with cirrolanid isopods showed that intra-and interspecific interactions would be of importance in population regulation (Defeo et al. 1997). Local populations tend to present density-dependent growth and mortality. Density-dependent and density-independent forces acting together can jointly explain population fluctuations over time, as shown for the yellow clam *M. mactroides* in Uruguay (Lima et al. 2000) and the guild of suspension feeders *M. donacium*, *D. peruvianus*, and *E. analoga* in Peru. The fact that density-dependent mechanisms are often manifested in dissipative beaches, with highest values of species richness and abundance, suggests that biotic interactions should be of utmost importance in these systems, whereas reflective beach populations should be mainly regulated by individual responses to the environment. Considering that these populations are highly spatially structured, potential compensatory and overcompensatory mechanisms have been shown to occur at small spatial scales in dissipative beaches with highest macrofauna abundance (Defeo 1996a). Other studies, however, suggest that environmental harshness leaves limited scope for competitive interactions, as suggested for Chilean beaches (Jaramillo 2001).

Predation by birds, gastropods, crabs, fishes, and insects usually generates high rates of size-dependent mortality, particularly at dissipative beaches with high macrofauna biomass. Research efforts were mainly focused on food habits

of juvenile fishes in the inner surf zones of dissipative sandy beaches (e.g., Monteiro-Neto and Cunha 1990). Predation could act as a selection pressure determining body size of macrofauna, and thus higher growth rates should be directed to diminish predation risks. An active selection of the site could also be invoked to decrease predation and desiccation risks.

Sandy beach ecosystems receive a variety of increasing anthropogenic impacts such as forestation, exploitation of coastal species, coastal development, pollution (disposal of liquid and solid wastes, freshwater discharges from wide plain basins used for agriculture and cattle rearing, agrochemicals), and unplanned recreational use (Lercari and Defeo 1999).

Sandy beach fisheries in South America rely on species with different life histories. Harvested stocks are the intertidal clams *Mesodesma* and *Donax*, mole crabs *Emerita* and supra-littoral ghost crabs *Ocypode*. The main artisanal/recreational fisheries are based on the extraction *M. mactroides* (handpicking: Brazil, Uruguay and Argentina) and *M. donacium* (handpicking and diving: Peru and Chile). These fisheries have shown to be notoriously difficult to manage, because ocean beaches are open and extended systems readily accessible to commercial and recreational users, but also to unauthorized harvesters. Management measures are difficult to enforce and appear to be beyond the finances of most management agencies. High uncertainty in stock estimates, lack of basic biological knowledge, improvements in fishing power, low operating costs and a risk-prone management attitude, determined a trend towards overexploitation (Castilla and Defeo 2001). Nevertheless, a successful management experiment was documented for *M. mactroides* of Uruguay, which included the experimental manipulation of the system based on a fishery closure for 32 months (Defeo 1996a, 1998). The long-term study (8 years) showed that abundance, age composition, age-specific survival, fertility, and population growth rate of the yellow clam were significantly affected by adult density, recruitment variability, and fishing intensity (Brazeiro and Defeo 1999). Population structure of the sympatric clam *D. hanleyanus* showed inter and intra-annual fluctuations of recruits and adults, with uneven periods of abundance related to fluctuations in the fishing effort targeted on *M. mactroides* (Defeo and de Alava 1995). Operational management tools based on area-specific management plans (legal sizes and catch per fisher and fishing ground) were implemented in Chile through "Management and Exploitation Areas," defined as concessions allocated to fisher communities (Castilla and Defeo 2001).

Disposal of liquid and solid wastes, accidental oil spills and deposition of mine tailings (Castilla 1983) modify community composition, abundance, population structure, fecundity, and zonation of macrofauna in South America. Freshwater discharges from man-made canals also affected

the habitat, community structure and abundance, fecundity and growth rates of *D. hanleyanus* (Defeo and de Alava 1995) and *E. brasiliensis* (Lercari and Defeo 1999). Mechanical disturbance of sands occasioned by recreational users during the summer seems to have little effect on macrofauna (Jaramillo 2001).

Information is comparatively scarce or absent for plankton, meiofauna, vagile megabenthos, and nekton of the surf zone, as well as for birds and sub-terrestrial fauna inhabiting sand dunes. An ecosystem approach is needed for modeling networks of interactions between different components of food webs. Other processes and mechanisms affecting structure and functioning of sandy beach macrofauna (e.g., predation, commensalism, parasitism, and mutualism) are still little documented. There is considerable scope to elucidate the role of competition in structuring sandy beach communities, and its relative contribution according to morphodynamic states, exposure, and tidal regimes.

Biogeographic patterns in life history traits have not been adequately assessed in sandy beach ecology. Further research should be directed to clarify large-scale patterns, the relative contribution of factors influencing beach fauna and to decipher cause–effect relationships. At present, our knowledge of how dissimilar responses of populations could also result from locally adjusted genotypes or a combination of plastic and genetic responses is limited, and should be addressed by genetic studies throughout biogeographic ranges. Little is also known about dispersive abilities of meroplanktonic larval phases, and the mechanisms influencing larval distribution and connectivity between populations are still poorly understood (Defeo 1996b). Research should also focus on planktonic stages and physical-oceanographic information (e.g., nearshore hydrodynamics) to determine the spatial scales at which the population dynamics is to be considered an open process, that is, if it is more related to the arrival rates of larvae than to post-settlement processes.

Long-term studies are scarce at both the community and population levels, and the consequences of natural or human-induced disturbances on the structure and dynamics of macrofauna are poorly known. Recent mass mortalities that occurred on a geographical range have been poorly documented and understood. Different approaches are needed to perform well-designed experimental and field studies directed to critically assess environmental impacts in these fragile ecosystems. This should be complemented by laboratory (microcosm) experiments to understand ecophysiological effects of pollutants and responses to abiotic factors coming from different human sources.

Sandy beach populations may be partially regulated by density-dependent processes of unknown extent. Future work should emphasize scale-dependent experimental manipulations of abundance, both through field and laboratory experiments. The spatial analysis of populations and the

environment should be useful for monitoring changes in abundance, structure, and dynamics of populations.

Accumulation of toxins, such as those associated with algal blooms of toxic algae, can cause mass mortalities of suspension feeders or render them unsafe for human consumption. This limits the potential utilization of many species and creates the need for careful monitoring and management. As human impacts on coastal waters continue, blooms of toxin-producing phytoplankton could affect beach suspension feeders more. This should be a focus for future studies.

Rocky Shore Ecosystems in South America

Rosana Moreira da Rocha

Rocky shores are important ecosystems along the South American coast because of the great diversity of species and economic importance of some of these species, such as oysters, mussels, crabs, and fish. As a transitional ecosystem between terrestrial and marine environments, the rocky shore can be divided into three parts. The first, upper part of the rocky substrate is permanently exposed to the air (the supralittoral zone), the second part is only exposed during low tides (the intertidal or mid-littoral zone), and the third part is always submerged (the sublittoral zone). While the rocky substrate does not permit organisms to burrow, crevices, pits, and the accumulation of boulders create a three-dimensional matrix that provides different microhabitats for many kinds of organisms. Most of the organisms attach to the rock surfaces, thus providing additional micro-habitats on which other organisms may settle. The sessile fauna on the intertidal zone is usually distributed in horizontal belts of dominant species, giving the zone a striped appearance called zonation. In the sublittoral zone, this distribution is much less marked and only some species show a depth zonation (see “► [Rock Coast Processes](#)”).

On the Atlantic side of South America, traveling along the coast from north to south, rocky-shore communities first appear in the Santa Marta area of Colombia (11°13'N 74°14'W–11°20'N 74°05'W). For the most part, the coastal mountains, part of the Sierra Nevada de Santa Marta, plunge abruptly into the sea. However, a low and narrow rock (phyllite or quartz-diorite) platform exists, at least in the bays, which are often split up into small promontories and islands. The spatial distribution of this community was studied by Battaström (1980), who described the zonation as follows: the supralittoral is inhabited only by the gastropods *Littorina ziczac*, *Littorina angustior*, *Nodilittorina tuberculata*, and *Tectarius muricatus*; the upper intertidal zone has a bare appearance but is covered by blue-green algae and represents an impoverished upper part of the lower barnacle–vermetid zone; the lower intertidal zone is a

narrow zone formed by barnacles (*Tetracrita* sp., *Chthamalus angustitergum*, *Megabalanus stultus*) and/or vermetids (*Petalconchus varians*, *Spirogyllus annulatus*); the sublittoral fringe is covered by a mixture of algae (*Ralfsia expansa*, *Sargassum* sp., *Laurencia papillosa*, and *Ectocarpus breviarticulatus*) and invertebrates (barnacles, chitons, limpets, boring organisms, and a multitude of crabs and snails living inside the algae); the sublittoral zone is characterized by the presence of encrusting coralline algae overgrown by a mat of macroalgae (most important are the rhodophyceans *Amphiroa* sp., *Jania* sp., *Hypnea* spp. *Ceramium nitens*, *Centroceras* sp., and *Laurencia* spp.) and many sessile and semi-sessile invertebrates (sponges, horny and scleractinian corals, the zoanthids *Palythoa* sp. and *Zoanthus* sp., sea anemones, scattered barnacles and chitons, the gastropods *Fissurella angusta* and *Acmaea* spp., the bivalves *Isognomon radiatus* and *Isognomon bicolor* and the ascidians *Styela canopus*, *Herdmania momus*, and *Pyura vittata*).

In Venezuela, rocky shores are present in the regions of Trinidad and the Paria Peninsula (11°N), but they are apparently poorly studied and no literature was found describing this area. Rocky shores are absent from the delta of the Orinoco River to the northern coast of Brazil, where the littoral is covered by a mangrove formation.

A few rocky shore ecosystems exist along the northeastern Brazilian coast, but hard substrates are commonly available as sedimentary rocky fringes along the shallow coast. Algae, coral communities, and other sessile invertebrates such as sponges, tunicates, bryozoans, and cnidarians usually inhabit these fringes. One of the few granitic rocky formations already studied is an intertidal and shallow sublittoral platform of boulders and gravel at Ponta Cabo Branco, Paraíba (7°S), which joins the fringing sedimentary reefs along the coast. These rocks form a unique ecosystem in the area composed of a rich sessile community of sponges (*Chondrilla nucula*, *Haliclona* sp., *Tedania ignis*, *Halichondria* sp.), tunicates (*Didemnum duplicatum*, *Didemnum psammotodes*, *Polysyncrator amethysteum*, *Eudistoma* spp.), cnidarians, macroalgae (*Ulva* sp., *Gelidium* sp.), oysters (*Crassostrea rhizophorae*) and associated fauna. In the sublittoral, corals (*Siderastrea stellata*, *Mussismilia hartii*, *Mussismilia hispida*, *Montastrea cavernosa*, *Agaricia agaricites*, and *Porites astreoides*) and the zoanthids *Palythoa* sp. and *Zoanthus* sp. are common invertebrates.

From central Brazil (Espírito Santo, 20°S) to Laguna in the south (Santa Catarina, 28°S), rocky shore environments are formed by granitic or basaltic rock, resulting from the erosion of the border of the Serra do Mar mountain chain, which lies parallel to the coastline. It is not a continuous ecosystem, but forms more or less extended outcroppings between sandy beaches. This is the principal rocky shore ecosystem along the coast of Brazil and also the most known. The first published descriptions of the community appeared between

late 1940s and early 1950s (Oliveira 1947, 1950; Joly 1951, 1957) and since then, information on species distributions and communities has been found through many species surveys and ecological studies.

The following synthesis provides an overall picture of the community, and is based on work by Joly (1951, 1957), Oliveira (1947, 1950), Nonato and Pérès (1961), and Oliveira-Filho and Paula (1983). The supralittoral mostly comprises bare space used by the periwinkle gastropods *L. ziczac* and *Nodilittorina lineolata*, which are the most common and characteristic organisms at the lower part of this zone. Isopod crustaceans of the genus *Lygia* are also very common at the supralittoral.

The upper intertidal contains a dense belt of the barnacle *Chthamalus bisinuatus*, while *Tetracrita* and *Megabalanus* are found lower in the intertidal zone, but are not dominant space occupiers. Below the *Chthamalus* belt, *Brachidontes solisianus* is the dominant in terms of space used. However, in sites exposed to waves, *Perna perna* mussels can form dense beds in the mid and lower intertidal. More recently, *I. bicolor* has invaded the southeastern and southern coasts of Brazil and is replacing *B. solisianus* in some areas. The mid-intertidal zone is also colonized by many algae, such as *R. expansa*, *E. breviarticulatus*, *Centroceras clavulatum*, *Jania adhaerens*, *Acantophora spicifera*, and *Ulva* spp. On wave-exposed sites one can add *Porphyra* spp. (in winter), *Chaetomorpha antennina*, and many species of fleshy macroalgae, and coralline algae both articulated and encrusting. On the low intertidal and sublittoral the oyster *C. rhizophorae* can be abundant and common algal species are *L. papillosa*, *A. spicifera*, *Jania capillacea*, *Amphiroa fragillissima*, *Hypnea cervicornis*, *Rhodomenia pseudopalmata*, and *Corallina officinalis*. In many sites along the coasts of São Paulo, Paraná and Santa Catarina, the sabelariid polychaete *Phragmatopoma caudata* forms extended sandy reefs along the low intertidal zone. In the latter two states, the colonial ascidian *Eudistoma carolinense* forms a narrow belt below *Phragmatopoma* and comprises an important intertidal microhabitat for more than 117 species (Moreno and Rocha 2001). Among the vagile invertebrates characteristic of the intertidal zone are some herbivorous mollusks, such as the limpets *Collisella subrugosa* and *Fissurella clenchi*, besides the periwinkles, which can migrate down from the supralittoral zone; the predator whelks *Stramonita*, *Pisania*, and *Leucozonia*; and the crabs *Pachygrapsus gracilis* and *Pachygrapsus transversus*.

A dense *Sargassum* spp. bed usually marks the upper sublittoral fringe and is probably the most abundant macroalgae in both tropical and subtropical sublittoral zones. It forms dense beds usually covering a thin layer of encrusting coralline algae, which are dominant space occupiers in many sublittoral sites, especially at places where grazing pressure is high. The most important herbivores at the sublittoral are the

urchins *Arbacia lixula*, *Echinometra lucunter*, *Lytechinus variegatus*, and *Paracentrotus gaimardi*; chitons; gastropod mollusks of the genera *Aplysia*, *Astraea*, and *Tegula*; and fish, such as damselfish (*Stegastes* sp.), surgeons (Acanthuridae), and parrotfish (Scaridae). Sessile invertebrates common to the sublittoral zone are cnidarians of the genera *Palythoa* and *Zoanthus*, which form large encrusting colonies, and a variety of small arborescent hydrozoan colonies. Sponges are also very common and diverse – more than 120 species are known within the merely 25 km long São Sebastião Channel on the north coast of the State of São Paulo (Hajdu et al. 1996). The most conspicuous bryozoan is the encrusting *Schizoporella*, but many small arborescent colonies are frequent, especially from the genus *Bugula*. Common ascidians are the encrusting colonial didemnids and the solitary pyurids and styelids (Rodrigues et al. 1998).

The southern coast of South American, between Rio Grande do Sul (Brazil) and Rio de la Plata estuary is formed basically by a long sandy beach of 750 km in length. Very little rocky substrate is available along this coast that could represent an important geographical barrier for sessile organisms. Close to the borderline between Santa Catarina and Rio Grande do Sul (29°S), there is a small rocky outcropping formed by very tall volcanic rocky structures, called Torres (“towers” in English) because of the height of the rocks. Artificial rocky substrates are also present, such as the rocky jetties at the entrance of Lagoa do Patos, which also supports a diverse encrusting community.

Very few experimental studies have been undertaken to understand the dynamics of both intertidal and sublittoral rocky shore communities along the Brazilian coast. Caging experiments showed that littorinid grazing activity controls the abundance of microalgal populations in the supralittoral zone (Apolinário et al. 1999). Experimental analysis of succession on cleared substrates in the intertidal zone revealed that *C. bisinuatus* recruits more on the *Brachidontes* zone and prefers granitic to basaltic substrates (Tanaka and Duarte 1998). On the belt formed by *Sargassum cymosum* var. *nanum* in the low intertidal zone, succession is maintained in its earlier stages due to desiccation stress causing widespread algal mortality during the early summer (Paula and Eston 1989). Experimentally cleared areas of the sublittoral zone revealed that *Sargassum stenophyllum* was both the competitive dominant with slow growth and an opportunist colonist (Eston and Bussab 1990).

South to the Rio de la Plata River estuary in Argentina, there are loess platforms separated from a coastal cliff of 7–8 m by sand strips. The upper intertidal is inhabited by crusts of blue-green algae, while the mytilid bivalve *Bachidontes rodriguezi* is the dominant space occupier in the mid-intertidal zone with densities up to 33,000 in. m⁻² and an associated community of around 40 species. The cause of this dominance is not only competitive abilities but also the

absence of important predators at the intertidal zone (Lopez Gappa et al. 1990). The barnacles *Balanus amphritite* and *Balanus glandula* are present in areas with fewer *Balanus rodriguezi*. The first species appears in Quenquén Harbor (38°34'S; 58°42'W) and the second has become abundant in Mar del Plata Harbor (38°0'S; 57°32'W) in recent years. There are no periwinkles in the Mar del Plata region, where they are replaced by the gastropod *Siphonaria lessoni*, which is abundant in both the intertidal and supralittoral zones.

Community distribution in the southern region of Argentina, the Chubut Province, is as follows: the supralittoral covered by various blue-green algae with the chlorophyte *Enteromorpha intestinalis* and the mollusks *S. lessoni* and *Brachidontes purpuratus* in tide pools; in the upper intertidal there is a striking absence of barnacles and the most important inhabitant is the gastropod *S. lessoni*; in the mid-intertidal there are three well-marked belts: *B. purpuratus*, the coralline alga *Corallinetum officinalis* and the phanerogams *Spartina montevidensis* and *Salicornietum ambigua*; the low intertidal is occupied by a mytilid belt (*Mytilus edulis platensis*, *Aulacomya ater*, and *B. purpuratus*); the sublittoral does not include the usual phaeophytes but instead has a wide belt of the chlorophytes *Codium fragile* and *Codium vermilara* establishing its upper limit (Olivier et al. 1966).

Chile has an extended coast, about 2,600 miles long, which can be topographically divided into two very different regions: south of Chiloé (41°29's), the coast is discontinuous, with mountains along the shore rising up to 3,000 m, comprising an eroded tectonic pattern of glaciated and non-glaciated fjords. North of Chiloé the coastline is very regular and fully exposed to the prevailing winds and waves, but there are geological differences: from Chiloé up to Navidad (33°57'S) the coastal range is made up of metamorphic shale of low elevation, from Navidad to Antofagasta (23°38'60S) it is mainly granitic rock, and north of Antofagasta it consists of volcanic rocks with sedimentary intrusions (Stephenson and Stephenson 1972; Santelices 1991).

Along the southern coast of Chile, south to Chiloé Island, the bad weather conditions and the limited access result in a lack of knowledge about the intertidal and sublittoral communities in these wave-exposed open coasts. Nevertheless, community descriptions are available from sheltered islands in the southernmost tip of South America. For instance, in the Beagle Channel the supralittoral zone contains several bands of lichens and the upper intertidal is covered by a mixture of algae (*Bostrychia mixta*, *Hildenbrandia lecanellieri*, *Pilayella littoralis*, *Adenocystis utricularis*, *Enteromorpha* spp., *Porphyra* spp., *Spongomorpha* spp.). Next, the mid-intertidal has mussels and barnacles together with high densities of the gastropods *Acmaea*, *Collisella*, *Nacella*, and *Siphonaria*, while the low intertidal is covered by pink encrusting coralline algae and high densities of *Nacella magellanica* and *Nacella mytilina*, with the brown alga

Lessonia vadosa marking the lowest limit of the intertidal zone. The sublittoral communities of these habitats consist of belts of *Macrocystis pyrifera* in sheltered and semi-sheltered sites (Santelices 1991).

In central Chile the supralittoral inhabitants are barnacles (*Jehlius cirratus*) in reduced densities, aggregations of *Littorina araucana* and *Littorina peruviana*, the algae *Porphyra columbina* and dark-red crusts of *H. lecanellieri*. The upper intertidal is covered by pure or mixed stands of chthamaloid barnacles (*Chthamalus scabrosus*, *J. cirratus*), while the mid-intertidal usually contains a belt of the mussel *Perumytilus purpuratus* mixed with the algae *Centroceras clavulatum*, *Enteromorpha compressa*, *Iridaea laminarioides*, *Ulva rigida*, and *Polysiphonia* spp. (Santelices 1991).

Towards the northern coast of Chile, this same pattern of species distribution is observed for wave-exposed rocky habitats, but there is a northward reduction in the number of belt-forming algae species. *Macrocystis pyrifera* disappears north of Concepción (36°49'60S) where the dominant algae are *Lessonia trabeculata* in the sublittoral and *Lessonia nigrescens* along the sublittoral-intertidal fringe (Santelices 1991). Rocky sheltered communities have not been well studied and the only description available is from Guiler (1959) in which the intertidal biota in Antofagasta Bay was studied. There the rocks are shale that forms a wide erosion platform covered by an association of *Pyura praeputialis* and *Corallina chilensis* immediately below the barnacle belt. The lower limit of the *Pyura* belt might be controlled by predation by the starfish *Heliaster helianthus* and *Stichaster striatus*. At the lower intertidal *Ulva* replaces *Corallina*; the large barnacle *Austramegabalanus psittacus* occurs in the lowest part of the *Pyura* belt and both organisms are covered by *Ectocarpus confervoides* and *Halopteris hordacea*. *Pyura chilensis* is widespread along the coast of Chile and do not form belts as *P. praeputialis* in Antofagasta region (Santelices 1991).

Much experimental work has been done to examine species distributions in both the intertidal and sublittoral zones of the Chilean coast. *P. praeputialis*, for instance, maintains its dense intertidal beds by intraspecific self-facilitating mechanisms that enhance recruitment to the border of previously settled individuals (Alvarado et al. 2001). The intertidal rocky community in Mehuin (39°25'60S), in the southern coast, is regulated by herbivory and competition; there are apparently no intertidal carnivores capable of controlling herbivore densities and high herbivore densities can destroy the red alga *Iridaea boryana* cover; depending on the season of the disturbance, the community may be dominated by the alga or by barnacles plus crustose algae (Jara and Moreno 1984). In the sublittoral zone, experimental kelp canopy removal revealed that *M. pyrifera* control the species composition of understory algal community (Santelices and Ojeda 1984), and, in central

Chile, the removal of *L. nigrescens* results in a community of calcareous crustose alga *Mesophyllum* sp. in the presence of herbivory and large patches of *Gelidium chilense* in the absence of herbivory (Ojeda and Santelices 1984). In northern Chile, the abundance of herbivores, algae morphology, plant density, water movement, and the egg case of elasmobranchs which tie plant stipes together are the most important ecological factors for the persistence and stability of *L. trabeculata* beds (Vasquez 1992).

A high degree of eco-geographic isolation seems to be a general characteristic of the intertidal and shallow sublittoral rocky communities of Chile. The result is a high degree of endemism and several of the endemic species occupy unique ecological niches with unknown parallels in comparable habitats elsewhere (Santelices 1991). The causes of this isolation, and factors limiting the distribution of these species and communities are unknown and offer an important area of future research.

Oceanographic conditions and upwelling are strong influences on the coastal communities of Peru. This upwelling area associated with highly productive waters stretches from 4°S southward to 40°S, in central Chile, but the most intensive effects are seen in the Peruvian coast, especially during the winter (Tarazona and Arntz 2001). Because of the upwelling process, intertidal and shallow sublittoral zones of northern Chile and Peru have high species diversity. In Peru, the forest-forming kelps (*Lessonia* spp. and *M. pyrifera*), the mussel *Argopecten purpuratus*, the gastropod *Thais chocolata*, the crabs *Cancer setosus*, *Cancer porteri*, and *Platyxanthus orbigny*, and the sea urchin *Loxechinus albus* all occur in great numbers with a large biomass. Kelp forests form a sublittoral belt about 15 m wide and harbor numerous associated species. Mussels (*Perumytilus purpuratus*, *Semimytilus algosus*) also form beds in the intertidal zone, with more than 70 associated species. The structure of these communities tends to be controlled by grazers and predators, while the population dynamics of the algae *Macrocystis* is influenced by the upwelling and El Niño events (Tarazona and Arntz 2001). As an example of this influence, the intense El Niño event of 1982–83 caused mass mortality of key species like the mussel *Semimytilus* and brown algae (*Macrocystis* and *Lessonia*) along the Peruvian rocky shores (Tarazona et al. 1988).

In Colombia, the tectonic processes along the Pacific coast have given origin to abundant steep cliffs (>45° slope) and rocky shores with more gently sloping platforms, composed of boulders, pebbles, and gravel from cliff erosion. Along the northern coast and in the interior of Buenaventura Bay (3°52'60N), species have the following distributional patterns: desiccation-tolerant blue-green and green algae, a lichen species, periwinkles (*Austrolittorina aspera*, *Littoraria zebra*), the crab *Grapsus grapsus*, and the isopod *Lygia baudiana* inhabit the supralittoral.

About 20 species occupy the upper intertidal zone, among them the periwinkle *L. zebra*, the crab *Pachygrapsus transversus* on less wave-exposed cliffs, while barnacles (*Tetraclita*, *Chthamalus*), limpets (Fissurellidae, Acmaeidae, Siphonariidae), crabs (*G. grapsus*) and some green algae occur on exposed cliffs. The mid-intertidal zone is dominated by bivalves (*Brachidontes* sp., Isognomidae, and oysters) and the associated fauna is comprised by crabs of the families Xanthidae (*Eriphia squamata*) and Grapsidae (*P. transversus*), and the red coralline alga *Lithothamnion*. In the lower intertidal zone it is possible to find barnacles, anemones, sponges, the gastropods *Acanthina brevidentata* and *Thais kiosquiformis*, and some crabs. At this level an increasing number of boring organisms (bivalves of the families Pholadidae, Petricolidae, and Mytilidae and the ghost shrimp *Upogebia tenuipollex* contribute to the bioerosion of the rocky cliffs. Erosion rates can be as much as $300^3 \text{ m}^{-2} \text{ month}^{-1}$ for igneous rocky cliffs and $450 \text{ cm}^3 \text{ m}^{-2} \text{ month}^{-1}$ for sedimentary rocky cliffs (Cantera and Blanco 2001).

The most important stresses on rocky shores are the trampling effects of tourists and fisherman, oil and sewage pollution, and selective collecting for food and obtaining mussel seeds for cultures. All these processes have been studied in different rocky shore ecosystems and showed marked influences in the intertidal community structure. Although many rocky shore ecosystems are located in protected areas inside ecological reserves, usually the goal of the reserve is to protect either the terrestrial or the underwater ecosystem while neglecting the intertidal zones. One of the few well-studied reserves is the Estación Costera de Investigaciones Marinas (ECIM) in Las Cruces ($33^{\circ}30'S$), Chile, with a 5 ha human-exclusion exposed shore established in 1982. The intertidal community changes and food-web cascading effects, which occurred inside the reserve, showed that humans act as an efficient and selective keystone predator (Castilla 1999).

The rocky shore communities throughout the South America coasts are reasonably well described. Yet, almost no information exists on the Patagonian coast of Argentina and the Ecuadorian coast. On the other hand, mechanisms of community structure are best studied in Chile and, very recently, Brazil. Apparently, the conservation of rocky shores has not been of primary importance for most countries environmental conservation policies. The few well-monitored known marine reserves are located in Chile.

Coral Reefs Ecosystems in South America

Beatrice Padovani Ferreira and Mauro Maida

Coral reefs are highly diverse ecosystems that have been compared with the tropical rain forest. They are formed by

coelenterates that secrete a calcium skeleton as well as by other calcium secreting organisms such as mollusks, coralline algae, and sponges (see “► Coral Reefs”). The structure formed by these organisms offer shelter and support for an incredible diversity of life, making coral reefs crucial to the culture and livelihood of millions of people in tropical coastal environments. Corals have on their living tissue symbiotic algae called zooxantellae that contribute with oxygen and organic compounds and receive protection from the coral colony. The phenomena known as bleaching is caused when the zooxantellae leaves the coral, usually due to stress related causes. In the last decades, mass bleaching events that have caused the death of many coral reefs have been reported from several parts of the world. These events, related to climate change, have added to other anthropogenic stresses, which are causing an alarming rate of degradation of coral reefs. The concerns with the health and conservation of coral reefs have led government and organizations to establish several programs for monitoring and protecting these ecosystems. The Global Coral Reef Monitoring Network (GCRMN) was established in 1995 with the objective of encouraging and coordinating monitoring of coral reefs at the government, community, and research levels around the world. Brazil, Colombia, and Venezuela are part of the Node of GCRMN for southern tropical America, together with Panama and Costa Rica (Wilkinson 2000).

Coral reef formations in tropical South America are more developed in the Atlantic coast, as several important cold upwelling areas inhibit reef development in the Pacific coast. Coral reef formations on the Pacific side are found in a few reef patches on the coast of Colombia and on offshore islands. The Atlantic coast of South America is under strong continental influences, which introduces large amounts of sediment and inhibits the development of extensive coral reef formations in some areas, especially around large rivers such as the Amazon, Orinoco, and Magdalena. Coral reefs occur along the Atlantic coast of Colombia, Venezuela, and Brazil.

Colombia is the only South American country with both Pacific and Atlantic coasts and coral reefs. There are about $2,700 \text{ km}^2$ of coral reefs in the Caribbean waters, of which 75% are located in oceanic reef complexes. Along the mainland coast, there are fringing reefs on rocky shores such as the Santa Marta and Urabá areas (Garzon-Ferreira et al. 2000).

In Venezuela, reefs occur along three Caribbean areas out of total 2,875 km of Caribbean and Atlantic coastline. Along the continental coast of Venezuela, the more developed coral formations occur in the Morrocony National Park and adjacent reefs, where more than 30 coral species can be found. The best reef formations are found 100 km offshore at Los Roques Archipelago, with 57 coral species and reefs growing up to 50 m depth (Garzon-Ferreira et al. 2000).

Brazil is located in the central-oriental portion of South America with approximately 7,408 km of coastline running from 4°25'N to 33°45'S. Coral reefs are sparsely distributed along almost 3,000 km of coast (Maida and Ferreira 1997), and their distribution and location is still poorly known (Castro and Pires 2001). Laborel (1969) provided the most thorough qualitative description of Brazilian reefs (see also Maida and Ferreira 1997 and Castro and Pires 2001 for reviews). The Brazilian coast can be divided into three biogeographical realms. Intertropical (northern coast): comprises the northern coast from the French Guyana border to Cabo de São Roque. Tropical: the largest portion of the Brazilian coast, from Cabo de São Roque to Cabo Frio, Rio de Janeiro State. Subtropical: from Cabo Frio to the border with Uruguay. Reef formations, including some that are not true coral reefs, are present mostly along the tropical Brazilian coast (northeastern coast), although some coral growth also occurs at the northern region and in the southeastern coast up to São Paulo State. Coral diversity is low, with only 18 hard coral species, but 10 of these are endemic to Brazil. Of those, three species have an even more restricted distribution, only occurring on the reefs of Bahia State. Brazilian reefs present only one species of reef-dwelling soft coral, *Neopongodes atlantica*, which is an endemic form (Maida and Ferreira 1997).

Reef formations such as those that are typical of the northeastern Brazilian coast are rare elsewhere, not displaying the distinctive zones generally observed in reefs around the world (Leão et al. 1988). One of the main characteristics of the Brazilian reefs, are the constructions made by calcareous algae, from the group of the Melobesiae, and vermetid gastropods of the genus *Petalonchus* and *Dendropoma*. These formations can be found on crystalline and eruptive rock, but are especially common on the seaward side of sandstone banks and coral reefs. They grow in the upper part of the reef front forming structures that are similar to the algal ridges of the Indo-Pacific reefs (Laborel 1969). Endemic species like *Favia leptophylla*, *Mussismilia braziliensis*, *Mussismilia hartii*, and *Mussismilia hispida*, that among the principal reef builders, are archaic forms, the remnants of a tertiary fauna that was preserved in a refugium provided by the seamounds of Abrolhos bank during the last glaciation (Leão et al. 1988).

On the northern coast the main geographic feature is the immense Amazonian estuary with a width of more than 350 km. The water of the Amazonian estuary, loaded with vast amounts of sediments, is transported by littoral currents to the north up to Guyana, forming an extensive barrier for coral reef development. According to Laborel (1969), the region has no reefs, only scattered coral growth, but further down the coast, about 80 km off São Luís, the capital of Maranhão State, lays a large coral bank called Parcel de Manuel Luis, whose existence has been known to navigators

since the 17th century, mostly because of the danger it represented to navigation.

Reef formations are present along the northeastern coast from Cabo do São Roque (Rio Grande do Norte state) to the south of Bahia State. In Cabo de São Roque there is a group of oval shaped reefs located a few miles from the coast. These reefs are simple structures, usually formed by numerous pinnacles in a shallow sandy base. On the reef flat only two species of coral occur, *Siderastrea stellata* and *Favia gravida*. Calcareous algae Melobesiae and the vermetid gastropods form an algal ridge. The seaward crest is dominated by *Millepora alcicornis*, followed by *M. hartii* on the slope and *Montrastea cavernosa* at greater depths. According to Laborel (1969) the main reef builder in this region is *S. stellata*.

In this region, from Natal to São Francisco River, the principal characteristics are the coastal sandstone banks and superficial coral reefs, disposed in various lines running parallel to the coastline along more than 600 km. The sandstone banks and superficial coral reefs form lines that are not continuous, and in the places where these formations are interrupted, the coast takes the form of small bays, normally with mangrove swamps whenever creeks and rivers are present (Maida and Ferreira 1997). The sandstone banks are structures that can reach up to 10 km in length and 20–60 m wide. They can appear directly adjacent to the beach, or as submerged formations on the high tides (Dominguez et al. 1990). In some areas up to three lines of reefs can be seen at one coastline, forming an effective protection to the shore. The length of each reef varies from 1 to 4 km, for the reefs that are exposed at low tide, and up to 10 km for the submerged reefs, such as the reefs located off Itamaracá Island, north of Recife (Dominguez et al. 1990). The depth of surrounding waters is seldom greater than 10 m. In this area, the region of more extensive coral development is located between Recife and Maceió Cities. The Tamandare reef complex is the most studied area due to the presence of the Oceanographic Institute in Recife, established in 1958 by the Federal University of Pernambuco, and due to the work of Laborel (1969), who presented a detailed description of the area (Maida and Ferreira 1997). Coral reefs in the area present a distinctive feature, given by their growth as isolated columns of 5–6 m high and expanded laterally on the top. Where the growths of these reef columns are dense, the reefs coalesce at their tops creating large structures with open spaces below the surface, forming a system of interconnected caves (Maida and Ferreira 1997). The coral fauna of the reefs in this region is richer than up north. From the 18 species of stony coral described for the Brazilian coast, 9 species were described for this coast. The main reef builders in this region are the species *M. hartii* and *M. cavernosa*. The fish fauna is similar to the Caribbean fauna, but less diverse, with basically the same families but less species.

Further south, reefs disappear in the area around the mouth of the São Francisco River. The São Francisco River, with an average run-off of $3,300 \text{ m}^3 \text{ s}^{-1}$, discharges large quantities of fine material and exerts considerable variation in salinity in the seawater for several kilometers off and along the coast, forming a large barrier for coral reef development over 100 km wide.

Coral formations are observed again south of the river mouth, along the coast of Bahia State. This coast has the higher diversity of reef formations, presenting superficial reefs; fringing reefs; large isolated mushroom shaped pinnacles, called “chapeirões,” and large platform bank reefs (Leão et al. 1988; Dominguez et al. 1990). *M. braziliensis*, an endemic species to the Bahia coast, is the most important coral species, and the main reef builder. In the coast of Bahia State the continental shelf widens up, reaching over 200 km in the southern part of the coast. In this region, “Abrolhos,” the largest and more diverse reef complex on the Brazilian coast, is located. Coral reef formations in the Abrolhos area are spread over an area of $6,000 \text{ km}^2$, up to 15 km long and 5 km wide. The whole region has been a National Marine Park since 1983, being the first Marine Park established in Brazil. A special permit is required from IBAMA (Brazilian Environmental Institute) and the Brazilian Navy to land on these islands. In Abrolhos there is observed the highest diversity of corals in Brazil. All hermatypic scleractinian corals and hydrocorals found in the Brazilian Coast are present in Abrolhos, from which seven species, including the principal constructors, are endemic forms. Two of the Brazilian species of scleractinian corals, *M. braziliensis* and *Favia leptophylla*, and the hydrocoral *Millepora nitida* only occur in the coast of Bahia.

South of Abrolhos, variations in water temperature are greater throughout the year, and there is a greater vertical temperature gradient. Reefs gradually disappear on the northern coast of Espírito Santo State in Rio de Janeiro State, few scattered points of coral growth are found in Cabo Frio, Angra dos Reis, and Ilha Grande, between the barriers represented by the cold upwelling waters around Cabo Frio, and the mouths of the rivers São Mateus, Mucuri, and Doce. The southern limit of coral growth is São Paulo State.

As in most regions of the world, anthropogenic impacts have been the main cause of degradation on coral reefs in South America. The human related activities that affect reefs in the region are the same that threaten most coral reefs around the world, such as land use practices that increase sedimentation, industrial, domestic and agricultural pollution, mining, over-exploitation of reef resources, and uncontrolled tourism. Coral reefs in the region support important biodiversity reservoirs and an expanding tourism industry (Garzon-Ferreira et al. 2000). According to the report of the Status of Coral Reefs of the World (Wilkinson 2000), bleaching events

appear to have increased in frequency, but decreased in severity, throughout the 1990s, due to the global warming phenomena.

Mangrove Ecosystems in South America

Simone Rabelo da Cunha, Mônica Maria Pereira and Tognella-De-Rosa

Mangroves are transitional ecosystems between land and sea. Mangrove plants are adapted to salinity variation, water-logged and hypoxic or reduced mud sediments (see *Mangroves*). Mangrove forests show best development in tropical protected shores, where low wave energy and abundant freshwater supply allow deposition and accumulation of fine organic mud and salinity range between 5 and 35 (Lugo and Snedaker 1974). The extension of forests inland depends on tidal range and topography, and large forest belts can extend several kilometers landward from the sea. Under these optimal environmental conditions and humid areas, mangroves can attain their maximum growth and productivity. Examples of these are reported for Ecuador and Colombia, where red mangrove (*Rhizophora*) trees reach 40–50 m in height and more than 1.0 m in diameter (Lacerda et al. 1993), and for coasts of Surinam, French Guyana, and northern Brazil, where black mangrove (*Avicennia*) trees reach 30–45 m in height and 0.7–1.0 m in diameter (Lacerda and Schaeffer-Novelli 1992). The productivity of these systems, in terms of litterfall, varies from 3 to 10 metric ton $\text{ha}^{-1} \text{ year}^{-1}$ (Lacerda et al. 1993; Twilley et al. 1997). Seaweeds can play an important role to mangrove productivity, representing 10–45% of total primary production of mangroves, not only in South America (Peña 1998; Cunha 2001), but also elsewhere (Steinke and Naidoo 1990; Rodriguez and Stoner 1990). Microalgae are certainly important, though not quantified. The high primary production, together with habitat heterogeneity, can sustain a large and diversified fauna, including transient and resident animals. Many of these are economically important, especially for artisanal users of mangrove. Around the world, many people are supported by mangroves' goods and services, as economic or food source (Hamilton and Snedaker 1984; Saenger et al. 1983).

South America mangroves, as much as New World mangroves, have a reduced number of tree species, contrary to Southeast Asia, which has nearly one hundred taxa of true mangrove trees (Tomlinson 1986). There are only four genera: *Rhizophora* (Rhizophoraceae), *Avicennia* (Avicenniaceae), *Laguncularia*, and *Conocarpus* (Combretaceae). The species *Rhizophora mangle*, *Avicennia schaueriana*, *Avicennia germinans*, *Laguncularia racemosa*, and *Conocarpus erecta* have wide distribution. The species *Rhizophora harrisonii* and

Rhizophora racemosa are more restricted to northern places of South America (Cintron and Schaeffer-Novelli 1992). *Avicennia* is the most tolerant genus to environmental stress, sometimes dominating highly saline substrates or low-temperature areas, being able to attaining greater structural development in low salinity, disturbance-free environments (Cintron and Schaeffer-Novelli 1992). Many plant species occur associated with mangrove forests, and the diversity can vary due to climatic conditions and proximity of other ecosystems, like rain forests. Despite this, some species frequently appear to be associated with mangroves around the world, including Atlantic and Pacific coasts of South America. The Malvaceae *Hibiscus tiliaceus* and the fern *Acrostichum aureum* are the most widespread (Cintron and Schaeffer-Novelli 1983). Macroalgae occur colonizing roots and trunks of mangrove trees, rocks, stones, or shell fragments into the mangrove. Red algae of the genera *Bostrychia*, *Caloglossa*, *Catenella*, and *Polysiphonia* and the green algae of the genera *Cladophoropsis*, *Rhizoclonium*, and *Bloodleopsis*, dominate the macroalgal communities. However, other genera can also occur, especially in northern areas, where water transparency and salinity are high (Cordeiro-Marino et al. Cordeiro-Marino et al. 1992).

South American mangroves occur on protected shores of all of the maritime countries except Chile, Argentina, and Uruguay. Along the Atlantic coast mangroves form a nearly continuous belt from northern countries to Laguna, Santa Catarina State, in southern Brazil (28°30'S), and the latitudinal limits are determined by the frequency, duration, and intensity of cold winter temperatures, rainfall and/or frost (Lacerda and Schaeffer-Novelli 1992). Along the Pacific coast, the southern mangrove limit is the Tumbes River estuary, in Piura, northern Peru (5°32'S). The restricted distribution at the Pacific coast occurs due to climatic constraints generated by oceanographic conditions along the Peruvian and Chilean coasts, where the Andes Cordilleras and the tropical air currents diminishes rainfall, and the upwelling of cold waters of Peru (Humboldt) Current suppresses convective activity. These result in extremely arid climates, high soil salinity and almost absent freshwater input, limiting mangrove occurrence (Lacerda and Schaeffer-Novelli 1992). The western limit of South American mangroves is the Galapagos Islands, off the Ecuadorian coast (Latitude 0° and longitude 91°00'W) (Lacerda et al. 1993), and the eastern limit is Fernando de Noronha Islands, off the Brazilian coast (State of Pernambuco, 3°34'S; 32°24'W) (Herz 1991). At the north of the South Atlantic coast, in Colombia, Venezuela, Guyana, Surinam, and French Guyana, mangrove occurrence is closely associated with protected areas and large rivers.

Peru has two mangrove forests. The larger one, Piura River has 58.5 km². The other one is located at Tumbes River estuary (5°32'S), and has only 3 km², and is the southern

distribution limit of mangroves along Pacific coast. These forests are under a subtropical climate, with annual rainfall ranging from 66 to 300 mm, and temperature ranging from 18°C to 32°C. Peruvian mangroves are under high natural environmental pressure due to low rainfall that results in high salinity (Echeverria 1993). Occasionally the "El Nino" phenomenon, which occurs during summer, causes rapid increases in rainfall. During these events, intense rains induce large geomorphological changes in the coast, which strongly affect the mangrove forests. Human pressure, mainly deforestation for shrimp culture, also occurs, although mangroves are legally protected (Echeverria 1993).

Mangrove forests in Ecuador occupy 161,770 ha, and more than 70% of these are located in the Gulf of Guayaquil (3°S, 80°W), in Guayas Province. This is the largest estuarine ecosystem on the Pacific coast of South America (Twilley et al. 2001), and its fauna is described in South American mudflats. The dominant species is *Rhizophora harrisonii* followed by *R. mangle*, *A. germinans*, *L. racemosa*, and *Conocarpus erectus* (Twilley et al. 2001). The structure of mangrove forest in this river-dominated estuary indicates optimum growth conditions with tree heights from 25 to 40 m, although some forests have trees up to 7 m (Twilley et al. 1997), probably due to differences in soil phosphorus concentration and other variations on edaphic conditions. Rainfall is seasonal, with more than 95% of the precipitation occurring from December to May, causing seasonal river discharge ranging from 200 m³ s⁻¹ during dry season to 1,400 m³ s⁻¹ during wet season. Seasonality does not occur to soil salinity, neither to total litterfall of the forests, but occurs to the leaf component of litter fall (Twilley et al. 1997). Mangrove forests with structural differences (mainly tree height) show different rates of litterfall, ranging from 6.5 to 10.6 ton ha⁻¹ year⁻¹ (Twilley et al. 1997). Owing to the construction of shrimp ponds and urban expansion along the shore of Estero Salado, mangrove areas of the Gulf of Guayaquil and the Guayas River estuary decreased from 159,247 ha in 1969 to 122,566 ha in 1995. Furthermore, defoliation of *Rhizophora*, *Avicennia*, and *Laguncularia* by larvae of *Oiketicus kirbyi* has contributed to loss of mangroves, though impacts vary interannually (Twilley et al. 2001).

Colombia has 358,000 ha of mangrove, 90% of these along the Pacific coast and 10% along the Caribbean coast. The most important estuaries are Buenaventura bay (3°54'N; 77°W), in the Pacific coast of Colombia, and the delta of the Magdalena River (Colombia's largest river), that includes the large lagoon complex of Ciénaga Grande de Santa Marta (11°N; 75°45'W), in the central Caribbean coast of Colombia. Climatic differences between the Pacific and Atlantic coasts of Colombia, due to the effects of the Inter-Tropical Convergence Zone, have resulted in Pacific humid tropical forests and Caribbean dry tropical forests. These differences strongly

influence mangrove forests structure and productivity, and Pacific mangroves show higher complexity and biomass, whereas Caribbean mangroves are smaller, with reduced crown and smaller biomass (Alvarez-Leon 1993). Climate and fauna for these areas are described in South American mudflats. Mangroves are the most conspicuous vegetation of Ciénaga Grande de Santa Marta (central Caribbean coast), comprising 50,000 ha. The dominant species is the black mangrove *A. germinans*, followed by the red mangrove *R. mangle*, the white mangrove *L. racemosa*, and buttonwood mangrove *C. erecta*. The lagoon is dotted with small islets of *R. mangle*, and on the alluvial plain a 700 m wide belt of *Avicennia* and *Laguncularia* fringes the shores of the lagoon (Polania et al. 2001). The average height of well-developed stands is 15 m, but 25 m tall trees and also dwarf stands are common (Alvarez-Leon 1993). The maximum diameter at breast height (dbh) is 30 cm in *R. mangle*, 40 cm in *A. germinans* and 13 cm in *L. racemosa*. About eight species of algae are associated with mangrove prop roots, but they have not been quantified yet.

Mangrove habitats in Buenaventura Bay (Pacific coast) are comprised of *R. mangle*, *R. racemosa*, *A. germinans*, *L. racemosa*, *C. erecta*, and the associates *Pellicera rhizophorae* and *Mora oleifera*, these two occupying consolidated substrate (Cantera and Blanco 2001). The mangrove habitats can be classified into three different physiographic types, according to Cintron and Schaeffer-Novelli (1983). Riverine mangroves are well developed (<35 m height, <20 cm dbh) along tidal creeks and estuarine zones of the bay. Bar mangroves are intermediate (<15 m height, <6 cm dbh), growing behind sand ridges near the mouth of the bay. Fringe mangroves growing in platforms resulting from bioerosion of sedimentary cliffs usually develop slowly, or are dwarf, but exceptionally they can exhibit development similar to riverine mangroves (6–9 m height, 4–11 cm dbh) (Cantera and Blanco 2001). Several red and green algae (*Bostrychia*, *Catenella*, *Caloglossa*, *Bloodleopsis*, *Cladophora*, etc), as well as benthic diatoms and blue-green algae grow on prop roots, pneumatophores or form beds (Peña 1998). The litter production ranges from 9.6 ton ha⁻¹ year⁻¹ in bar mangroves to 11.4 ton ha⁻¹ year⁻¹ in riverine mangroves, but is much lower in disturbed riverine mangroves, with 7.5 ton ha⁻¹ year⁻¹ (Cantera and Blanco 2001), and macroalgae contribute about 26% to total annual mangrove production (Peña 1998).

Mangrove forests of Venezuela cover 250,000 ha, and major mangrove areas are the Orinoco River delta, the estuaries of the rivers San Juan, Limon, and San Carlos, and the lagoons la Restinga, Tacarigua, Cocinetas, and Sinamaica, and the Gulf of Cuare-Morrocoy Bay. Except for Orinoco River mangrove forests, which are typically fluvial and can attain over 40 m in tree height, all other mangrove forests of Venezuela are located in arid and semiarid regions (Conde 2001). Annual rainfall in the Orinoco River watershed can be

lower than 1,000 mm in northern lowlands and as high as 8,000 mm in southern high relief areas. In the delta, rainfall is markedly seasonal with two dry and two wet periods. The dominant vegetation in the delta is mangrove, which comprises 73% of Venezuela's mangrove forests. The dominant species are *R. mangle*, *A. germinans*, and *L. racemosa*, though *R. harrisonii* and *R. racemosa* have also been reported. Mangrove forests tend to form a 100 m wide belt along the margins of channels, with some individuals growing up to 20 m height and dbh about 25 cm. Mangroves are replaced landward by tall stands of herbaceous *Montrichardia*, followed by halophobic species (Conde 2001).

The most part of mangrove forests in Trinidad and Tobago occur in Trinidad, which has about 7,000 ha of well-developed forests, used as timber resource and for tourism. These forests are important places for coastal birds. *R. mangle* is the most abundant tree species, and *R. harrisonii*, *R. racemosa* are apparently restricted to Trinidad. *A. germinans*, *A. schaueriana*, and *C. erectus* are also common (Bacon 1993).

Brazil has 7,400 km of coastline and mangroves occur in a patchy fashion on 92% of this entire coastline. They reach from Oiapoque, Amapá (4°30'N) to Laguna, Santa Catarina (28°30'S) (Schaeffer-Novelli et al. 1990), and mangrove area estimates range from 2,500,000 ha (Saenger et al. 1983) to a more realistic 1,376,255 ha (Kjerfve and Lacerda 1993). Brazilian tidal ranges decrease southward, from strongly macrotidal, with ranges greater than 4 m (8 m in some places) in the north, mesotidal (2 m) in most parts of the coast, and microtidal (0.2 m) in the south (Schaeffer-Novelli 1993). Because of the large latitudinal gradient, mangrove forests structure are much more variable in Brazil than in other South America countries. Brazilian mangrove species are *R. harrisonii*, *R. racemosa*, *R. mangle*, *A. germinans*, *A. schaueriana*, *L. racemosa*, and *C. erectus*. *Hibiscus* and *Acrostichum* are very common plants associated with mangrove forests. The smooth grass *Spartina alterniflora* is usual along the entire coast, fringing mangrove and zones of accretion (Schaeffer-Novelli 1993).

At the northern coast (4°30'N to 3°S) the climate is wet, with rainfall ranging from 2,000 to 3,250 mm year⁻¹. Mangroves develop very well north and south of the Amazon delta, reaching 20 m in height. Black mangrove *Avicennia* dominates mangrove forests, and *Rhizophora* stands are more frequent on estuaries under a more direct marine influence. *Laguncularia* is also common, especially in low salinity or backing *Rhizophora* fringes. In the Amazon delta region (1°40'N to 0°36'S) mangroves development and coverage are poor because of the overwhelming influence of the freshwater Amazon discharge, and are mixed with freshwater swamps (Schaeffer-Novelli et al. 1990). Eighty-five percent of Brazilian mangroves occur along the 1,800 km length of this part of the coast, extending more than 40 km inland following the course of estuaries and rivers in the states of

Pará and Maranhão. The Maranhão State has the most extensive (500,000 ha) and structurally complex mangrove forests (Kjerfve and Lacerda 1993). Mangrove trees may reach 45 m height, some with dbh exceeding 0.8 m and above ground forest biomass about 280 ton ha⁻¹. The extension and complexity of these mangrove systems reflect hydrological and topographical characteristics of the coast.

The northeastern coast (3–13°S) has a dry climate, with a long and pronounced dry season. Annual rainfall (1,100–1,500 mm year⁻¹) is usually lower than potential evapotranspiration, and seasonal droughts and hypersalinity are usual. This coast is exposed to high-energy waves, characterized by sandy beaches and dunes, with reefs offshore. Mangroves develop poorly due to lack of freshwater runoff and prolonged droughts (Schaeffer-Novelli et al. 1990), and have been restricted to protected areas in association with estuaries and coastal lagoons.

At eastern and southeastern coasts (13–23°S) a mountain chain (Serra do Mar) approaches the coast, restricting the width of the coastal plain. Shallow coastal lagoons are common behind narrow sandy spits. Rainfall (1,200 mm year⁻¹) is similar or higher than evapotranspiration, and there is no marked dry season. Where the Serra do Mar is very close to the coast, higher rainfall may occur. In areas protected from high energy, mangrove forests are large and well developed (Schaeffer-Novelli et al. 1990), and trees can reach 6–15 m in height, mean dbh from 0.08 to 0.12 m and above ground forest biomass about 65 ton ha⁻¹ (Kjerfve and Lacerda 1993).

In the southern distribution of mangroves (24–28°30'S), rainfall largely varies in small spatial scale, ranging from 1,100 to 2,000 mm year⁻¹, being usually higher than evapotranspiration. Despite the water surplus, mangroves are not as well developed as in northern places, especially due to lower temperatures. *Avicennia* trees are taller than *Laguncularia* or *Rhizophora* in this region, reaching 10–15 m in height (Schaeffer-Novelli et al. 1990). The mean of mangrove forests height vary from 2 to 7 m, with dbh from 4 to 12 cm and above ground forest biomass from 9 to 100 ton ha⁻¹, depending on position of the forest and flooding frequency (Tognella-De-Rosa 2000; Cunha 2001). South from 28°30'S cold winter temperatures inhibit mangrove occurrence, and salt marsh vegetation, as *Spartina*, *Scirpus*, *Juncus*, and other herbaceous plants colonize protected areas and tidal flats.

In all places in Brazil where detailed studies were made (as in Rio de Janeiro, São Paulo, Paraná, and Santa Catarina), there were not always clear zonation patterns for species. However, strong gradients for forests structure (mainly height and dbh) always occur from fringe to inner mangrove forests, reflecting the environmental gradients of flooding frequencies and certainly the geochemistry of the substrate (Schaeffer-Novelli et al. 1990). Latitudinal and local gradients of mangrove forests structure strongly influence the patterns of forests productivity. Well-developed forests usually present

higher biomass accumulation and higher litter fall than forests with shorter trees. In local scale, however, there is not always correlation between forest structure and litter fall, and it seems to be more related to differences in allocation patterns of production, and accumulation of biomass, which can vary due to edaphic differences (Cunha 2001 and references therein). Productive processes are generally related to temperature, rainfall, and nutrients availability, but seasonal patterns of litter fall largely vary, and are not always directly related to productivity of the trees. Litter fall measurements are available mainly for southeastern and southern mangrove forests and vary from 3 to 10 ton ha⁻¹ year⁻¹ (Kjerfve and Lacerda 1993 and references therein). Seaweed production was quantified only in Santa Catarina and varies from 0.3 to 1.8 ton ha⁻¹ year⁻¹, where litter fall varies from 2.3 to 3.8 ton ha⁻¹ year⁻¹ (Cunha 2001). In northern places, where temperature and water transparency are higher, seaweeds production certainly presents an important contribution to mangrove systems. Trophic interactions in mangrove systems are explained on South American mudflats.

Mangroves play an important role in South American economy, providing many goods and services for the human population (Tognella-De-Rosa 2000). These include: coast-line protection and stabilization, nursery for many economically important shellfish and crabs, and source of timber, firewood, charcoal, chemicals, medicine, and water-ways for transport. In the north of South America, a great part of the shrimp fisheries is based on species, which depend on mangroves for completing their development. Timber, firewood, and charcoal seem to be the major uses of mangroves in South America, and mangrove bark is still a source of tannin in most countries (Kjerfve and Lacerda 1993; Tognella-De-Rosa 2000).

Despite their ecological and economical importance, South American mangroves are threatened by diverse natural and anthropogenic disturbances. Dredging and filling of channels, industrial and urban pollution have also resulted in large losses of mangrove areas. Low sheltered embayments often containing extensive mangroves are used for establishment of large industrial complexes, resorts, and extensive mariculture projects (Kjerfve and Lacerda 1993; Schaeffer-Novelli 1993).

Deforestation in mangrove areas has increased in the last years for all countries, mostly for conversion into salt ponds and mariculture, mainly for shrimp ponds. Land reclamation for building condominiums and marinas and illegal occupation are responsible for the most part of the deforestation, mainly in Brazil (Kjerfve and Lacerda 1993). These contribute to the degradation of mangrove areas through physical fragmentation of landscape and the impoverishment or irreversible loss of genetic resources. Despite this alarming situation, mangrove forests are protected in almost all South American countries, and there is an increase in the

preoccupation of public opinion in conservation matters. Simultaneously there is an increase in the quantity and quality of research in all South American countries in the recent years, especially for sustainable use and conservation of coastal ecosystems, including mangroves. Earlier studies had focused mainly on structural aspects of mangrove forests, but in the last 10 years many research groups around South America have been studying productive and litter dynamics, herbivory and competition, population and community dynamics, valuation of mangroves (economic ecology) biogeochemical and hydrological processes, and restoration and recreation of mangroves. Most of these works are just beginning, but the results already found have helped to plan the use and conservation of these amazing ecosystems.

Mudflat Habitats in South America

Carlos Emilio Bemvenuti

When shores become more protected from wave action, they become finer grained and accumulate more organic matter, thus, they become muddier. Mud particles accumulate where currents are low and most of the tidal flats of the world are associated with estuaries and similar embayments. Despite mudflats being characteristic of sheltered habitats, they are far from being static entities, since they represent areas of changing balance between erosion and deposition (Little 2000). The coastal slope and the tidal range determine the areal extension of the tidal flats (see “► [Muddy Coasts](#)”). In regularly flooded coasts, the tidal cycles determine the frequency and length of low tide exposure. In irregularly flooded areas, this exposition reflects the environmental unpredictability, mainly, due to the wind, rainfall, and evaporation effects. Waves, tides, and currents transport and sort the sediment particles, and also, determine their distribution, stability, and composition. Together with local climate and geomorphology, these factors constitute the environmental matrix of tidal flats, affecting the composition, abundance, and the distribution patterns of the organisms (Reise 1985).

The intertidal flat is a rigorous environment for plants and animals, as they are intermittently exposed to the heat of the sun and to the drying action of air and wind. The variety and numbers of organisms gradually increases toward the lowest intertidal zone, which is exposed only during the lowest tides. Here, you will find not only typical intertidal species but also some essentially subtidal ones, able to survive out of water for short periods of time (Little 2000). For mangrove flora, see “► [Mangroves, Ecology](#),” in this entry.

The Gulf of Guayaquil (3°S, 80°W) of the coastal province of Guayas, Ecuador, is the largest (12,000 km²) estuarine ecosystem on the Pacific coast of South America. Surface water temperatures vary between 21.5°C and 25°C during

the dry season and increase by 3°C during the wet season. Salinity decreases from 34 in outer regions to 30 (20 during the wet season) inside the gulf. Tides are 1.8 m near the upper boundary of the Gulf and increase to 3–5 m in the Guayas River estuary near the city of Guayaquil (Twilley et al. 2001). The oysters *Crassostrea columbiensis* and *Crassostrea iridiscens*, the mangrove crab *Ucides cordatus*, the pelecypods *Mytella guayanensis*, *Mytella strigata* live in the intertidal of mangrove habitats, which are dominated by *R. harrisonii*. Mangrove fall rates vary from 6.47 to 10.64 ton ha⁻¹ year⁻¹. A model of leaf litter dynamics suggests that geophysical energy (tides, river, discharge) controls the fate of mangrove leaf litter, though highest litter turnover rates are associated with the activity of the mangrove crab *Ucides occidentalis* (Twilley et al. 1997). High tide predators like the blue crab *Callinectes sapidus* and the shrimps *Penaeus stylirostris*, *Penaeus vanamei*, that occur near mangrove areas (Twilley et al. 2001), develop a severe predation on infauna in mudflats.

The Buenaventura Bay at the central Pacific coast of Colombia, due to the effects of the Inter-Tropical Convergence Zone and intense precipitation, is one of the most humid places in the world, with a mean annual air temperature of 25.9°C, and 228–298 days of precipitation per year. At high slack water, salinity ranges from 18 to 27 at the mouth of Buenaventura Bay to 4.8 at the Dagua River in the inner bay. Complex estuarine conditions and diverse habitats (sandy beaches, rocky shores, mangrove swamps, and mudflats) that maintain rich biological communities, characterize Buenaventura Bay as a diverse and productive ecosystem (Cantera and Blanco 2001). Extensive mudflats occur around creeks in the inner bay with a rich macrofauna (157 spp.), dominated by deposit-feeders, most of which occupy the aerobic upper subsurface layers of sediments, while some bivalves and polychaetes burrow into deeper layers. The high temperatures and desiccation determine a poor diversity on upper littoral, with few dominants that remain buried most of time. Species richness (32) increases in the mid-intertidal and the lower intertidal zone (up to 67 species) displays high diversity and evenness and is inhabited by gastropods (*Natica*, *Nassarius*, *Anachis*, *Cerithium*), bivalves (*Tagelus*, *Anadara*, *Chione*) polychaetes (Amphinomidae, Capitellidae, Glyceridae, Nereidae), crabs (*Panoeus*, *Callinectes*), and gobiid fishes (Cantera and Blanco 2001).

The lagoon complex of Ciénaga Grande de Santa Marta, located on the central Caribbean coast of Colombia, is part of the eastern delta of the Magdalena River (Colombia's largest river). The delta and the lagoon complex (1,321 km²) comprise the Ciénaga Grande (450 km²), the Ciénaga de Pajarales (120 km²), several smaller lagoons, creeks, and channels (150 km²), and mangrove swamps. The lagoon complex can be considered a euhaline–mixohaline system, with mean annual temperature of about 30°C. Temporal and

spatial salinity gradients are common, resulting from variable runoff, seawater intrusion, rainfall, and evaporation. During the dry season (249 mm rainfall), salinity varies between 30 and 40. During wet season periods (1,268 mm), salinity varies between 15 and 20 (short wet season) and close to zero on the long wet season. The Ciénaga Grande complex is a high productive system, where sequential pulses of phytoplankton ($990 \text{ g cm}^{-2} \text{ year}^{-1}$) and seasonal export of mangrove detritus (mangrove litter of $15.7 \text{ tons ha year}^{-1}$), contributes to the carbon budget and sustains high secondary production in the lagoon (Polania et al. 2001). Among the invertebrate fauna, mollusks are represented by approximately 98 species (66 genera and 48 families), of which 61 are of marine origin. Six species occur in mangrove sediments, *Melampus coffeus* being the most abundant among the three gastropods. The fiddler crabs *Uca rapax* and *Uca vocator*, the most abundant crabs in areas of mangrove sediments, can construct their burrows on tidal flat area and ingest detritus and macroinvertebrates there. The crabs *Eurypanopeus dissimilis*, *Pachygrapsus gracilis*, and *Petrolisthes armatus*, species of amphipods and the polychaete *Nereis virens* are abundant on muddy areas (Polania et al. 2001). This polychaete is recognized as an important infaunal predator, which can control the non-predatory infauna densities inside mudflat sediments (Reise 1985). Fiddler crabs (*Uca* spp.), land crabs (*Cardissoma guanhumi*, *Ucides cordatus*), and grapsid crabs (mainly the genus *Sesarma*) are often fairly general feeders, depending mainly on scavenging the deposits and a certain amount of predation. The family grapsidae shows a distinct propensity to herbivore, that may include the scraping of epiphytic algae from the surface of mangrove roots, trunks and branches, and also eating the leaves and the reproductive products of mangrove trees. Others important predators in the Ciénaga Grande are the swimming crabs *Callinectes* spp., which invade the intertidal mangroves or mudflats from the subtidal areas, during flooding periods.

Another Caribbean habitat is the Maracaibo system, which acts as an assemblage of interactive brackish water bodies ($>220 \text{ km}^{-2}$), comprised of the Gulf of Venezuela, Tablazo Bay, and the Maracaibo Strait, which connect Lake Maracaibo in the interior of the basin to the Caribbean Sea. Water temperature follows a seasonal pattern, with a minimum mean temperature of 29°C at 1-m depth in February and a maximum of 32.5°C in September. The resuspension of bottom sediments is the principal source of nutrients in the gulf Tablazo Bay, and the strait. Elevated phosphorus levels in the lake are largely due to terrestrial runoff and the nitrogen is a limiting factor under most conditions in the lake (Rodríguez 2001). Dense mangroves (*R. mangle*) occur on muddy intertidal, and in subtidal waters the widgeon grass *Ruppia maritima* dominates. In muddy intertidal bottoms there are the pulmonate gastropod *Melampus coffeus* and high densities of fiddler

crabs (*Uca cumulanta* with $330 \text{ burrows m}^{-2}$ and *U. rapax* with $113 \text{ burrows m}^{-2}$) and the swamp ghost crab *U. cordatus* ($10 \text{ burrows m}^{-2}$). The ubiquitous clam *Polymesoda solida* and the blue crab *C. sapidus* are abundant components in the submersed meadows. The sublittoral community of the Maracaibo system is not well known, however, the fact that 165 species of mollusks have been recorded suggests a diverse subtidal fauna (Rodríguez 2001). Dominant components are the mytilid pelecypod *Mytella maracaiboensis* (up to $158 \text{ g dry weight}^{-2}$), the tubiculous amphipod *Corophium rioplatense*, and polychaetes, like *Heteromastus filiformis* (980 ind. m^{-2}), *Sigambra* sp. ($1,380 \text{ ind. m}^{-2}$), *Streblospio* sp. ($4,000 \text{ ind. m}^{-2}$), *Capitella capitata* (160 ind. m^{-2}), and *Nereis succinea* (360 ind. m^{-2}). The composition and distribution of species clearly follows a salinity gradient (Rodríguez 2001).

In northeastern Brazil, the Itamaracá estuary (824 km^2) shows characteristics of a tropical hot and humid ecosystem. Both salinity (27) and water temperature (26.8°C) are lower in the rainy season (February to August) than the salinity (34.1) and temperature (30.1°C) in dry season (September to January). Mangrove forests (28 km^2), dominated by *R. mangle* and *L. racemosa*, occupy the lowlands along the Santa Cruz Channel and the lower part of tributaries. On mudflat habitats the detritus is the main diet item of macroconsumers, such as fiddler crabs *Uca* spp., the shrimps *Penaeus schmitti* and *Penaeus subtilis*, and the mollusks *Neritina virginea*, *Heleobia australis*, and *Tagellus plebeius* (Medeiros et al. 2001).

In the southern part of the coastal plain of São Paulo State, Brazil (25°S , 48°W), the lagoon region and estuary of Cananéia extends for approximately 110 km. Tropical air masses prevail from the end of spring (September) and the end of summer (February) and cold fronts occur in April and May. The salinity is highly variable (0–34), and the input of freshwater and the introduction of sediments into the system, during the last 150 years, have modified physiographic and hydrologic characteristics, which influence biological structure and ecological functions of the lagoon region (Tundisi and Matsumura-Tundisi 2001).

Among the invertebrate fauna (73 spp.), polychaete, crustacea, and mollusca dominate in number of species, while deposit feeding polychaetes characteristic of mudflat areas like *Loandalia americana*, *Laonice japonica*, *Clymene* sp., and *Clymenella* sp., are the most common species. The seasonal changes in macrobenthic abundance and diversity are attributed to changes in salinity, redox potential, sediment granulometry, and organic matter concentration (Tundisi and Matsumura-Tundisi 2001).

Paranaguá Bay (612 km^2) a subtropical estuarine system on the coast of Paraná State in southeastern Brazil ($25^\circ30'\text{S}$, $48^\circ25'\text{W}$), is comprised of two main water bodies, the Paranaguá and Antonina bays (260 km^2) and the Laranjeiras and Pinheiros Bays (200 km^2). Mean salinity and water

temperature in summer and in winter are 12–29 °C and 23–30 °C and 20–34 °C and 18–25 °C, respectively (Lana et al. 2001). Mangroves colonize most intertidal areas around the bay (*R. mangle*, *A. schaueriana*, *L. racemosa*, *C. erectus*). *S. alterniflora* marshes colonize tidal flats or creeks as mono-specific, discontinuous narrow (up to 50 m wide) belts in front of the mangroves (Lana et al. 1991).

While salinity and environmental energy gradients appear to control large-scale distribution patterns in the bay (Lana 1986), plant architecture and food availability seem to be the main source of small-scale macrofaunal variability (Lana et al. 1991). The opportunistic gastropod *Heleobia australis* is dominant on the mudflats in the inner part of Paranaguá Bay. Two assemblages are characteristics of the low-energy areas in the central part of the bay: (1) one dominated by the polychaete *Clymenella brasiliensis* and the gastropod *Turbonilla* sp., in low-energy environments; (2) the other, dominated by the polychaetes *Owenia fusiforme* and *Magelona* spp., in moderate energy environments. Another sector of the bay, with silty-clay sediments on the lower intertidal flats, showed low diversity macrobenthic invertebrates, probably due to drastic fluctuations of salinity in 4–5 h periods. In this kind of environment, the dominants are the polychaetes *Laeonereis acuta* and *Heteromastus similis* (Lana 1986).

The Patos Lagoon is a huge choked lagoon with a surface of 10,227 km². It stretches in a NE–SW direction from 30°S to 32°12 S, where in the south part there are 971 km² of estuarine area (approximately 10% of the lagoon). The estuarine region exchanges water with the Atlantic Ocean through a 20 km long and 0.5–3 km wide inlet (Asmus 1997).

As a consequence of reduced tidal influence (mean of 0.47 m) in the inlet and in the estuarine area, the salinity distribution lacks tidal variability but does correlate with wind forcing and variations in freshwater input on scales of hours to weeks (García 1997). The high frequency and low predictability of the salinity variations characterize the estuarine region of Patos Lagoon as an area chemically highly unstable (Niencheski and Baumgarten 1997).

The macrobenthic community in the estuary is composed of approximately 40 spp., most of which are r-strategists with pronounced seasonal and interannual variations in abundance. The long and narrow entrance channel with unstable bottoms, the reduced tidal oscillations, the unpredictable wind and precipitation patterns, that cause a general absence of conservative gradients of salinity, may account for the low diversity of the macrobenthic fauna in the estuarine area (Bemvenuti 1997a; Bemvenuti and Netto 1998).

The larger part of the estuarine shallow shoals is dominated by intertidal and shallow mudflats (<1.5 m), either with or without the occurrence of the widgeon grass *Ruppia maritima* beds, but with epibenthic microalgal growth and occasional aggregations of macroalgae (mainly *Enteromorpha* spp.). The

motile epibenthic organisms of mudflats are decapods like the blue crab *C. sapidus*, the shrimp *Farfantepenaeus paulensis*, the grapsid crab *Cyrtograpsus angulatus*, and the mud crab *Rhithropanopeus harrissii*. During summer juveniles of decapods and fishes exert a severe predation pressure on the macrobenthic community on estuarine mudflats (Bemvenuti 1997a).

The epifauna of mudflats is mainly represented by the opportunist gastropod *H. australis* (Hidrobiidae), which densities may exceed 40,000 in. m² and achieves highest biomass of 246 g m⁻², though pronounced spatial and temporal changes in density are common. The densities of amphipods, isopods, and tanaidaceans, which are typical of lower reaches of intertidal flats, increase as a result of macroalgal aggregations, which supply habitat, food, and shelter against predators (Bemvenuti 1997a).

The infaunal deposit feeder polychaete *L. acuta* reaches densities of 5,127 in. m⁻² and a biomass of 28.26 g m⁻² on intertidal and shallow mud habitats of Patos Lagoon estuary. The polychaetes *Nephtys fluviatilis* and *H. similis* have higher densities in mudflats but lower densities in the subtidal bottoms. Adults of both, *H. similis* and *L. acuta* attain refuge against predators by burrowing deeply (approximately 20 cm) in mud bottom habitats (Bemvenuti 1997b). The deeply burrowing pelecypod *T. plebeius* form patches in mudflats and the deeper tube dwelling tanaid *Kalliapseudes schubartii*, in spite of a severe predation by fishes and decapods, attains densities up to 10,000 in. m². This tanaid is a typical r-strategist, with intense reproductive activity in summer months, embryos marsupial protection, and intense recruitment, which maintains elevated densities in mudflats. In shallow habitats the pelecypod *Erodona mactroides* suffer high mortality during the first year and despite high densities (3,722 in. m⁻²), the mean biomass rarely exceeds 105 g m⁻² (Bemvenuti 1997a, b).

Macrobenthic invertebrates display diverse feeding habits but detritus appears to be an obligate food item for most species in mudflats. Infaunal deposit feeders like *L. acuta* and the epifaunal *H. australis* occupy the first level of consumers. In general, the decapods are typically omnivorous and opportunistic feeders, which exploit different trophic levels as food items become available. The predation of infaunal polychaete *N. fluviatilis* on *H. similis* represents an important intermediate link between the non-predatory infauna and epifaunal predators (Bemvenuti 1997c). The environmental stress of oligohaline estuarine systems are especially enhanced in the Patos Lagoon estuary and may have contributed a soft bottom community with wide trophic niches and abbreviated food chains (Bemvenuti 1997c).

The Quequén Grande River (approximately 38°S and 59°W) that drains a basin of around 7,800 km², has its estuarine area located between the cities of Necochea and Quequén, Buenos Aires Province, Argentina. Salinity shows remarkable fluctuations within the estuary (6–26), mainly to

tidal cycle but also due to freshwater inflow (López Gappa et al. 2001).

The infaunal macrobenthic community of intertidal flats of Quequén Grande River showed a very low species number, being mainly composed of four annelid species: the nereid *L. acuta*, the spionid *Boccardiella ligérica*, the tubificid *Ilyodrilus* cf. *frantzi*, and a species of *Capitella* (López Gappa et al. 2001). The low biodiversity seems to be a characteristic feature of the brackish water environments of the Buenos Aires Province, since there are just three infaunal polychaetes in Quequén Grande estuary (López Gappa et al. 2001), four in Samborombon Bay–Rio de la Plata (Ieno and Bastida 1998), and five in Mar Chiquita coastal lagoon (Olivier et al. 1972).

Bahía Blanca is a mesotidal coastal plain estuary in southwest of the Buenos Aires Province, Argentina. The estuarine area extends over about 2,300 km², with extensive tidal flats (1,150 km²). The mean water temperature is 13 °C, while salinity shows drastically differences (17) between the mouth and the head of the estuary (Perillo et al. 2001). In the mudflats the polychaetes *L. acuta* and *Eteone* sp. occupy the lower and middle mesolittoral. The mollusks *H. australis* and *T. plebeius* inhabit the middle and upper mesolittoral. Dense populations of *Chasmagnathus granulata* represent the third association in the upper intertidal of salt marshes and mudflats (Elias 1985). The grapsid crab *C. granulata* diet in marshes is dominated by pieces of marsh grass, while in the mudflats polychaetes, diatoms, ostracods, and nematodes predominate (Iribarne et al. 1997).

The diverse and rich mudflats in the coastal habitats of South America are fascinating areas for ecological works, but in general, these areas are still not very well known. An evident lack of knowledge exists on the spatial–temporal patterns of the populations and communities, and of estimates of the secondary production of the macrofauna. Besides, it is also necessary to accomplish studies on the main processes that govern the mudflats. The biological interactions and the related physical variables must also be tested through field experiments. It is also strongly advisable that more attention be paid to the processes of low predictability, such as those related to the success of the establishment and the recruitment of invertebrates with pelagic larvae. It is also urgent that the development of long-term studies, which are powerful tools that allow distinction among the anthropic effects of the natural effects, reach the variables and the resources of the ecosystems.

Seagrass Beds in South America

Joel C. Creed

Submerged macrophytes in general, and sea grass beds in particular, are recognized as ecologically important features

of the coastal zone (Larkum et al. 1989). Seagrasses actively construct and maintain extensive tidal flat structures in South America. Seagrasses provide food for herbivores, a habitat for other organisms, stabilize sediments, reduce or modify water movement and erosion, and present themselves as a substratum for colonization. In all these ways seagrasses add to the biodiversity and productivity of soft sediment environments and protect coastal areas from erosion.

The seagrasses of South America are not well known. Although South America's seagrasses continue to be the subject of some taxonomic debate, at least 10 seagrass species have been reported for the continent. Remarkably, seagrasses are almost absent from the Pacific coast of South America, the only seagrasses on the western continental coast being a couple of small populations of *Heterozostera tasmanica* in northern Chile at Coquimbo (Phillips 1992). Intriguingly, this species is otherwise known only from Australia and it has been suggested that these are remnants of formerly widely distributed Chilean populations (Phillips 1992). No seagrasses are known for Peru or Ecuador.

In the Caribbean, on the coasts and islands of Venezuela and Colombia, seagrasses can form very extensive meadows. *Thalassia testudinum*, turtle grass, is probably the most abundant seagrass, although it has not been reported further south. Plants are erect, with shoots up to 1 m, and grow in intertwined turf, forming extensive meadows on shallow sand or mud substrates from the lower intertidal to 20 m (Littler and Littler 2000). *Syringodium filiforme* (manatee grass) has a similar distribution, the southernmost known populations occurring in Venezuela. This species differs from *Thalassia* in having cylindrical, narrow leaves, which form canopies up to 45 cm high. It grows in sand and mud down to 20 m (Littler and Littler 2000).

Shoal grass, *Halodule wrightii*, which is common throughout the Caribbean and Brazil, has small supple, grass-like leaves. It is found growing on sand and mud from the intertidal down to 5 m (Littler and Littler 2000). It has a tropical–subtropical distribution and is found in Colombia and Venezuela and in Brazil from Ceará to Paraná States (Phillips 1992). The closely related *Halodule emarginata* is endemic to Brazil, occurring from Bahia to São Paulo States (Oliveira et al. 1983). Two other species of *Halodule*, *Halodule lilianeae* and *Halodule brasiliensis*, have been reported as endemic to Brazil but the separation of these three endemic species, based on leaf tip characteristics, has been questioned and some authors consider them forms of *H. wrightii* (Creed 2000).

Three *Halophila* (sea vine) species, *Halophila baillonii*, *Halophila engelmannii*, and *Halophila decipiens* grow in finer sands and sediments. *H. decipiens* is found in deepwater (to 30 m) in the southern Caribbean and has a tropical–subtropical distribution in the southwest Atlantic, stretching from

the Brazilian State of Ceará to Rio de Janeiro (Oliveira et al. 1983). *H. decipiens* can be found very shallow where turbid conditions exist because of run-off or pollution such as at Guanabara Bay, Rio de Janeiro, Brazil. *H. engelmannii* is a species restricted to the Caribbean, which has been reported at two locations in Venezuela and is only found down to 5 m depth (Littler and Littler 2000). *H. baillonii*, which grows deeper, is found throughout the Caribbean but has also been reported twice (in 1888 and in the 1980s, though not since) at Itamaracá Island in the northeast of Brazil (Oliveira et al. 1983).

Ruppia maritima, widgeon grass, is found sporadically from Venezuela down to Argentina, where it forms the southernmost populations of seagrasses in the world, at the Magellan Straits (Short et al. 2001). Such records reflect the species' wide latitudinal distribution and tolerance to variable environmental conditions, as it can be found growing in coastal lagoons and estuaries with salinities from 0 to 39. At the Patos Estuarine Lagoon in southern Brazil, a large (about 120 km²) area of *R. maritima* dominates the benthos and local primary productivity (Seeliger et al. 1997).

An unattached leaf of what was reported as *Zostera* (Setchell and Gardner 1935) has been found at Montevideo, Uruguay. Phillips (1992) commented that the leaf tip resembled that of *H. tasmanica* but that it was unlikely that it came from so far away as Chile. Seagrasses have not been reported for Guyana or Surinam, where information is very limited, although there is indirect evidence of seagrass beds in French Guyana. It is possible that *T. testudinum*, *H. wrightii*, *S. filiforme*, and *Halophila* spp. will be reported in the future, when surveys are carried out.

In South America, seagrasses are most abundant in the Venezuelan Caribbean region, in Colombia and sporadically at specific locations along the coast of Brazil. *Thalassia*, and to a lesser extent *Syringodium*, can be dominant, but in the Caribbean it is common to find *Thalassia* in mixed species stands with *Syringodium* or *Halodule*. *Halophila* species can also be found in monospecific or mixed species stands (Short et al. 2001). In Venezuela, *T. testudinum* is widely distributed on the western, central, and eastern Venezuelan coast as well as around the islands (Vera 1992). In Sucre State, northeastern Venezuela, *Thalassia* occupies 70% of the Cariaco Gulf, an area of about 290 km².

In classic models of Caribbean seagrass succession, *Halodule* (and sometimes *Syringodium*) are considered to be pioneers species (Gallegos et al. 1994). *Halodule* is better able to occupy mobile sediments and facilitates the subsequent growth of *Thalassia* and *Syringodium*. In Brazil, where the competitors *Syringodium* and *Thalassia* are absent, *H. wrightii* is the seagrass that most frequently occurs in shallow waters where it can form extensive monospecific meadows, such as at Itamaracá Island, Pernambuco State (Magalhães and Eskinazi-Leça 2000).

Of the deeper seagrasses, *H. decipiens* may be of great ecological importance because it may form extensive meadows most of which remain to be discovered. For example, at the Abrolhos Bank, southern Bahia State, Brazil, the suspicion that *Halophila* may be very abundant was recently confirmed in a Rapid Assessment Program (RAP) carried out in the region. Of 45 sites visited, *Halophila* was found at 18 at 5–22 m depth (Figueiredo, personal communication). As these sites were distributed over an area of about 6,000 km², the potential importance of *H. decipiens* in the region, especially in terms of primary productivity, could be enormous.

South American seagrasses are often found spatially close by or closely trophically linked to other marine and coastal ecosystems and habitats and this juxtaposition results in higher diversity (Creed 2000). *Halodule*, *Syringodium*, and *Thalassia* are associated with shallow habitats without much freshwater input, such as reefs, algal beds, coastal lagoons, rocky shores, sand beaches, and unvegetated soft-bottom areas and nearby mangroves without too much salinity fluctuation. *Halophila* is associated with deeper reefs, algal and marl beds, and deeper soft-bottom vegetated areas. *R. maritima* can be found in lower (coastal lagoon, estuary, fish pond, mangrove, salt marsh, and soft-bottom unvegetated) and higher salinity (coastal lagoon, salt pond, soft-bottom unvegetated) habitats. Numerous physical, chemical, and biological interactions take place between these habitats.

Seagrass beds in South America are known to be important habitat for a wide variety of plants and animals and an enormous diversity of organisms is associated with the South American seagrasses. Groups that contribute most to the richness of seagrass systems in the Caribbean are polychaetes, fishes, amphipods, decapods, foraminifers, gastropods, and bivalve mollusks, macroalgae, and diatoms. Oligochaetes, nematodes, coelenterates, echinoderms, bryozoans, and sponges are other ecologically important groups. About 540 taxa (to genus or species level) of organisms are known to be associated with the Brazilian seagrasses (Creed 2000). Over 100 macrofaunal and 46 epiphyte floral taxa have been identified in or on *Halodule* at Itamaracá Island, Brazil (Alves 2000). In Venezuela, at the Caiaco Gulf, 112 mollusk species have been identified in *Thalassia* beds and 127 species of macroinvertebrates are associated with *Thalassia* beds at Mochimba Bay (Vera 1992). 372 fish species have been reported associated with seagrasses at Bahia de Chengue, Colombia (UNESCO 1998). Corals are also found in seagrass meadows throughout the Caribbean and Southwest Atlantic. For example, the corals *Meandrina brasiliensis* and *Siderastrea stellata*, the latter being endemic to Brazil, grow unattached in *H. wrightii* beds and are sold as souvenirs locally (Creed 2000). At Bahia de Chengue, on the Caribbean coast of Colombia, corals of the genera *Manicina*, *Siderastrea*, *Millepora*, *Diaporina*, *Porites*, and *Cladocora* grow within the *Thalassia* beds (UNESCO 1998).

Leaves of seagrasses are a substratum for epiphytic algae and invertebrates. Macroalgae in seagrass beds typically consist of members of the orders Dasycladales and Caulerpales, calcified chlorophytes (*Acetabularia*, *Halimeda*, *Penicillus*, *Udotea*) and rhodophytes such as *Acanthophora* and *Hypnea*. Calcareous chlorophytes which grow in the sediment in seagrass beds are major contributors to the sedimentary cycles of tropical South American shallow-water environments. Other macroalgae grow unattached within the bed as “drift” algae.

Seagrasses are extremely palatable to herbivorous fish, such as parrotfish and surgeonfish, and urchins. Seagrasses are not thought to be chemically defended, and their palatability is such that pieces of *Thalassia* have often been used in *in situ* bioassays to determine relative herbivore pressure on reefs. Such studies have elucidated the strong trophic link between coral reefs and seagrass beds in South America, which is best exemplified by grazing “halos.” Herbivorous fish that graze seagrass beds abutting them create such halos. Because of predators, the herbivores do not distance themselves greatly from the reef; hence the halo. Sea urchins such as *Lytechinus variegatus* and *Tripneustes ventricosus* are also major herbivores of seagrass blades, consuming *Thalassia* and *Halodule*. Echinoids have a double role in the trophodynamics of the seagrasses. Not only are they major grazers on sea grass but also their feces are a food source for detritivores.

Because of the diversity of species assemblages, seagrass beds have been recognized as among the most productive fisheries areas in the Caribbean and southwestern Atlantic. Most of the living resources are associated directly (as food, habitat, or foraging areas) or indirectly (through export of primary production, larvae and juveniles to other habitats) with seagrass areas. Soft-bottom demersal fisheries exploit scianoids, mullets, sharks, penaeid shrimp, conch, loliginid squid, and octopods over seagrass beds and rough-bottom demersal and coral reef fisheries exploit snappers, groupers, grunts, and lobsters which are also found in seagrass habitat. Inshore pelagic fish may also hunt over the seagrass meadows. Other resources which are seagrass associates are turtles, crabs, oysters, sea urchins, sponges, stony corals, and seaweeds.

While the seagrasses of South America contribute to coastal protection, local productivity, and thus fisheries, there is hardly any information available about the value of seagrasses to the local economy. Economically important fish species such as the bluewing searobin (*Prionotus punctatus*), whitemouth croaker (*Micropogonias furnieri*), and mullet (*Mugil platanus*) are found and fished in Brazilian sea grass beds (Creed 2000). Local fisheries exploit commercially important crustaceans such as blue crabs (*C. sapidus*), stone crab (*Menippe nodifrons*), lobster (*Panulirus argus* and *Panulirus laeviscauda*), and shrimp (*Penaeus brasiliensis*

and *Penaeus paulensis*), all of which are associated with seagrass beds. Other shellfish which are commercially collected from seagrass beds are clams (*Anomalocardia brasiliana*, *Tagelus plebeius*, *Tivela mactroides*), volutes *Voluta ebraea*, rockshells (*Thais haemastoma*), oysters (*Ostrea puelchana*), and cockles (*Trachycardium muricatum*) (Creed 2000). In Chile, the Chilean scallop (*Argopecten purpuratus*) preferentially settles in *H. tasmanica* beds (Aguilar and Stotz 2000).

Two threatened species, which feed directly on seagrasses from the Caribbean to Brazil, are the green turtle *Chelonia mydas* (Creed 2000) and the West Indian manatee *Trichechus manatus* (Magalhães and Eskinazi-Leça 2000). Both have benefited from specific conservation action sponsored privately and by the Brazilian Environmental Agency IBAMA (green turtles by the Projeto TAMAR and manatees by the Projeto Peixe-Boi Marinho). The black-necked swan *Cygnus melancoryphus* and the red-gartered coot *Fucila armillata* also feed directly on *R. maritima* in southern Brazil and Argentina but are not endangered (Seeliger et al. 1997). Recently, the semi-aquatic capybara (*Hydrochaeris hydrochaeris*), which is the world's largest rodent, was observed feeding on *R. maritima* near Rio de Janeiro.

As seagrass beds occupy shallow, nearshore depositional environments, they are highly susceptible to damage by human activity. Pollution from land-based sources varies from country to country. Activities related to human settlements, agriculture and industry have been identified as major contributors to the pollutant loads reaching coastal and marine waters. The greatest threats are sewage, hydrocarbons, sediments, nutrients, pesticides, litter and marine debris, and toxic wastes. River loads are enhanced by erosion of watersheds caused by deforestation, urbanization, and agricultural activities. On the continent, the impact of moderate cultural eutrophication on seagrass ecosystem results in greater epiphyte levels and lower shoot density, leaf area and biomass of the seagrass. Sewage pollution has been reported throughout the continent.

Direct reports of impacts on seagrasses on the continent are few. However, pollution by heavy metals from sporadic mining and metalworking activities, by polychlorinated biphenyl congeners and organochlorine compounds and by nutrients from agricultural runoff and sewage discharge have all been reported. Effects of physical damage by anchors and trampling on seagrass and associated macroalgae have also been identified. Loss of water area, because of sediments produced after erosion due to deforestation, infilling for construction and dredging activities have also reduced the area occupied by South America's seagrasses. *R. maritima* has suffered from reduced freshwater inputs because of rice irrigation, population growth and lock construction (Seeliger et al. 1997). It has been estimated that 100% of Chilean and Argentine, and 40% of Brazilian seagrasses are “highly threatened” (Creed 2002).

Thirty-six percent of Brazil's seagrasses are "moderately threatened" and 24% are in "low threat" areas.

There are five research groups currently studying seagrasses in South America (Colombia: Instituto de Investigaciones Marinas y Costeras (INVEMAR), Santa Marta; Venezuela: Instituto de Tecnología y Ciencias Marinas, Universidad Simón Bolívar, Caracas; Brazil: Laboratório de Ecologia Marinha, Universidade do Estado do Rio de Janeiro; Universidade Federal Rural de Pernambuco, Recife; Fundação Universidade Federal do Rio Grande – FURG, Rio Grande). Researchers are studying basic biology and ecology of the sea grasses in their region and participate in important regional or global programs. For example, in Venezuela, seagrasses are being monitored by the Coastal Ecosystem Productivity Network in the Caribbean (CARCOMP) Program at the Parque Nacional Morrocoy, which has been the subject of numerous studies, and at Punta de Mangle on the Isla de Margarida. There is also a CARICOMP site in Colombia at Bahía de Chengue within the Parque Natural Tayrona. These sites are used for comparisons of productivity throughout the Caribbean region (UNESCO 1998). In Brazil three seagrass beds were recently included in the Global Sea grass Monitoring Network – SeagrassNet. At the Patos Lagoon, Rio Grande do Sul State, Brazil, a site where *Ruppia* beds are widespread and which has been studied for the past 25 years, is included in the Brazilian Long-term Ecological Research Program (PELD). In one way or another, all these programs aim to provide a perspective on long-term change in seagrass growth and productivity at regional or global scales.

Marine Mammals of South America

André S. Barreto

The coastal environments of South America are inhabited by a great diversity of marine mammals, from the orders Sirenia (manatees), Carnivora (otters and pinnipeds), and Cetacea (whales, dolphins, and porpoises). Sirenians and cetaceans have a fully aquatic existence, hardly ever coming ashore intentionally, while otters and pinnipeds leave the water from time to time, specially during the reproductive season. In order to better describe the species that occur along the coast of South America, each order will be dealt separately.

Order Sirenia: Manatees

This order is divided in two families, Dugongidae and Trichechidae, and only the latter is represented in South America, by two species: *Trichechus inunguis*, the Amazonian manatee, and *Trichechus manatus*, the West Indian manatee. The Amazonian manatee is restricted to freshwater on the Amazon basin and will not be discussed here. The West

Indian manatee inhabit coasts, estuaries, and major rivers from Rhode Island, United States, to Alagoas, Brazil (Reeves et al. 2002). However, the distribution is not continuous, with gaps along this range. Despite the manatees' ability to move thousands of kilometers along continental margins, strong population separations between most locations were observed on the phylogenetic structure of *T. manatus* (Garcia-Rodriguez et al. 1998). This is very important for its conservation, since that although the species has a relatively wide distribution, there is low genetic exchange between locations. Therefore, they should be managed separately. In Colombia this species is considered endangered, with hunting apparently increasing but incidental capture with nets still representing the species' major direct threat (Montoya-Ospina et al. 2001). On the Venezuelan coast, a remnant manatee population exists in Lake Maracaibo, but none was found to occur along the more than 1,500 km of Caribbean coastline (O'Shea et al. 1988). It is considered as critically endangered on the coast of Brazil, being subject to intentional and accidental mortalities and degradation of its habitat. Total populational size on the Brazilian coast is estimated to be approximately 500 animals (IBAMA 2001).

Sirenians are the only herbivorous aquatic mammals, feeding on a large variety of coastal and freshwater vegetation, including several species of seagrasses, floating freshwater plants (*Hydrilla* and water hyacinths), and even the leaves and shoots of emergent mangroves (Berta and Sumich 1999). In Brazil, it was found that shoalgrass, *H. wrightii*, is an important item of the manatee diet, being consumed together with many species of algae from the divisions Chlorophyte (*Caulerpa cupressoides*, *Caulerpa prolifera*, *Caulerpa racemosa*, *Caulerpa sertularioides*, *Halimeda opuntia*, and *Penicillus capitatus*) and Phaeophyte (*Dictiopteris delicata*, *Dictyota* sp., *Lobophora variegata*, and *Sargassum* sp.) (Cardoso and Picanço 1998). Since each manatee may consume from 29.5 to 50 kg of plants per day (Würsig et al. 2000), they are probably important in the ecology of seagrass beds.

Order Carnivora

Different from their terrestrial counterparts, aquatic carnivores are well adapted for foraging in water, being excellent swimmers and divers. Along the coasts of South America they are represented by one species from the Mustelidae (marine otter, *Lutra felina*), two of the Phocidae (southern elephant seal, *Mirounga lionina*, and leopard seal, *Hydrurga leptonyx*) and five of the Otariidae family (Juan Fernandez fur seal, *Arctocephalus philippii*; South American fur seal, *Arctocephalus australis*; Galápagos fur seal, *Arctocephalus galapagoensis*; Galápagos sea lion, *Zalophus californianus*; and South American sea lion, *Otaria flavescens*).

The Marine Otter is the only species of the genus *Lutra* that lives exclusively in marine environments. It ranges from the coasts of central Peru south to Cape Horn and along the Atlantic coast of Tierra del Fuego (Reeves et al. 2002). Very little is known about this small otter, due to its secretive nature. It is believed that they feed mainly on shellfish, near-shore marine fish, and freshwater prawns.

The seven species of pinnipeds mentioned above are the ones most commonly found on the South American coasts. Vagrants of other species have been reported, but are not regularly observed. The Juan Fernandez and Galápagos fur seals, the Galápagos Sea Lion and the Leopard Seal have breeding colonies in islands around South America, while the other three species have breeding sites on the coast and are more commonly found in the nearshore environment. The breeding systems of these species are polygynous, with males competing among them for territories in the otariids or for dominance in a hierarchical system in the elephant seal. All species feed on fish (demersal or pelagic), cephalopods (octopuses and squid), and more occasionally crustaceans.

Order Cetacea

This order is divided in two suborders, Mysticeti (baleen whales) and Odontoceti (dolphins, porpoises, and sperm whales). Cetaceans occupy marine and freshwater environments in tropical, temperate, and polar regions. Of the 78 accepted species of cetaceans 50 are found on waters around South America (Jefferson et al. 1993). They range from coastal specimens that do not go into waters deeper than 30 m, such as the Franciscana, *Pontoporia blainvillei*, to truly oceanic, deepwater species that only occasionally come closer to shore as the Sperm Whale, *Physeter macrocephalus*. Only seven species of cetaceans are endemic to South America: the Amazon River Dolphin, *Inia geoffrensis*; Commerson's Dolphin, *Cephalorhynchus commersonii*; the Chilean Dolphin, *Cephalorhynchus eutropia*; Tucuxi, *Sotalia fluviatilis*; the Estuarine Dolphin, *Sotalia guianensis*; Peale's Dolphin, *Lagenorhynchus australis*; and Burmeister's Porpoise, *Phocoena spinipinnis*.

They occupy different trophic levels, with whales being trophically more basal, consuming zooplankton, and most odontocetes being top predators consuming fish, cephalopods and even other marine mammals, as in the case of Killer Whales, *Orcinus orca*. Considering their large size and abundance, they probably play an important role in the structuring of marine communities. Even allowing for the unpredictability of strandings, cetacean carcasses are an important food source for terrestrial and benthic scavengers (Katona and Whitehead 1988).

The occurrence of baleen whales on the coast of South America is restricted to the second semester of the year, during the austral winter. Humpback and Southern Right Whales (*Megaptera novaeangliae* and *Eubalaena australis*,

respectively) use coastal areas and islands as breeding grounds, but do not feed while there. Thus their impact on local ecosystems is relatively low, apart from being a food source for other top predators (e.g., sharks) or scavengers when dead. However, seagulls have been observed feeding on skin from the back of Right Whales when they are on the breeding grounds (Rowntree et al. 1998; Groch 2001). On the other hand, odontocetes are present all year round and have important roles in the structuring of aquatic food chains, consuming a great variety of organisms. As a rule they are opportunistic, consuming the resource that is most available, including pelagic and demersal fish, cephalopods (octopuses and squids). However, some species are highly specialized, as the Sperm and Beaked Whales (families Physeteridae and Ziphiidae), which feed almost exclusively on squids.

Cross-References

- [Beach Processes](#)
- [Coral Reefs](#)
- [El Niño–Southern Oscillation \(ENSO\)](#)
- [Estuaries](#)
- [Human Impact on Coasts](#)
- [Mangroves, Ecology](#)
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- [Sandy Coasts](#)
- [South America, Coastal Geomorphology](#)
- [Vegetated Coasts](#)

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South America, Coastal Geomorphology

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Investigation into the coastal geomorphology of Latin America does not have a lengthy history, but there are some indications that Latins and others are turning their attention to this important area (Psuty 1970; Tavares Corrêa 1996; Bittencourt et al. 1999; Klein et al. 2002). Large portions of the coastal zone remain unstudied in detail, and frequently only general descriptions exist (Putnam et al. 1960; Dolan et al. 1972; Bird and Schwartz 1985). Questions about regional correlations of depositional and erosional features must await basic research into topical problems. However, the following interpretation of the coastal geomorphology of South America is presented as a base upon which to build many future layers of information.

Colombia

From Venezuela to the east to the border with Panama to the west, the Caribbean coast of Colombia is approximately 1,030 km in length. The Guajira Peninsula is the northernmost arm of the Andean mountain system, and the extreme north-eastern tip is fronted by an active coral reef. Some evidence has been presented (Anderson 1927) that there are raised coral platforms and fossil mollusks representing tectonic displacement of the land terminus. Sandy beaches without accompanying coral line the northwestern margin of the peninsula southward to Cape San Juan de Guía, where crystalline cliffs occur with pocket beaches.

South of Santa Clara the coastal zone is dominated by the delta of the Magdalena River. There is a broad arcuate delta in this area with several active distributaries. Great quantities of sediment are transported to the ocean to be reworked. Mangrove-covered mudflats are characteristic of this area, as is beach ridge and chenier topography. The fluctuating activity of the distributaries is responsible for episodes of accretion and erosion of the shore. Beach ridges and cheniers (*q.v.*) on the delta show the old coastal alignments, and their truncated forms provide evidence of reorientation of the coastline in past times. Near the Colombian–Panamanian border the Atrato delta provides the source of the sediments that form the shore features, but there is little change. Mangrove forest, distributary channels, and sand ridges along the distal margins continue to characterize the coast.

The 1,300 km-long Pacific coast of Colombia begins as a high-cliffed coast at the Panamanian border. A portion of the flank of the Andes comes to the sea to create steep, vegetation-cloaked precipices plunging directly into the water. West (1956) described the region from south of Cape Corrientes to the border with Ecuador as having short drainage systems leading from the steep mountainous ridges, a shore consisting of fluvially derived sediments, and, with the exception of three places where the coast is cliffed, narrow sandy beaches lying between the ocean and dense mangrove forests. Large mudflats occurred in front of the beaches and in front of the fluvial plain at the foot of the mountains. In 1995, Martinez et al. reported for the first time in the periodical literature that most of the Cape Corrientes to Tumaco shore consisted of a series of barrier islands. Subsequently, Martinez et al. (2000) and Morton et al. (2000) described barrier washover events resulting from a relative sea-level rise. They attributed the relative rise in sea level variously to long-term subsidence, short-term seismic subsidence, and El Niño events.

Ecuador

The coast of Ecuador, including the shores of the Puna, Jambeli, and Galapagos islands, is about 2,500 km in length (Ayón and Jara 1985). The northern coast is drenched with rainfall, and supports a dense vegetation down to the shore. However, at the southern margin of the country, the coast is stark and almost devoid of all vegetation. At both extremes, the coast consists of high cliffs fronted by a sandy fringing beach or pocket beaches. The high cliff extends consistently from near the border with Colombia southward to near Santa Elena. Near the latter location the coast takes a right-angle bend and trends southeastward. Along this portion of the coast, Sheppard (1930) described a series of marine terraces rising step-like to an elevation of 60 m. He interpreted these features as evidence of tectonic displacement. Ayón and Jara (1985) reported continental uplifting in this region as high as 200 m, evidenced by Pleistocene beachrock benches (locally called *tablazos*).

The Gulf of Guayaquil is an estuarine system whose shore is fringed by dense stands of mangrove on mudflats (Twilley et al. 2000). Near the distal margins of the estuary, beach-ridge systems make up several of the exposed points. However, the southern margin of the estuary remains a cliffed coast. Southwestward toward Tumbes in Peru, a high-cliffed coast dominates with either a sandy fringe or pocket beaches.

Cliffs, shore platforms, and pocket beaches border the volcanic Galapagos Islands (Ayón and Jara 1985). Eroded craters have resulted in steeply plunging cliffs, and shore platforms are cut into former lava flows. Uplift of some of the islands by about 100 m is shown by exposed sedimentary rocks and submarine lavas.

Peru

The hyperaridity of coastal Peru is responsible in large part for the specific geomorphologic features found there. The lack of precipitation presently limits the contribution of fluvially transported sediments to the littoral zone, and thus limits the development of beaches. During the Holocene, as the sea level was rising and melt-waters were coursing through the river valleys leading to the Pacific Ocean, there were great quantities of sediment discharged to be reworked by the waves and currents. Large sediment volumes were sent into the small estuaries at the river mouths and onto the narrow continental shelf. However, only a few rivers continuously reach the sea at present, and the situation has changed from that of several thousand years ago. The rise in sea level has continuously encroached upon the small estuarine locations, and scattered tectonic displacement has produced further alterations of the coastal features.

Peru is characterized by the presence of a 2,300 km-long cliffed coast interspersed with pocket beaches and beaches fronting river mouths. The pocket beaches tend to be located where more easily eroded rock formations of sandstone, shale, or marl are exposed at the shoreline. The rocky promontories are usually composed of crystalline rock units. In a few places the cliffed coasts are cut into colluvial deposits that have moved down spacious interfluvial areas during the Quaternary. At locations where the cliffs are in colluvial gravels and cobbles, a narrow fringing beach usually exists.

South of Pisco there is little or no coastal plain, and the coast usually consists of towering cliffs or small crescent-shaped embayments with narrow beaches. From Pisco north to Chiclayo there are a number of rivers that reach the sea and contribute to the development of localized coastal plains. According to Parsons and Psuty (1975), the mouths of the major river valleys have had a similar geomorphic history and a similar assemblage of landforms. Archaeologic evidence suggests that the fluvial plains found in these river mouths obtained their present characteristics about 1,000–1,500 year BP. At that time the embayments were filled, and the leading margin of the fluvial plain included a classic beach/dune profile. The profile has been migrating inland since that period over a narrow backmarsh that has formed between the beach and the fluvial plain. Coastal sand dunes reach impressive size: 6 m high at the shore fringing the Sechura Desert and 20 m high at Negritos (Bird and Ramos 1985). At the southern margins of the valleys, there is frequently a dynamic beach area where sand is being moved inland by the prevailing winds out of the southwest. The coastal dune ridge is frequently enlarged in this area and transformed into longitudinal dunes extending inland. In some instances, barcan dunes break off the sand sheet and migrate independently across the terrace surfaces.

Coastal displacement is evidenced in many locations of coastal Peru by raised shore platforms and raised beaches. Bird and Ramos (1985) reported Quaternary marine terraces (tablazos), incised by transverse valleys (quebradas), south of Zorritos. Other investigators have identified as many as 10 coastal terraces at elevations of up to 250 m at the mouth of the Ica River (Broggi 1946) and up to 75 m in the Sechura Desert area. However, there is some question whether the displacement is part of regional uplift or local movement. Several detailed inquiries (Craig and Psuty 1968; Parsons and Psuty 1975) have shown that the movement is local because the platforms do not extend great distances. Further, flanking terraces in the valleys are Pleistocene depositional features. Most of the major river valleys contain no evidence of poststillstand uplift (Psuty 1978). In northern Peru, Richards and Broecker (1963) collected marine shells from several terraces that implied emergence of the platforms. A low terrace at 4.5 m had shells collected on its surface dated at 3,000 yr. BP, whereas a high terrace of 75 m had shells assayed beyond 30,000 year BP.

Chile

The coast of Chile trends parallel to the uplifted longitudinal coast ranges for about 4,400 km, or a total length of 35,000 km when all of the inlets and islands are considered (Araya-Vergara 1985). The northern two-thirds of coastal Chile has characteristics similar to those of Peru. The Andean coastal range comes down to the sea, and the cliffed coast is interrupted by small pocket beaches or alluvial embayments where infrequent streams lead down to the shore. Numerous terraces appear along the cliffed headlands fronting the foothills. According to Börgel (1967), there are a series of steps reaching to 200 m, thought to be terraces of abrasion, north and south of Valparaíso. Paskoff (Fuenzalida et al. 1965) believes that the highest terraces found in northern Chile at 250–400 m are probably Pliocene, whereas those below 250 m are likely to be Pleistocene. Numerous investigators (Fuenzalida et al. 1965) have identified the considerable number and variety of terraces found in coastal Chile. It is unlikely that the terraces can be considered wholly as products of eustatic sea-level changes, as was suggested in the early investigations. Rather, the lack of uniformity in number, elevation, and kind points to localized tectonic events. However, one terrace at the 85/100 m level does appear to persist throughout much of northern and central coastal Chile. It is considered to have extensive deposits of marine gravel on its surface and also marine mollusks. It is possible that this surface represents an episode of regional displacement.

Rainfall increases southward in Chile, but there are no major coastal alluvial plains developing. The coast range comes to the sea and provides for only modest embayments.

Pocket beaches prevail south from Valparaíso to Chiloé. However, the shores near the mouth of the Bio-Bio River and south of the Mataquito River are well developed, with sizeable active dunes penetrating inland. In this region of the Aruaco Province, Tavares Corrêa (1996) has investigated the potential of coastal dune management and “According to the aeolic potential, the dune environment of the three sectors were classified as for Conservation (South Sector), Stabilization (North Sector) or a combination of them (Central Sector). At the same time, these sectors were divided in small units which were assigned for activities like afforestation, stabilization, conservation, agriculture and grazing.” There is also estuarine development at Valdivia, where a small fluvial plain has accumulated at the mouth of a river and a narrow belt of coastal features bounds its shoreward margin. Principal Component Analysis of sand from 16 beaches in the vicinity of Valdivia by Pino and Jaramillo (1992) identified two main groups, fully reflective sites with coarser particles and steeper profiles as compared with dissipative sites with intermediate characteristics. Jaramillo et al. (2002) subsequently studied the reflective–dissipative effect upon sandy beach microfauna on both sides of a concrete seawall in the same region.

South of the latitude of Puerto Montt the coastal configuration is dramatically changed. The coastal range becomes broken. The island of Chiloé retains many of the characteristics of the northern coast, but to the south the coast is altered by the processes of glaciation. A fjord coast is present south of 43°S latitude, with many channels extending entirely across the crestal portion of the coastal range into the interior passage. Many of the pocket beaches are cobble-strewn and exist only in sheltered areas. In this region of tidal marshes associated with environments of glacial retreat, earthquake-induced subsidence, and braided river channel migration, Reed (1989) investigated variations in marsh topography and morphology, and found “distinct morphological differences between the marshes developed on glacial outwash and river deposits, and those formed after a major subsidence event.”

Weischet (1959) reports that southern Chile has several terrace levels, but they are not continuous through the region. They may be dissected remnants of a larger surface or products of local tectonic activity. However, the degree of glaciation is so thorough in this area that it is the characteristics of glaciation that must be considered rather than those of coastal processes. Patagonia is a region of straits, fjords, drumlins and roches moutonnées, with glaciers that reach the sea (Araya-Vergara 1985). In an eighteenth century readvance the bay at Lagun de San Rafael was overrun by glaciers.

Argentina

Argentina is characterized primarily by a 5,700 km-long cliffed coast with a narrow beach zone before it. The cliffs

vary from only a few meters to the spectacular elevations of greater than 500 m south of Comodoro Rivadavia. Occasional areas of sedimentary accumulations do exist either as beaches fronting the cliffs or as substantial areas of beach ridges and coastal dunes.

The Río de la Plata estuary, 15,000 square km in extent, dominates the northern portion of the Argentinian coast (Schnack 1985). From the mouth of the river to Cabo San Antonio, the shore is a tidal mudflat. Wave energies are low, and the fine-grained sediments derived from the fluvial system are not reworked into a beach form. However, Urien (1972) has interpreted beach ridge and chenier forms that were created in the period of 7,000–3,000 year BP along the southern shore of the estuary. These features were part of the sand wedge that was being pushed up the continental shelf as the sea level was rising, and they developed prior to the silting of the estuary. Urien suggests that tectonic displacement of these forms has raised them to elevations of 9 m, where they and a marine terrace form the margin of Samborombón Bay. From Mar del Plata to Bahía Blanca, the coast consists of a low-cliffed shore with a narrow beach before it. Occasionally there are large dune fields leading from the beach. This portion of Argentina is an extension of the Pampas coming to the coast. South of Bahía Blanca the Negro and Colorado rivers transport considerable quantities of sand to the shore, and the beaches are extremely broad. The alluviation is not complete, however, for there are rocky islands and points located between these two river mouths. The Colorado delta is extensive, and there is evidence of coastal aggradation in the form of beach ridges and distributary elongation. Though the Negro delta is not nearly so extensive, it manages to fill its estuary. For further studies of sediment transport along the coast of Argentina see Isla (1997), Mar Chiquita; Kokot (1997), Punta Médanos; and Cuadrado and Perillo (1997), Bahía Blanca.

With the exception of well-developed beaches and associated landforms at the Gulf of San Marcos and the Gulf of San Jorge, the southern half of Argentina is a cliffed coast. Investigators have identified a number of terraces that have been used to describe tectonic or eustatic displacement. Terraces ranging to 140 m have been noted in Patagonia. A 9 m terrace at Comodoro Rivadavia has been dated at 3,000 year BP (Richards and Broecker 1963), and has been used to suggest local tectonic movement. Dating beach gravels and beach ridges along the Patagonia coast, Rutter et al. (1990) found the penultimate or older glaciation shore at 24 and 41 m above mean sea level, the “intermediate” age shore at between 20 and 28 m, and “young” Holocene beach ridges at 8–12 m. Urien (1970) has indicated that a 9 m terrace also exists in Tierra del Fuego along parts of the Río de La Plata estuary and bordering the Paraná delta. However, he cautioned against attempting any broad correlations. More recent radiocarbon dating

with corresponding altimetry over present sea level, at 15 locations along 3500 m of the Argentina coast, by Codignotto et al. (1992) indicated relative uplift rates between 0.12 and 1.63 m/1000 year during the Holocene. A twentieth century sea-level rise of from 1.6 to 3.5 mm/year along much of the same sector has been reported by Lanfredi et al. (1998).

From Santa Cruz to the eastern tip of Tierra del Fuego the cliffs are cut into glacial morainic material. Occasional outcrops of bedrock are noted, as are pocket beaches (Etchichury and Remiro 1967). At Punta Dungeness there is an excellent series of beach ridges created where currents converge at the point. The more protected western margin shows considerable accretionary history, whereas the exposed southeastern flank gives evidence of truncation of the ridges and extensive dune fields extending inland. Schnack (1985) has described the coast of Islas Malvinas as being indented and cliffy, with some raised Quaternary shore terraces.

Uruguay

The 600-km-long coast of Uruguay is extremely diverse for such a relatively short sector. The northern area consists of an extension of the barrier island system of southern Brazil. The sandy beach continues into Uruguay but narrows and becomes discontinuous, so that it becomes a series of sandy embayments set into Uruguay's crystalline Eastern Highlands Shield (Jackson 1985). These embayments are not products of marine processes but, rather, the prior irregular topography of the shield encroached upon by a rising sea. In several places, the embayments contain small lagoons behind a sand barrier. This feature is the product of incomplete filling of small estuaries along the margin of the shield (Chebataroff 1960).

From near Maldonado westward, the coast is the margin of the Río de la Plata estuary. For nearly this entire length there is a cliffed coast with a sand beach lying at its base. Delaney (1963, 1966) suggested that there are terraces cut into this cliff. And it was proposed by Urien (1970) that the beach sand fronting the cliffs is the product of local erosion of the bluffs rather than fluvial transport. The exposure of this portion of Uruguay to the southeast would tend to allow greater wave energies to reach this shore and favor the necessary erosion and sorting responsible for sandy beaches, while much of the rest of the estuary is a shallow muddy tidal flat. Occasionally mudflats dominate along the cliffed coast. The Río de la Plata is formed by the confluence of the Paraná and Uruguay rivers, which flow between Uruguay and Argentina, and owing to the shallow depth across this region, navigation can only be accomplished by dredging channels for large ships (Vieira and Lanfredi 1996).

Brazil

The great size of Brazil allows for considerable diversity of coastal exposure and geomorphologic development. There are three principal portions of the 9,200 km-long coast. The first is the area influenced by the Amazon River and its sediments; the second is the narrow coastal margin fringing the huge Brazilian Shield, creating an escarpment nearly adjacent to the ocean; the third is the southern area, where considerable quantities of sediments have accumulated to provide a barrier island formation.

The mouth of the Amazon River is a great estuary stretching for nearly 1,600 km (1,000 miles). Large quantities of sand and especially silt and clay are discharged by the river and accumulate along the shore margins. From the border with Surinam eastward to the Gulf of San Marcos, the fine-grained sediments blanket the shore and are cloaked with mangrove. The Gulf of San Marcos is another, much smaller, estuary that similarly contributes large quantities of fine-grained sediment.

East of the Gulf of San Marcos the shore begins to be characterized by sandy beaches lying before low hills. The sand beaches are interspersed with mangrove stands that dominate where local deltaic buildout occurs in association with short drainage systems leading off the eastern margin of the Brazilian Shield. Beginning in Rio Grande do Norte and continuing southward to the coastal margin of Alagoas state, the beach zone is severely attenuated. The dry climate and the short drainage systems limit the transport of sediment to the ocean margin. Further, this portion of Brazil is bordered by a fairly extensive coral reef. Where beach sediments have accumulated, there are also likely to be well-developed dune forms migrating inland over terrace surfaces.

South of Recife the coast is cliffed. The combination of cliffed coast and the presence of coral reef extends for 480 km (about 300 miles). Near Recife a small promontory has been investigated for evidence of a high sea level. Van Andel and Laborel (1964) have dated a fossil biogenous limestone that is encrusting granite a few meters above modern sea level as having been active 1,200–3,600 year BP. The authors interpreted these dates to hypothesize that sea level was higher in that time period.

The sandy beach backed by an escarpment begins near the Alagoas–Sergipe border and continues south to Rio Grande do Sul. The beach often broadens in large curvilinear embayments, and there may be local mangrove stands, beach ridges, and deltaic buildout. However, the escarpment dominates the horizon, and the coastal geomorphic features occupy a small niche on the continental margin. Bittencourt et al. (1999) have proposed large-scale tectonic flexure as controlling geomorphic characteristics in this region. In the state of Paraná there is an extensive area of beach ridge development associated

with the Maciel river. These ridges appear to resemble cheniers in that they are bounded by clayey deposits rather than forming a broad sandy surface. The beach ridges attain elevations of 10 m in their interior location, and gradually decrease to elevations of 2–3 m near the shore. Bigarella (1965) believed this was further evidence for a fluctuating and generally falling sea level. Where the state of São Paulo borders that of Paraná, Tessler and de Mahiques (1993) have shown that coastal sedimentary features, such as spits and sandbanks, are clearly indicative of longshore sediment transport directions.

Along the coast of Santa Catarina state there are a number of headland bay beach systems bounded by headlands or rock outcrops, where the shore assumes a curvature form. Klein et al. (2002) have investigated short-term beach rotation processes at three such sites. They found that erosion–deposition events varied with incident wave direction. And while there was no loss of sediment from a sector, original realignment was attained with a return to the prior wave direction.

The coastal margin of the state of Rio Grande do Sul is distinct from the rest of Brazil—it consists of a classic barrier island–lagoon sequence. Delaney (1963, 1966) described the geologic origin of the barrier island sequence as occupying a unique position in a geologic depositional basin with sediments being transported into it from several directions. Toldo Jr. et al. (2000) have suggested an average depositional rate of 0.75 mm/yr. during the Holocene in the Lagoa dos Patos estuary. Certainly the positive sediment budget that existed in this area with the changing sea level caused the particular assemblage of broad sandy beach extending along the coast for 640 km and incorporating 120 km-wide (Cruz et al. 1985) beach ridge systems and large coastal dunes reaching 25 m in elevation. The northern margin of this Holocene coastal plain comes against a terrace surface with elevations of 15 m. It is probably of Pleistocene age, but whether its origin is wave-cut is unknown.

French Guiana, Surinam, Guyana

The coast of what was once referred to as the European Guianas is somewhat similar for its entire length of over 1,100 km. Basically, the shore is the product of massive quantities of fine-grained sediments discharged by the Amazon River that proceed to drift westward. Some small quantities of sand are contributed by the Amazon and by the smaller streams leading from the Guiana Shield to the shore.

The beach and inland coastal geomorphology of the Guianas is characterized by the active development of cheniers. These ridged features are coarse-grained deposits of sand and shell accumulating on a mud or

clay foundation. Intermittent development of these coarser accumulations creates a series of sandy ridges bordered on either side by finegrained sediments. At times the shore is a broad mudflat rather than a sandy beach. Several investigators (Zonnenfeld 1954; Vann 1959; Wells and Coleman 1978) have described the massive clay waves that migrate along this coast and extend far out into the ocean. Turenne (1985) has described westerly migration of these waves or mudbanks as being at a rate of 1.3 km per year.

There is a type of pattern to the chenier or sand ridge development along this coast. A kind of apex forms near the west bank of the river mouth from which a series of ridges extends. The number and extent of the ridges decrease westward to the point where the coarser sediments are no longer found. Some of the river mouths, such as those of the Marowijne and the Suriname, show evidence of west bank progradation for several kilometers by means of mudflat development interspersed with sandy ridges. The ridge trends also provide evidence that the progradation sequence has not been continuous, because numerous ridges are truncated and new fulcrums have developed from which a fan-shaped series of ridges has spread.

Much of the mudflat area is colonized by mangrove. These trees line the shores in places as well as occupy the troughs between the ridges. Wave action appears to be attenuated by the extremely turbid waters, and thus the coastal clays are only infrequently disturbed. However, during the infrequent higher wave energies, there is considerable movement of the clay waves (see above). Large units of clay are displaced, and the sand and shell are sorted to accumulate as a beach on top of the clay base. These beaches continue to develop to form the cheniers.

Coastal lengths of these sectors are, respectively, French Guiana, 370 km (Turenne 1985); Surinam, 350 km (Psuty 1985); Guyana, 434 km (Schwartz 1985).

Venezuela

Venezuela derives its name from late fifteenth century explorers finding lake dwellings on piles in mangrove swamps, and thus calling the area "Small Venice" (Ellenberg 1985). Coastal Venezuela, 3000 km in length, tends to be dominated by the northern terminus of the Andean mountain system. The distal portion of the Andes splits into two north-trending prongs and between them creates the depression occupied by Lake Maracaibo and the Gulf of Venezuela region. The eastern prong makes a sharp bend due east and establishes the northern margin of the country for most of its length. Modifications of the mountainous coastal topography are caused by deltaic development and by breaks in the Andean ridges.

The eastern coast of Venezuela is completely the product of the Orinoco River and its deltaic forms. A fairly large arcuate delta occupies the position from the Gulf of Paria south to the border of Guyana (Van Andel 1967). The southern third of the delta tends to be a series of coalescing distributaries that retain their fluvial forms as they discharge into the Atlantic. However, the northern two-thirds of the delta is lined with cheniers forming a well-developed shore. Mangrove forests occur at the frontal lobes of the southern margin of the delta and at the northern margin. However, in the area of chenier development, the mangrove is in the protected troughs between the ridges rather than at the exposed coast. Warne et al. (2002) have traced the evolution of the delta from a late Pleistocene sea-level lowstand to its present stand in the mid-Holocene. They have also reported that the Volcán Dam, built on the Cãno Manamo to prevent flooding, expand agriculture, and enhance navigation, has had a negative effect on farming due to the presence of pyrite in the soil.

The Paria Peninsula is bounded on its southern side by sand beaches and fairly broad mudflats. The sand beaches are found in association with small streams that drain the high peninsula and contribute their coarser sediments. The eastern half of the south side of the peninsula is a steep rocky cliffed coast, as is the entire northern portion of this Andean extension. There are some pocket beaches, but not until the Gulf of Barcelona is a well-developed beach found. The shore consists of a broad curvilinear beach that has characteristics of barrier island development because several lagoons are formed behind it.

Westward beyond the Barcelona embayment, a cliffed coast appears once again with a number of pocket beaches. A well-developed sandy beach is located at the Triste Gulf; and there is a low, narrow isthmus connecting the Paraguaná Peninsula with the mainland, the result of recent uplift (Ellenberg 1985). Along the Gulf of Venezuela most of the shore is sheltered from marine processes, and thus the features are of fluvial origin and the shores are marshy or lined with mangrove. However, that portion of the gulf exposed to swell waves from the northeast does have prominent coastal forms. The exposure to the northeast is also the dominant fetch direction for the tombolo to Paraguaná and the Triste Gulf beaches as well.

Tanner (1970) has shown that the western shore of the Gulf of Venezuela has a barrier island formation with a series of beach ridges prograding seaward over a distance of 7 km. The ridges are (low, (only about 1 m local relief) and there have been several interruptions in the accumulation sequence. These erosional breaks in beach ridge development are marked by longitudinal and parabolic dunes at the erosional shoreline whose form extends inland over older beach ridges. A number of beach ridge sectors along the Venezuelan coast are backed by sabkha salt lagoons (Ellenberg 1985).

Cross-References

- [Antarctica, Coastal Ecology and Geomorphology](#)
- [Cheniers](#)
- [Cliffed Coasts](#)
- [Coral Reef Coasts](#)
- [Deltas](#)
- [Estuaries](#)
- [Mangroves, Ecology](#)
- [Mangroves, Geomorphology](#)
- [Middle America, Coastal Ecology and Geomorphology](#)
- [Muddy Coasts](#)
- [Sandy Coasts](#)
- [South America, Coastal Ecology](#)

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Spits

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General Features of Sand Spits

Longshore currents lose their sediment transporting capacity at the down-current side of the break in coastline orientation, where the wave energy abruptly drops off, causing the

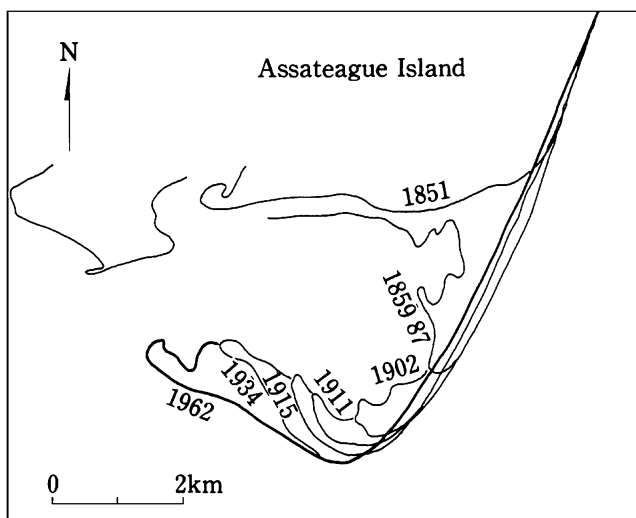
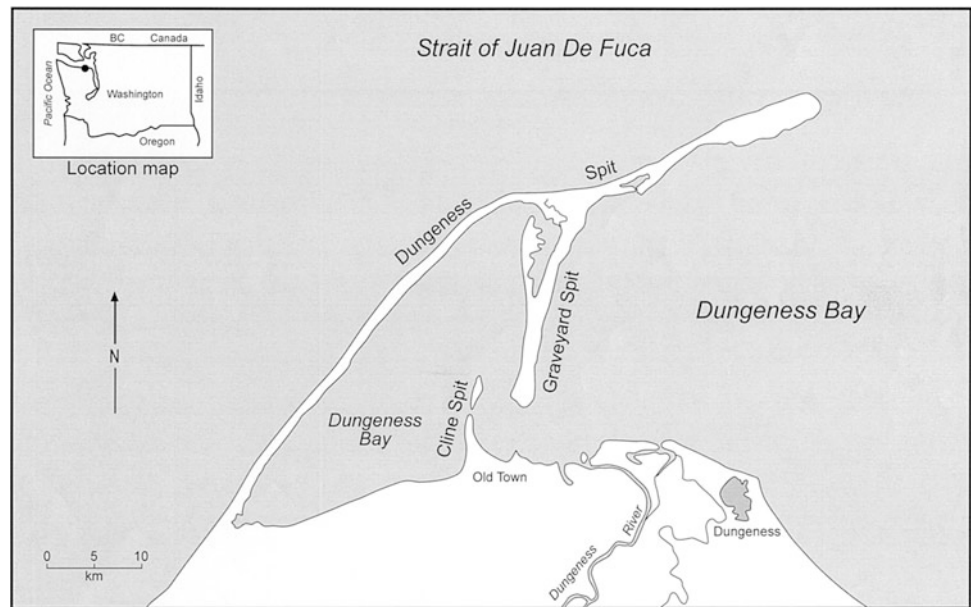
currents to weaken (Horikawa 1988). Sediment accumulates at this location, forming elongate spits, which grow in the direction of the predominant longshore drift. The spits are classified according to their plane shape configuration as simple spits (with relatively linear features), recurved spits (with a distal end hooked landwards), and complex spits (with plural hooks). Figure 1 shows an example of the actual plane shape of a complex spit. Material supplied from the upcoast is transported alongshore to the distal portion, resulting in the extension of the spit. The configuration of the spit terminus depends mainly on (1) the longshore sediment transport rate; (2) wave diffraction and refraction processes; and (3) the interaction between dominant waves and waves arriving from different directions (Zenkovich 1967). The episodic growth of hooks characterizes the development of a complex spit, as illustrated in Fig. 2. A double spit, which is a set of spits extending toward each other from two adjacent locations, frequently develops on a coast where the sediment is transported alongshore in one direction at one time interval of long duration, and in the opposite direction at another time interval.

Longshore growth of spits results in the formation of barriers, with lagoons at their landward side. In addition to longshore spit elongation, barriers can be formed in close association with the postglacial sea-level rise: (1) a straight offshore bar emerged with sea-level rise and migrated landward to form a barrier; or (2) a preexisting subaerial sand mound in the vicinity of the shore moved landward keeping pace with sea-level rise to develop a barrier. Separation of a barrier by estuaries or tidal inlets leads to the formation of a barrier island. The origin of an actual barrier island is complex reflecting local wave climate, sediment supply, and relative rise of sea level (Schwartz 1973).

Experiment of Formation of Sand Spit and Modeling

Sand spits can be reproduced by the movable bed experiment (Uda and Yamamoto 1991). The initial contours of semicircular shape are made by using sand of median diameter of 0.28 mm in a wave tank, and the bed slope is one-fifth as shown in Fig. 3. The initial shoreline is given by a semicircle of $r = 1.5$ m. Given incident wave height of 3 cm and wave period of 0.8 s, subsequent beach changes due to waves were measured. Figure 4 illustrates the shoreline changes under the 2.5-h wave action. The shoreline upcoast of No. 10 retreats, while it progrades downcoast of No. 10. The retreat rate of the shoreline in the eroded zone is faster in earlier stages, and its rate decreases with time. Sand transported by longshore drift accumulates near No. 13 and No. 14 located in the wave

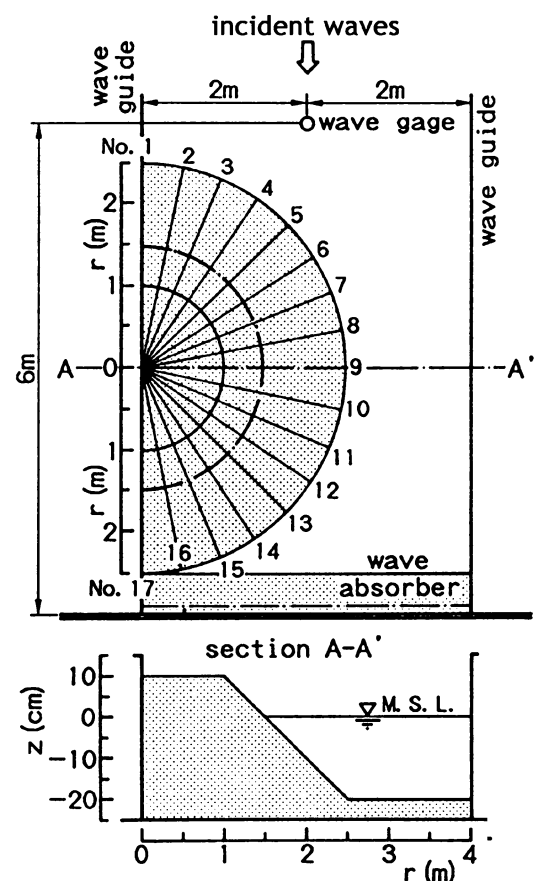
Spits, Fig. 1 Dungeness Spit, a complex spit on the west coast of the USA. (From Schwartz et al. 1987, with permission of the *Journal of Coastal Research*)



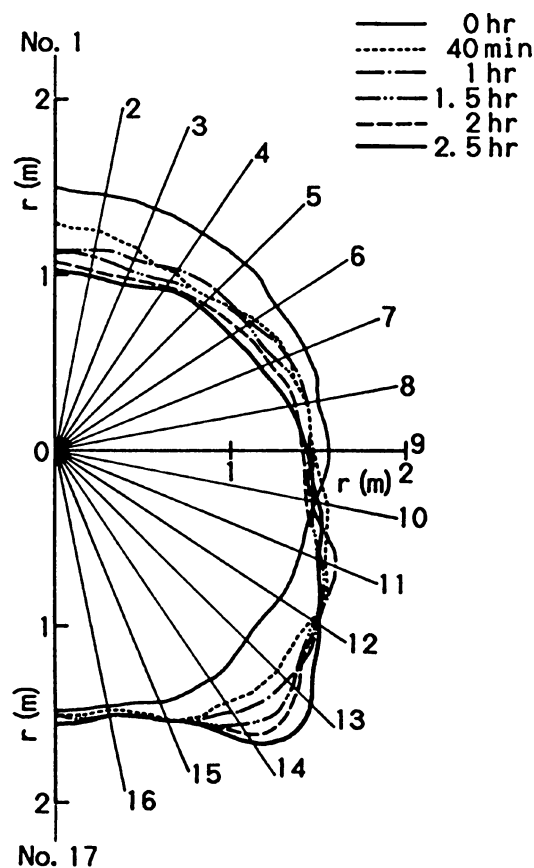
Spits, Fig. 2 Growth of complex spit, located at southern Assateague Island, Virginia, USA. (After Finkelstein 1983)

shadow zone to form a sand spit. The growth rate of the spit is also faster in earlier stages, and it gradually decreases with time.

Figure 5 shows the bottom contours measured after the wave action of 3 h. The interval of the contour lines between 10 and 0 cm on land is very narrow at No. 1 through No. 7, whereas the interval between -2 and -8 cm is wide. This means that the scarp was formed on the fore-shore, whereas a gentle slope is formed on the sea bottom due to the erosion. At No. 11 through No. 15, the interval of the



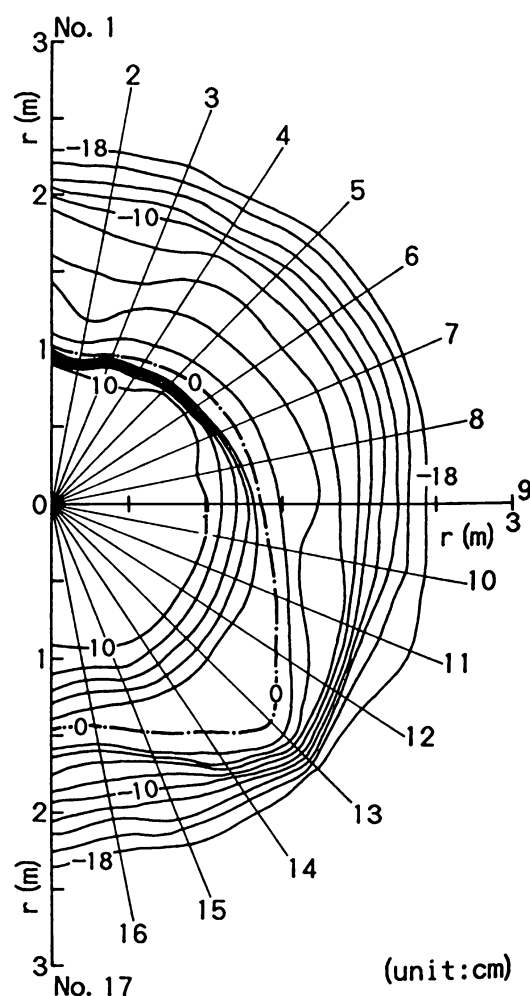
Spits, Fig. 3 Model beach and alignment of measuring lines. (From Uda and Yamamoto 1991, with permission of the American Society of Civil Engineers)



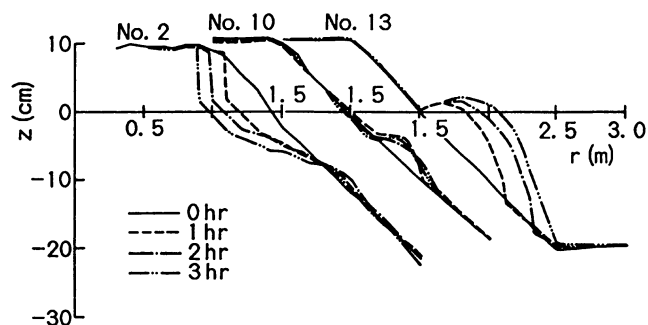
Spits, Fig. 4 Change in shoreline configuration. (From Uda and Yamamoto 1991, with permission of the American Society of Civil Engineers)

contour above 2 cm is similar to those at the initial state, but the contour intervals between 2 and 0 cm are widened and those between -2 cm and -16 cm are narrowed, indicating the formation of a flat plane on the foreshore and a steep slope below the sea surface.

The typical beach profiles representing the erosion and accretion zones, and the one in a neutral zone without erosion and accumulation of sand, are compared as shown in Fig. 6. Along No. 2, located in the eroded zone, the shoreline significantly retreats with the formation of a scarp. The foreshore slope is constant, and the scarp is formed above the foreshore. During the erosion process, the slope of this scarp and the fore-shore slope remain constant, implying that the steep slope on land deforms as if it moves in parallel. Furthermore, the closure depth is 8 cm and a gentle slope is formed between the closure depth and the depth of 2.5 cm. Along No.13 in the accretion zone, most topographic change is observed below the sea surface. Sand accumulates within the wave run-up zone forming a flat beach surface. During the progradation of the shoreline, the foreshore slope remains constant. The bottom slope remains almost constant



Spits, Fig. 5 Beach topography after the formation of sand spit. (From Uda and Yamamoto 1991, with permission of the American Society of Civil Engineers)



Spits, Fig. 6 Beach profiles in eroded, accreted and neutral zones. (From Uda and Yamamoto 1991, with permission of the American Society of Civil Engineers)

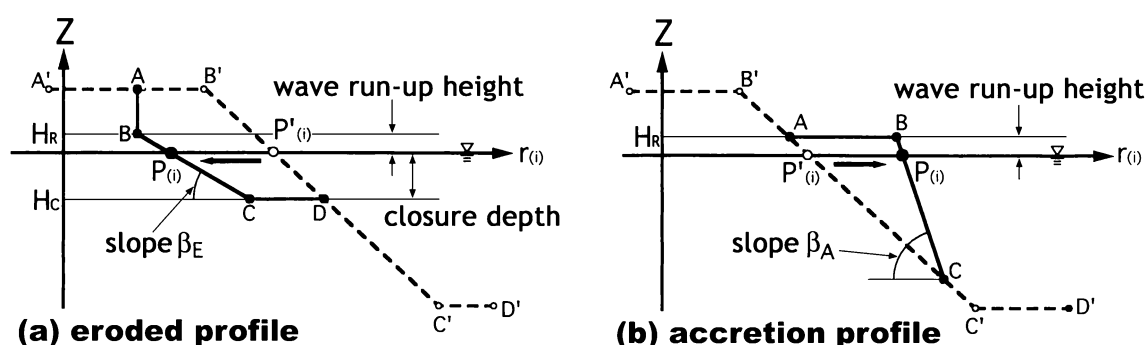
in the zone deeper than -2 cm, and the steep slope deforms as if it moves in parallel. In the eroded zone, no topographic deformation takes place below the critical depth for sand movement.

In summary, the initial topography does not change in the upper part of the scarp. In the accretion zone, the initial topography does not change at levels higher than the maximum wave run-up height. Similarly there is no topographic change in the zone deeper than the critical depth where sediment is deposited. These profile changes can be modeled, as schematically shown in Fig. 7.

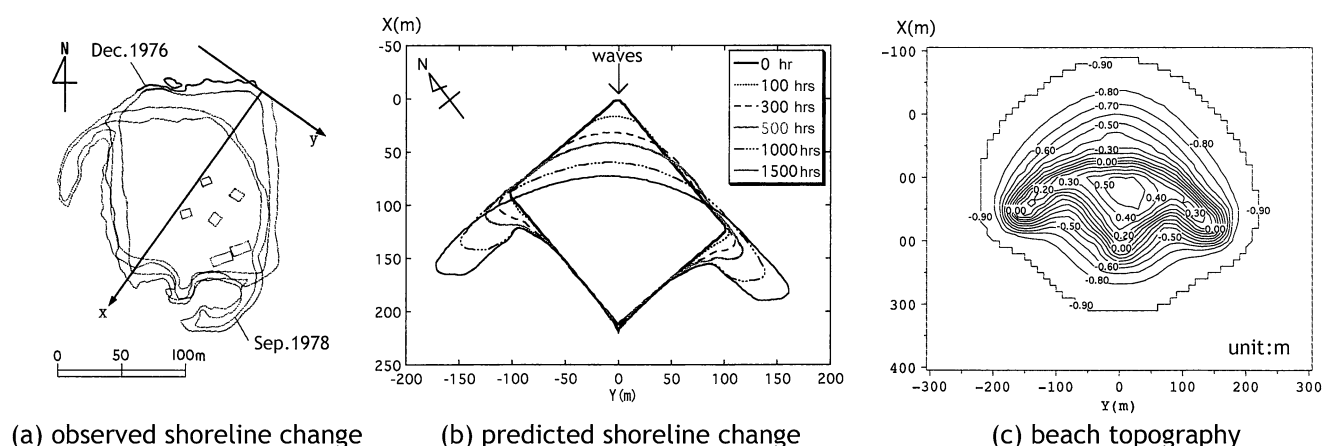
Numerical Simulation of Formation of Sand Spit

Ashuton et al. (2001) show that coastline instability takes place when the incident angle of the deepwater waves relative to the shoreline exceeds 45° , and a sand spit can be automatically formed from an infinitesimal disturbance of the shoreline. In their method, the shoreline is divided into the infinitesimal meshes in the x - and y -axes, and unidirectional sediment movement forming a sand spit is reproduced by using this coordinate system. However, longshore sand transport is not provided by the deterministic method from the wave field. Watanabe et al. (2002) show that successive development of a sand spit can be numerically predicted

using a one-line model with orthogonal curvilinear coordinate system fixed on the ever-changing shoreline. In their model, the breaker height and breaker angle are determined first on this coordinate system and longshore sand transport rate is obtained from these values, resulting in the shoreline calculation based on the continuity of sand. A new sea bottom shape is inversely assessed from the new shoreline configuration at some steps, assuming the profile changes in eroded and accreted areas, as shown in Fig. 7, and a new wave field is calculated using this sea bottom data. Furthermore, smoothing of the shoreline configuration and redistribution of grid points are carried out to ensure the stability of the calculation. Figure 8a is the sand spit formation around an artificial sandy island measured in Alaska (Gadd and Czerniak (1979) and Gadd et al. (1979)). A rectangular island of 137 m length and 99 m width was created in the very shallow sea of the depth of 0.91 m. Comparison of the shoreline is made between December 1976 and September 1978. Sand was transported from the northeast corner of the island due to the predominant waves from the northeast, and sand deposited to form two sand spits in the wave shadow zone at the southeast and northwest corners.



Spits, Fig. 7 Model of beach profile change (Watanabe et al. 2002)



Spits, Fig. 8 Change in shoreline configuration of an artificial island. (From Gadd and Czerniak 1979; Gadd et al. 1979, with permission of Tetra Tech Inc.)

Figure 8b shows the calculation results assuming the incident wave height of 0.3 m and the wave period of 3 s. The wave field is predicted based on the wave orthogonal theory assuming irregular waves. In the calculation of the shoreline change using a finite difference method, time step of $\Delta t = 5.0 \times 10^{-2}$ h is selected and the shoreline change was calculated up to 30,000 steps. Renewal of the depth data and resulting wave field was carried out at 300 steps (15 h). Furthermore, the closure depth, wave run-up height, $\tan \beta_E$ and $\tan \beta_A$ as shown in Fig. 7 are set to be 40 cm, 20 cm, 0.1 and 0.3, respectively. The initial shoreline of a rectangular shape deformed with the successive recession of the shoreline at the northeast corner of the island. Eroded sand was transported by longshore drift forming two sand spits at the northwest and southeast corner of the island. Observed results are well reproduced by the numerical simulation. Figure 8c shows the seabed topography after change. A mild seabed was formed in the eroded zone, whereas in the accretion area, a flat foreshore was formed due to the berm formation as well as the steep slope due to the falling of sand into deep sea. Thus, in the prediction of the development of the sand spit, coupling of changes in waves and sea bottom is important.

Cross-References

- ▶ [Accretion and Erosion Waves on Beaches](#)
- ▶ [Beach Features](#)
- ▶ [Beach Processes](#)
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- ▶ [Modeling Platforms, Terraces, and Coastal Evolution](#)
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- ▶ [Wave Refraction Diagrams](#)

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Storm Surge

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Definition

A storm surge is the increase in ocean water level near the coast generated by a passing storm, above that resulting from astronomical tides.

The atmosphere acts on the sea in two distinctly different ways. A drop in atmospheric pressure reduces the vertical force acting on a column of water beneath the sea surface, causing the sea water to rise and vice versa. A decrease in atmospheric pressure of 1 mbar will produce an increase in sea level of around 1 cm. This change is called the *inverse barometer* effect. The inverse barometer effect is seldom exactly observed in nature, because of the complex ways in which shallower waters of the continental shelves interact with passing atmospheric pressure systems.

Another major meteorological process contributing to the surge is the drag or stress on the sea surface due to the wind, measured as the horizontal force per unit area. Wind stress depends upon the wind speed and air density. The wind strength and direction relative to the coastline are important factors in elevating the sea surface. Wind blowing toward the coast causes a much greater rise in sea surface than wind blowing parallel to the shore. The effect of the wind on sea level increases inversely with water depth. Thus, the surge is amplified when wind blows over wide regions of shallow water, such as the North Sea or the northern end of the Bay of Bengal, off the coast of Bangladesh. The superposition of wave run-up on a high surge increases the storm hazard.

Two distinct types of meteorological disturbances produce major surges leading to coastal flooding and beach erosion (Dolan and Davis 1994). These include tropical and extra-tropical cyclones. Tropical cyclones (known as *hurricanes* in the Atlantic basin, *typhoons* in the western Pacific, and *cyclones* in the Indian subcontinent) are intense low-pressure systems that strengthen over the ocean at low latitudes, where sea surface temperatures are at least 27°C (Zehnder 2015). Tropical cyclones are usually small in extent but very powerful. The storm surge produced by a hurricane is a dome of water raised by the low barometric pressure, coupled with the strong wind shear (maximum sustained wind speeds of at least 119 km/h; surges 1–2 m), particularly on the right side of the low-pressure system. The hurricane's strength is greatest in the upper right-hand quadrant, as the high-velocity

winds flowing counterclockwise around the eye (in the Northern Hemisphere) reinforce the lower-velocity steering winds which direct the forward motion of the cyclone (Zehnder 2015; Coch 2015). No changes in tropical cyclone frequency or intensity have been detected since the 1970s, although the intensity of the strongest North Atlantic cyclones has increased (Elsner et al. 2008) and will likely continue to strengthen in the future (Peduzzi et al. 2012; Knutson et al. 2010). The most intense tropical cyclones have also shifted poleward within the past three decades (Kossin et al. 2014). Increasing coastal populations, economic development, and greater cyclone intensity will magnify future flooding risks and damages (Estrada et al. 2015; Hallegatte et al. 2013; Peduzzi et al. 2012; Mendelsohn et al. 2012).

Extratropical cyclones typically have lower wind speeds than hurricanes. Yet they often inflict considerable damage along the coasts because they cover a much wider area (>1,000 km for extratropical cyclones versus 100–150 km for hurricanes) and are longer in duration, often persisting over several tidal cycles at a particular location (Colle et al. 2015; Dolan and Davis 1994). South of Cape Hatteras, along the East Coast of the United States, tropical storms occur more frequently than extratropical storms, whereas further north, extratropical cyclones dominate (Zhang et al. 2000).

In addition to weather systems, surges are influenced by the configuration of the coastline and bathymetry. Surges are amplified by a wide continental shelf and where the coastline makes a right-angle bend, as, for example, at the apex of the New York Bight (Coch 2015) or along the coast of Bangladesh (Murty and Flather 1994). Interaction between surges and tides due to local topography may considerably change surge amplitude and phase.

The surge is amplified significantly when it coincides with astronomical tides (Wood 1986). Above-average tides occur at new or full moon (spring tides) and at the solstices and equinoxes. These fortnightly or seasonal high tides are reinforced when they coincide with perigee (the closest approach of the moon to the earth in its orbit) or perihelion (the closest approach of the earth to sun). Tides, especially at high latitudes, are further enhanced during the 18.6-year lunar nodal cycle, when the angle between the plane of the moon's orbit and the ecliptic is minimum (see ► [Tides](#)).

Many major extratropical storms have struck the eastern North American coastline at times close to perigean spring tides (Wood 1986). One of the worst in almost 100 years was the “Ash Wednesday Storm” of March 6–7, 1962, which affected the entire Atlantic coast from South Carolina to Portland, Maine. This storm owed much of its destructiveness to its duration over five tidal cycles. Another major perigean spring storm was the “Saxby Tide” or “Saxby Gale” of October 4–5, 1869, which caused enormous damage along the Bay of Fundy, Canada (Desplanque and Mossman 1999). Tides

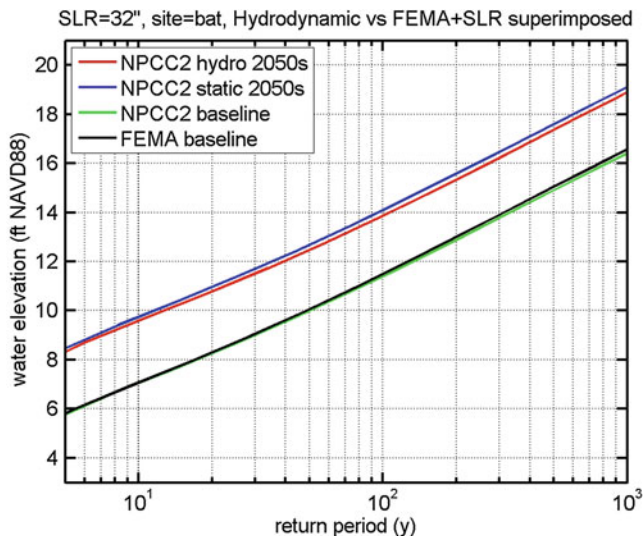
were nearly 2 m above the previous record. More recently, Superstorm Sandy (a hybrid extratropical/tropical storm in NYC), on October 29, 2012, generated the highest recorded water level (3.38 m) at the Battery (the southern tip of Manhattan) in nearly 300 years (Orton et al. 2016; Talke et al. 2014), due to strong easterly winds, surge amplification, historic sea-level rise (0.44 m since 1856), and maximum storm surge coinciding with high tide and full moon. The storm caused 43 fatalities, hospital and nursing home evacuations, inundation of subways and tunnels and 17% of the city's land area, power outages affecting nearly two million people, and major transportation disruptions. Over 70,000 buildings were flooded and/or damaged. Storm damages totaled an estimated \$19 billion (SIRR 2013).

Historical Observations

Historical storm activity has been monitored using meteorological data and tidal records. Dolan and Davis (1994) have calculated wave heights from wind field data for each important storm along the East Coast, United States, between 1942 and 1992. They developed an extratropical storm classification scheme, similar to the Saffir-Simpson scale used for hurricanes, based on significant wave heights, duration, and storm “power,” which they define as the square of significant wave height times duration. While this approach provides an index of storm damage potential, it does not specifically measure surge levels associated with these storms.

Another approach is based on measurement of hourly water levels recorded by a tide gauge. Since the water height includes surge, astronomical tide, and long-term sea-level trends, the latter two components must be subtracted. Twentieth-century patterns of storm activity along the East Coast, United States, have been reconstructed from hourly tide gauge data with records longer than 50 years, after removal of astronomical tide and sea-level factors (Zhang et al. 2000). Storm indices based on surge number, duration, and “integrated intensity” (area under surge curve greater than 2 standard deviations) showed considerable interdecadal variations, but no consistent long-term trend during the twentieth century. A review of past US East Coast extratropical cyclone activity (Colle et al. 2015) and storm surges (Booth et al. 2016) reached similar conclusions.

The likelihood of future storm surge events can be calculated from daily or monthly surge maxima measured at individual tide gauge stations, after removal of the effects of tides and secular sea-level trends (e.g., Booth et al. 2016; Tebaldi et al. 2012). Statistical analyses of such data can then be used to prepare return period curves which show the probability of occurrence of a given surge (or flood) height (Fig. 1). The total flood level (surge + tide + sea-level rise) is a more meaningful elevation for assessing impacts to coastal



Storm Surge, Fig. 1 Flood return curve for the Battery, New York City (after Orton et al. 2015; Horton et al. 2015). The present 100-year flood level is 11.3 ft (3.4 m) (*lower black line*). For a projected sea-level rise of 2.5 ft (0.76 m) by the 2050s, the 100-year flood height increases to 13.8 ft (4.21 m), with a reduction in the return period to 28 years (*upper curves*)

settlements and installations. The 100-year flood level is a commonly employed reference level for coastal flood risk analyses and insurance purposes (FEMA 2011). It represents the flood height that has a 1% probability of occurrence in any given year.

Modeling Surges

The empirical data needed to derive return period curves are not available at all locations. Therefore, the storm-induced surge is often calculated using mathematical models. The US National Weather Service has developed a numerical dynamic surge model, SLOSH, for real-time forecasting of hurricane surges on continental shelves, open coastlines, and estuaries (Jelesnianski et al. 1992). The model is activated by inputs of simple meteorological parameters (minimum central pressure, radius, storm track, and speed along track). The storm model balances surface forces including surface friction. Short period wave “run-up” is omitted; other longer-term wind-wave effects are generalized. SLOSH surge predictions are regularly applied to hurricanes affecting the US East and Gulf Coasts. Updated versions of SLOSH are still used for forecasting worst-case scenarios of hurricane storm surges and to issue recommendations for evacuation, if necessary.

A more sophisticated model, the ADCIRC (Advanced CIRCulation) hydrodynamic model, is used by the US Army Corps of Engineers (Luettich et al. 1992). ADCIRC solves three-dimensional, dynamic, shallow-water wave equations of fluid motion, incorporating information on

bathymetry, topography, wave, and meteorological data in order to simulate coastal flooding. It is frequently coupled with SWAN in order to simulate wave/hydrodynamic interactions, including wave setup (Booij et al. 1996).

Another type of three-dimensional hydrodynamic model simulates the flooding associated with both tropical and extratropical cyclones. For North Atlantic tropical cyclones, a suite of synthetic storms was developed from a statistical analysis of climatological data. The extratropical cyclone assessment utilized historical storm data together with multiple random tide-surge phases (Orton et al. 2016). While the combined storm set data were initially used to construct return period curves for New York harbor, the methodology can be adapted to other coastal cities, as well.

Effect of Sea-Level Rise on Coastal Flooding

Anticipated climate change is expected to accelerate rates of global mean sea-level rise by factors of 2–5, by the end of this century (IPCC 2013). The rise in sea level will come from thermal expansion of the ocean, melting of mountain glaciers, with a larger uncertainty surrounding possible contributions from increased melting of the polar ice sheets.

The rise in mean sea level will be superimposed on storm surges from storms and astronomical tides. Locally, sea-level rise could be higher than the global mean change, due to land subsidence, changes in ocean circulation, and other factors. Although the increase in flood height due to sea-level rise may only comprise a small proportion of the total elevation (assuming that the storm strikes at high tide), even a minor increase in sea level could significantly shorten the flood return period (Fig. 1). For example, in New York City, the present-day 100-year flood height is around 11.3 ft (3.4 m). For a projected sea-level rise of 30 in. (0.76 m) at the 90th percentile by the 2050s (over a base period of 2000–2004), the 100-year flood height increases to 13.8 ft (4.21 m), with a corresponding reduction in the return period to 1 in 28 years (Horton et al. 2015). This would lead to much more frequent episodes of coastal flooding, with associated damages in low-lying areas (see also ► [Natural Hazards](#)).

Coastal Surge Hazards

Areas of the world most vulnerable to surges from tropical cyclones include the low-lying coasts of Southeast Asia adjacent to major river deltas, the Southeastern United States, and the northeastern coast of Australia. The convergence of several conditions places Bangladesh especially at risk to storm surges. The coast is nearly at sea level, the continental shelf is

shallow especially in the eastern part of Bangladesh, the tidal range is high, and storm tracks have a tendency to recurve near the apex of the Bay of Bengal (Murty and Flather 1994). A severe cyclone in November, 1970, produced surges of over 9 m, killing several hundred thousand people (Pugh 1987). Another cyclone killed 11,000 people in May, 1985. Since then, Bangladesh has instituted a coastal storm warning system and evacuation centers, which have reduced the death toll in more recent storms. But in neighboring Myanmar, Cyclone Nargis killed over 138,000 people on May 2, 2008. Surge levels reached up to 5 m, not including 2-m high waves. The storm was tracked by weather satellites, so that deaths could have been avoided by issuing timely evacuation warnings (Fritz et al. 2009).

In the United States, the low-lying Gulf states, Florida, and the Carolinas are at high risk to hurricanes, although the Northeast is not immune. Hurricane Katrina flooded 80% of the city of New Orleans on August 29, 2005, killed ~1,200 people, and caused \$106 billion (2010 dollars) in damages (Blake et al. 2011). A hurricane that struck Galveston, Texas, on September 8–9, 1900, raised water around 5 m and killed around 8,000 people (Blake et al. 2011). In September, 1938, another hurricane swept across Long Island and Southern New England, pushing water levels 4.5–6 m above normal, and killed around 700 people (Ludlum 1988). Other major damaging hurricanes include Hugo (South Carolina, 1989, category 4 on the Saffir-Simpson scale), Andrew (Florida, Louisiana, 1992, cat. 4), Camille (Mississippi, Louisiana, 1969, category 5), and an unnamed hurricane that struck the Florida Keys in 1935 (category 5).

Extratropical cyclones can also result in severe coastal flooding. Regions prone to surges from extratropical cyclones include the east coast of North America north of Cape Hatteras, the coasts surrounding the North Sea, and the Adriatic Sea. The Ash Wednesday Storm of March 1962, affecting much of the US East Coast, was mentioned above. Other major “nor’easters” include the December 11–12, 1992, storm; the March 13–14, 1993, storm; and the “Halloween Storm” of October 31 to November 1, 1991 (also known as “The Perfect Storm” – subject of a best-selling novel and film).

In Northern Europe, the North Sea is exposed to Atlantic extratropical storms which move across shallow shelf waters. Furthermore, the gradual subsidence of the southern North Sea area makes low-lying regions of South East England and the Netherlands increasingly vulnerable to flooding from surges. In January 31 to February 1, 1953, a major storm broke through the dike defenses of the Netherlands, flooding a large portion of the country (Pugh 1987). The disaster led to the construction of the Delta Works – a massive engineering project designed to protect much of the coast from future coastal storm surges.

Summary

Storm surges are the most common and costly type of coastal hazard. The deleterious consequences of storm surges are amplified by geomorphological factors including bathymetry, width of continental shelf, and coastal configuration. Meteorological factors include maximum wind speed, storm track, and timing with respect to astronomical tides. Storm surge damages have grown in coastal areas due to increased development. Storm surges of a given flood height will recur with greater frequency as sea level rises, even without additional changes in storm characteristics.

Cross-References

- [Changing Sea Levels](#)
- [Meteorological Effects on Coasts](#)
- [Natural Hazards](#)
- [Sea-Level and Climate Change](#)
- [Tide Gauges](#)
- [Tides](#)
- [Waves](#)

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Strandflat

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The strandflat is a rim of gently sloping bedrock plain in front of higher land or coastal mountains (Klemsdal 1982, 1985). The plain has a very irregular terrain with small differences of height. Most of the bedrock plain is covered with a thin mantle of loose material; only locally does the loose material have forms of its own. The gently sloping, undulating, flat produces, when meeting the sea, an uneven coastline with numerous bays, coves, inlets, islands, islets and skerries, all characteristic features of the skerry zone. The gradient of the supramarine zone varies between 5 and 25 m per km. The topography of the supramarine and skerry zones continues into the sea. At a depth between 30 and 60 m a break in the slope occurs and steeper slopes lead down to the paleic, old forms of the bankflats constituting part of the continental shelf. Inland, the strandflat terminates at heights between 30 and 80 m, where in most places a steep slope leads up to higher land, though in some places the transition is more gentle. The coastline of the mainland and the islands of the strandflat is rather long. The shore-zone is mainly either a gently sloping ice-smoothed rocky shore, without any post-glacial alteration, or a stony beach composed of reworked till. Only locally have post-glacial littoral processes produced pocket beaches of sand or clay (Klemsdal 1982, 1985).

The term strandflat became a geomorphological term in 1894 when Reusch introduced it into the literature. Reusch put forth, from observation of strandflat localities along the Norwegian coast, a description of the strandflat and a discussion of its origin. Marine abrasion and frost weathering were said to be the main processes responsible for the development of the strandflat. Regarding the development of the strandflat, Nansen (1922) proposed frost weathering and sea ice frozen onto the shore, combined with waves breaking up the sea ice. Local glaciation, especially cirque glaciers, and also currents of inland ice, moving and spreading out in the coastal areas, were added to the picture by Holtedahl in 1929.

Today, the origin of the strandflat is closely associated with the general development of the landforms of the Scandinavian landmass. The land was exposed to denudation through the Mesozoic and the first part of the Tertiary. In a warm climate with dry and wet periods, denudation produced a land surface, the paleic surface (Gjessing 1967), with well-rounded, mature landforms, close to a plain in the peripheral parts of the landmass, a propitious starting point for the development of the strandflat.

In the Tertiary the Scandinavian landmass was elevated and tilted, giving a steeper slope towards the west and north-west, bringing the paleic surface to different heights above

sea level. Inland along the coast it varies from sea level to 500–700 m above sea level. Along the coast the elevation of the paleic surface may have been negligible.

Probably coincident with the uplift, the climate became temperate and more humid. Fluvial processes produced initial forms along a fluvial pattern in the paleic surface, most distinct in the west and northwest. Along the coast marine abrasion and denudation, favored by the old paleic surface, may have started the development of a peneplain, which can be easily fitted into the development of the strandflat.

Cirque glaciers, descending from higher coastal mountains onto the level surface along the coast, were important in the widening and splitting up of the strandflat. Valley glaciers and ice flows of the inland ice spreading out in the coastal areas (Holtedahl 1929; Dahl 1947), may also have taken part in the development. In interglacial times and at the beginning of the glacial periods frost weathering and marine abrasion were momentous, and together with mass movement and littoral transport, most active in the evolution of the strandflat.

Although its origin is still not settled, it is said that it is polygenetic (Holtedahl 1959, 1960; Klemsdal 1982, 1985); starting with the denudational processes of the paleic surface and stressing the frost weathering and marine abrasion, even though glacial and fluvial activity also have played a part in the development of the strandflat.

The distribution of the strandflat is closely connected with areas where Quaternary glaciations took place and where a harsh climate in Holocene favored frost weathering.

In Norway, the strandflat is found along the coast from the Stavanger area in the southwest to the western Finnmark in the northeast (Klemsdal 1982, 1985). In eastern Finnmark, the strandflat is only a narrow strip of land, some broader but still narrow submarine parts, and no skerry zone. The maximum width is approximately 40 km, but as a mean the strandflat is 16 km wide. The supramarine part of the strandflat is normally between 5 and 10 km in width, but can reach 15 km. In other places, the strandflat is a few hundreds meters wide.

Furthermore, the strandflat is found in the archipelago of Svalbard, on Novaya Zemlya; on the peninsula of Taimyr; the archipelago on the west coast of Scotland; on Iceland; along the southwestern coast of Greenland; and in the arctic areas of Canada and Alaska. In the Southern Hemisphere, the strandflat is observed along parts of the Antarctic Peninsula and neighboring islands. The mentioned localities of the strandflat are all in a harsh climate where frost weathering takes place and where Quaternary glaciations occurred or influenced the processes.

Cross-References

- [Cliffs, Erosion Rates](#)
- [Glaciated Coasts](#)

- [Ice-Bordered Coasts](#)
- [Paraglacial Coasts](#)
- [Rock Coast Processes](#)
- [Shore Platforms](#)
- [Weathering in the Coastal Zone](#)

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Submarine Groundwater Discharge

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Background and Definitions of Terms

Groundwater comprises about 95% of all the freshwater on earth (Winter et al. 1998). Of this groundwater, an unknown amount discharges to lakes, rivers, estuaries, bays, and oceans on an annual basis. Groundwater discharged to the marine environment (estuaries, bays, and the ocean) is known as submarine groundwater discharge (SGD), a term originally proposed by Johannes (1980). SGD occurs offshore as seepage (groundwater discharges from both unconfined and confined aquifers) (Simmons 1989; Simmons and Reay 1992), as recirculated seawater mixed with freshwater (Buddemeier 1996), and in springs (Kohout 1966). Additionally, SGD can occur in geo-thermal heat flows or vents (Kohout et al. 1979; Schwerdtfeger 1981; Nossin et al. 1987), upwelling in reefs (Simmons and Love 1987; Tribble et al. 1992) and oceanic islands (Rougerie and Wauthy 1993).

The chemical makeup of SGD varies spatially and temporally depending upon the composition of the aquifer material, residence time of the freshwater within the aquifer, onshore hydraulic heads, the amount of recirculation of seawater and groundwater in the nearshore, and the magnitude of the tidal fluctuations. The primary chemical characteristics (or *finger print*) of the water are functions of the original geochemical composition of the groundwater and the residence time in the aquifer and minerals contacted along the flowpath.

The geologic principle of *uniformitarianism* states that geologic processes today (including SGD) are similar to the processes of millions of years ago (Wicander and Monroe 1993). Therefore, SGD processes occurring today in beaches, estuaries, and inlets are thought to be the same as past processes, since the formation of the continental plates, changed only by the conditions that drive them. This is generally true for all types of seepage, except for seepage that is influenced by anthropogenic factors such as drainage canals and aquifer pumpage. SGD is not a recently discovered phenomenon. The earliest account of springs in the ocean was presented by Kohout (1966) in which he indicated that Pausanias, who lived in the second century AD, noticed that off the coast of western Italy . . . “there is water boiling in the sea and it has created an artificial island.”

Importance of SGD in the Coastal Environments

The significance of SGD has become more widely appreciated in the last two decades, especially in efforts to improve understanding of marine geologic and geochemical processes, nearshore and offshore water quality anomalies, nutrient loading, and ecological impacts (Kohout 1966; Johannes 1980; Simmons 1992; Zektser and Loaiciga 1993; Buddemeier 1996). Recent research, even by lay scientific journals, has been stressing the importance of SGD (Svitil 1996). Anthropogenic nutrient input from stormwater runoff to the oceans and estuaries was previously believed to have been a major cause of eutrophication of coastal waters (Zektser and Loaiciga 1993; Lapointe and Matzie 1996; US Environmental Protection Agency 1998). Although groundwater was traditionally overlooked or underestimated in previous investigations, it has been demonstrated that groundwater input can be a major source of nutrient input to the oceans. Recent research has identified nutrient enrichment as one of the major processes responsible for the degradation of coastal waters. This enrichment, which causes problems ranging from anoxia to harmful algae blooms, is directly related to the transport of nutrient-enriched groundwater into coastal waters through SGD as demonstrated by Lapointe et al. (1990), Shinn et al. (1994), and Harbor Branch Oceanographic Institute (1995). Groundwater and recharge areas are often contaminated by agricultural, urban fertilizers,

animal waste, shallow injection wells, septic tanks, cesspools, and other nutrient sources. The result leads to increased concentrations of nitrogen (N) and phosphorus (P) in SGD (Simmons and Love 1987; Lapointe and O’Connell 1989; Shinn et al. 1994, 1997; Finkl et al. 1995; Finkl and Krupa 2000, 2003) with disastrous long-term effects.

Governmental agencies and private organizations need SGD information to deal with water degradation issues, develop accurate numerical models, create tidal mixing models, make circulation and temperature predictions, and to establish correct boundary conditions for numerical groundwater and oceanic model simulations (Buddemeier 1996). Accurate determination of SGD is essential because flow volumes are used to calculate mass balances and in the determination of nutrient fluxes to the coastal waters. However, experience has shown that quantifying SGD input is difficult because of uncertainties in groundwater flux measurements.

Occurrence of SGD

The rate and direction of seepage is influenced by hydraulic heads within the aquifer and by the underlying lithology, watershed topography, and seasonal weather patterns (Freeze and Cherry 1979). Seepage is known to occur from the nearshore (Cable et al. 1997a, b) up to about 100 km offshore (Manheim 1967). The process of SGD is more complex than seepage to terrestrial water bodies because of the influence of bathymetry, micro-topology, tidal cycles, winds, currents, recirculation at or near the freshwater/saltwater interface, and the chemical, biochemical, geological, and ecological processes occurring at the freshwater/saltwater interface (Huettel and Gust 1992; Buddemeier 1996; Huettel et al. 1996, 1998; Robinson 1996; Shinn et al. 1997; Uchiyama et al. 2000).

Seepage occurs as baseflow (flow from saturated groundwater storage) and interflow (flow from subsurface storm flow) to the surface water (Davis and DeWiest 1966; Freeze and Cherry 1979). The base-flow component of seepage can be further divided into two primary velocity components. The first baseflow velocity component provides the normal daily discharge to the estuary/ocean and the second velocity component is controlled by daily tidal cycles. Interflow is the temporary saturated flow within the unsaturated portion of the aquifer. This flow generally begins some time after a rainstorm or seasonal climatic change (winter to spring) and occurs until hydraulic gradients within the aquifer have leveled or have dropped to background levels.

Identifying and Measuring SGD

Nearshore submarine groundwater discharge can be seen by visual observations and measured by pore water devices,

seepage meters and results of isotope analysis (Simmons and Love 1987; Harbor Branch Oceanographic Institute 1995; Nuttle and Harvey 1995; Moore 1996; Cable et al. 1997a, b). Offshore SGD can also be identified by side-scan sonar, near bottom echo sounders, seismic profiling, geophysical logging, conductivity, depth, and temperature (CDT) profiling, and deep-sea drilling (Manheim 1967; Zektser and Meskheteli 1988; Land et al. 1995; Merchant et al. 1996; Guglielmi and Prieur 1997).

Measurement of seepage was initially developed for onshore applications, but adaptation of equipment and practices to the marine environment has recently made offshore applications commonplace. Using this relatively new and sophisticated technology, SGD rates and water quality parameters can be determined in the field. A comprehensive SGD study along coastal regions generally requires a variety of methods to best determine seepage rates. For example, monitor wells and piezometers can be used to determine horizontal and vertical gradients while water quality, radiochemistry, and seepage meter data are used to estimate the quantity and direction of seepage.

Coastal Piezometers and Groundwater Wells

To determine the tidal effects from the horizontal and vertical movement of the groundwater/seawater boundary, Urish and Ozbilgin (1989) recommend the placement of one control well onshore, near the surf area. Onshore monitor wells and piezometers in shallow water can be installed using standard well technology, including casing advancement methods and jetting. Methods for installing monitor wells or piezometers underwater are adapted from land-based techniques, but are far more complicated and time-consuming (Lee and Cherry 1978; Shinn et al. 1994). Wells used to measure vertical gradients require short-screen intervals or open-ended pipes (Freeze and Cherry 1979; US Army Corp of Engineers 1993). Screen lengths can be as short as 0.15 m, but are generally 0.6 m. The screens should be sufficiently separated both horizontally and vertically, to calculate differences in hydraulic heads between the monitored groundwater zones and surface water (Harvey et al. 2000, 2002).

Continuous logging of water level data from the piezometers, monitor wells, and surface water should be corrected to equivalent freshwater heads, if they are completed on the same flow path, to allow comparison of data (Senger and Fogg 1990; Reich 1996; Rasmussen 1998). Tidal filtering of water level data should be done to obtain average vertical and horizontal gradients. To calculate the equivalent freshwater head, it is necessary to measure the transient salinities and/or total dissolved solids (TDS) concurrently with the water level measurements. All wells, piezometers, and the ground surface at each site should be surveyed vertically (elevation) and horizontally (longitude and latitude or state planar

coordinates) to ensure accurate gradient calculations and site maps preparation.

Seepage Meters

Since the 1930s, scientists have used a variety of chambers and seepage meters to study the benthic boundary layers associated with canal linings, rivers, and lakes (Gale and Thompson 1975; Sonzogni et al. 1977; Carr and Winter 1980). One commonly used seepage meter is the modified drum apparatus devised by Lee (1977). This meter, commonly called a "Lee meter" or a conventional seepage meter, has been used in numerous marine studies (Lewis 1987; Cable et al. 1997a, b; Shinn et al. 2002). The Lee meter is constructed from a standard metal drum (208 L) cut into thirds. The middle section is discarded and the two end sections are each prepared by using the existing hole on one end and drilling a similar hole on the other end. A connector is fitted to each hole; these are used to attach seepage bags after the meters are placed. The meters are installed into the ground with the open end downward, seepage bags are attached to the connectors and monitored.

Conventional seepage meters operate very simply: a change in water volume in the bag over time represents a seepage flux across the sediment interface. The quantity of water in the bag is compared with the initial volume in order to determine gain or loss of water. When the bag gains water, groundwater is discharging to surface water. Conversely, when the bag loses water, the surface water is recharging the groundwater. The change in quantity of water is averaged, per unit area, and usually converted to units such as milliliters per square meter per hour or centimeters per day.

Seepage can also be measured with automated seepage meters. These devices utilize electronic flowmeters located on the tops of seepage meters or ultrasonic flow meter sensors located on the top of a funnel apparatus (Reay and Walthall 1992; National Aeronautics and Space Administration 1992; Taniguchi and Fukuo 1993; Paulsen et al. 1997; Krupa et al. 1998; O'Rourke et al. 1999). Automated seepage meters can best deal with tidal cycles when continuous recorders are used. This allows the investigator to capture seepage rates over longer periods of time which includes storm events and normal tidal fluctuations, investigation duration (months vs. days) and finally increases the statistical significance of the flux readings. Manual seepage meters capture data only over the time the bags have been installed on the meters.

Seepage meters are typically positioned within the first 100 m of the coast, according to Bokuniewicz and Zeitlin (1980). Recent studies in Panacea, Florida and Perth, Australia, show that seepage typically decreases with distance from shore. However, multiple seepage meters are required at various locations in order to improve statistical significance and to evaluate spatial variability within a designated seepage location. The typical configuration of a Lee meter in the

offshore environment is shown in Fig. 1. Sediment changes, onshore topography, bathymetry, distance from shore, and ecological indicators such as sea grasses, corals, or algal mats should be considered when placing seepage meters. Figure 2 shows a typical nearshore seepage meter scenario with shallow and deep piezometer and conventional seepage meter.

Water Budget: Mass Balance Approach

The mass balance approach to estimation of SGD summarizes all inflows and outflows in a study area. Cherkauer (1998) states “the principle of conservation of the mass requires that the total quantity of groundwater entering the surface water must equal the net recharge within the watershed contributing flow.” This approach assumes the system is in steady-state and there is no net change in the groundwater levels. Thus, the net infiltration (rainfall minus transpiration, evaporation, and anthropogenic uses) to the aquifer is equal to the SGD. Several investigators have completed large-scale regional water budgets (Buddemeier 1996; Church 1996), while others have attempted continental water budgets for Australia (Zektser et al. 1983).

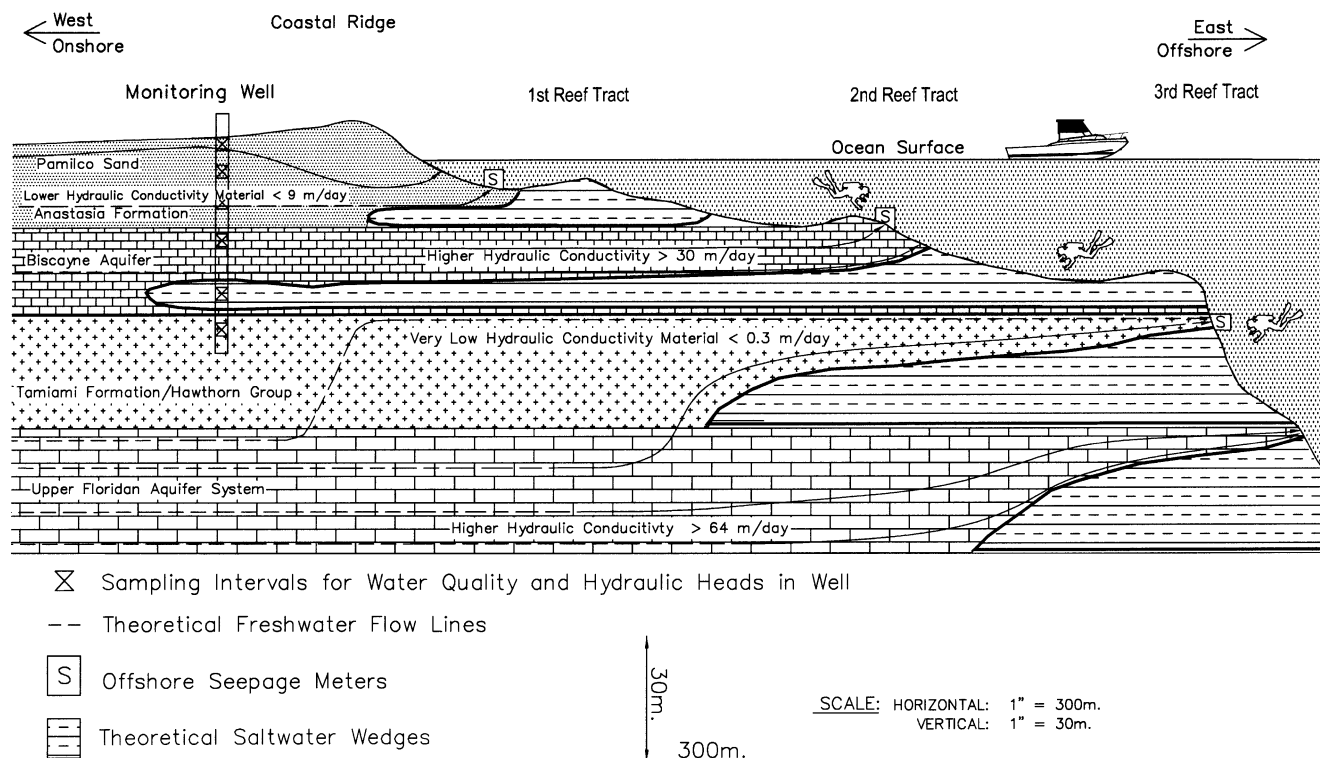
As with all models, the mass balance approach makes some assumptions about the system being modeled. These approaches and approximations overlook coastal system dynamics and may provide inaccurate results.

Chemical Mass Balance Method

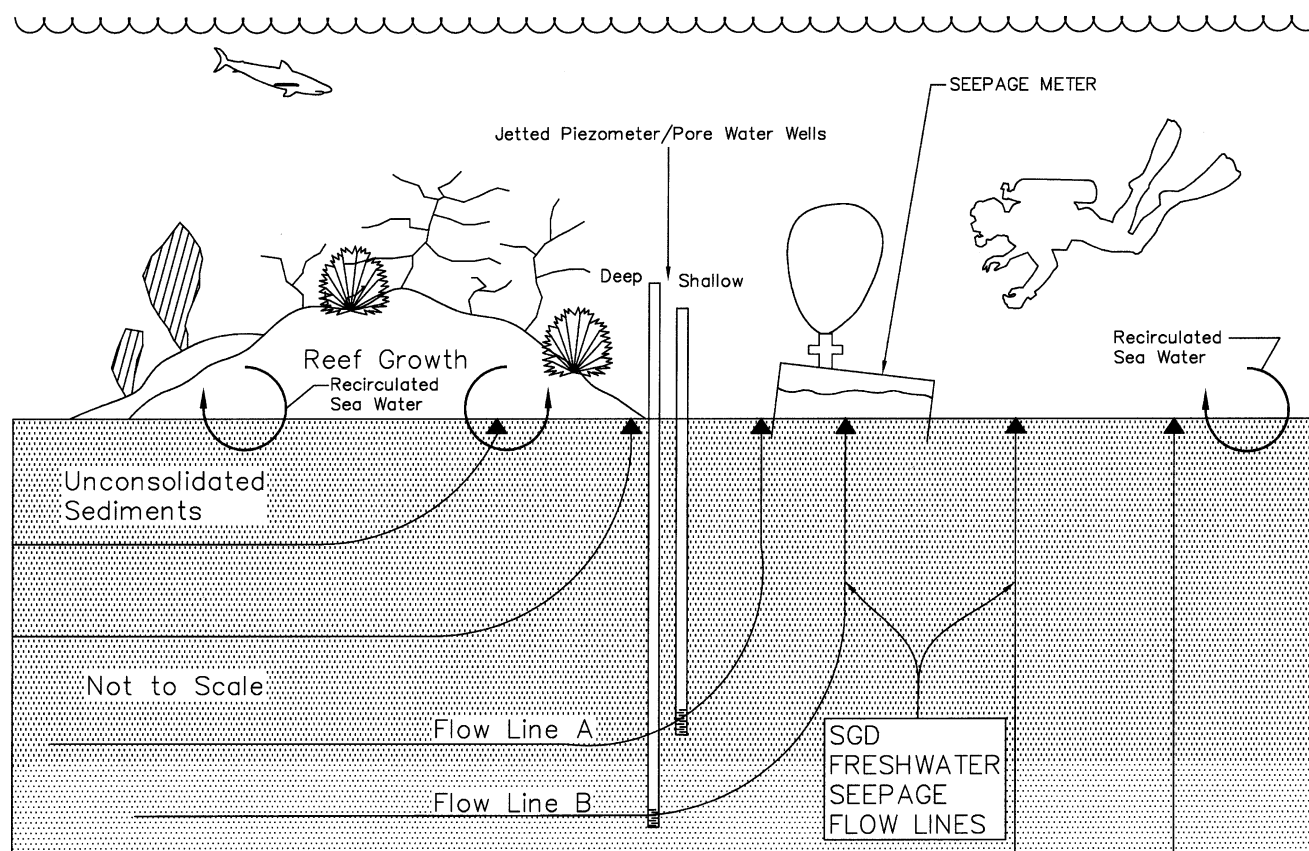
Natural tracers are also used to estimate SGD. Some common natural tracers include radioactive isotopes, silica, and chloride. Geochemical data can provide a long-term integrated picture of hydraulic processes in the aquifer and the groundwater/surface water exchange (Harvey et al. 2000, 2002). If a particular solute occurs in groundwater and is not in other sources (or is in reduced concentrations) and the solute concentration in each source is known, the information can be used to infer the magnitude of SGD (Church 1996; Moore 1996, 1997; Burnett et al. 2003). The magnitude of groundwater contributions has largely been ignored until recent isotope (^{236}Ra) work conducted in the South Atlantic Bight by Moore (1996). Moore indicates that groundwater contribution to the nearshore was approximately 40% of the riverine inputs during his study period.

Using Seepage Meters to Measure SGD

When using seepage meters to make direct SGD flux measurements, the seepage meters, piezometers, and monitor wells should be installed 2–3 days prior to sampling to allow the sediments to rebound and the system to return to equilibrium. SGD measurements made before equilibration may result in erroneous seepage rates (Belanger and Walker



Submarine Groundwater Discharge, Fig. 1 Generalized submarine groundwater discharge locations, lithostratigraphic, and theoretical flow line cross-section of the aquifer for Palm Beach County, Florida



Submarine Groundwater Discharge, Fig. 2 Typical nearshore seepage investigation set-up with theoretical freshwater flow lines, recirculated flow lines and deep and shallow monitoring piezometers

1990). The installation of traditional and automated seepage meters on the seafloor or lake/river bottom is a simple task. The installers usually stand on the seepage meter, gently rocking back and forth until the meter is pushed about 15–20 cm into the sediment surface. This can be difficult when the seepage meter is more than 3 ft (1 m) below the surface of the water. Placement of the meter is important; the threaded hole is usually placed on the upslope side of the sediment to allow the movement of water, air bubbles, and gases from the seepage meter to the bag.

Gas collection in the seepage meter can become a problem because gasses are produced by the decomposition of organic material within the seepage meters. Gas production creates additional challenges with automated seepage measuring devices. Several investigators using electronic flowmeters to measure SGD found that air bubbles sometimes render the device useless until a change in the tidal direction or wave action forces the bubbles to move (Krupa et al. 1998).

The sampling frequency of the traditional meters is variable and best determined by trial and error at each site. Broken seepage collection bags indicate higher seepage rates and require more frequent sampling. The sampling frequency may be as short as 0.5 h to as long as 12 h.

The seepage data must be adjusted to obtain accurate SGD calculations. Studies show that seepage meters require the use of a frictional coefficient to correct for the frictional losses in the meters and collection bags (Shaw and Prepas 1989; Belanger and Montgomery 1992). Recent tank test results by Belanger indicate that different meter designs require different correction factors, and each meter requires different correction factors for discharge and recharge situations. The seepage rate is multiplied by the coefficient to calculate the actual SGD rate.

Factors Affecting SGD Results

Several factors besides the actual groundwater discharge rates affect SGD measurements and calculations. Gaps in data collection may cause errors. Many of the SGD measurement methods collect data for limited time periods and may miss the enhanced groundwater discharge during storm events. The size of the study area is also a concern because studies have found that SGD varies spatially and temporally. These spatial and temporal differences can be challenging when measurements are required over large areas.

Care must be taken in estimating the groundwater flux. It is important to avoid including oceanic and fluvial fluxes more than once. Riverine groundwater discharge must be calculated between the lowest gauging station and the mouth of the river and biogeochemical processes must also be considered.

Sediment lithology of coastal aquifers will vary based on depositional environments. In the field, any combination of confined and unconfined aquifer scenarios exists. Davis and DeWiest (1966), Strack (1975), Bear (1979), and Todd (1980) include solutions for analytical solution of groundwater flow from the aquifer to the coast for various aquifer combinations including unconfined, confined, stressed (pumping and drainage), and natural flow conditions.

SGD is highly complex and is rarely treated in terms of fully three-dimensional (3D) density dependent miscible fluid flow in a porous medium (Reilly 1993). Instead, the system conceptualization is usually highly simplified to facilitate physical limitations and project budgets. The accuracy of analysis is dependent on the assumptions in the following three categories: (1) the mixing process, (2) the characteristics of the aquifer under study, and (3) the desired scale and detail of the resulting analysis (Reilly 1993). These issues are discussed in depth, along with a brief discussion on the mathematical basis of different solution methods in Reilly (1993).

Relating local measurements to regional fluxes can be an arduous task and can result in misleading conclusions. The calculations used to transform water-level measurements in wells, seepage meter rates, and radioisotope data into a flux for an area can be complicated and timeconsuming. Seasonal and tidal factors need to be understood and applied to *in situ* seepage results. Often, values obtained through analytical or numerical simulation were not verified in the field. Experience has shown that values determined by models may vary significantly from the estimates based on field measurements (Belanger and Walker 1990).

Traditional mathematical models do not consider the cyclic or transient discharges from the groundwater component. Many analyses, such as the Darcy solution, assume steady-state conditions (Davis and DeWiest 1966; Freeze and Cherry 1979) and overlook the transient vertical position of the land/ocean interface (Urish and Ozbilgin 1989).

Methodology can also affect seepage results. In traditional seepage meters, empty seepage bags appear to exert a slight suction that pulls water from the meter as the bags slowly unfold. Prefilling the bags reduces the anomalous short-term inflow and allows both discharge and recharge to be measured. Open water wave action and currents are reported to cause induced seepage rates (Libelo and MacIntyre 1994; Shinn et al. 2002). Harvey et al. (2002) collectively terms these methodology effects as apparent seepage rates.

A recent septic tank study near a tidally controlled estuary in Tequesta, Florida (US), demonstrates the tidal effect. *In situ* heat-pulse flowmeters in onshore monitor wells were used to measure horizontal velocities during high, low, and slack portions of the tidal cycles. The results showed groundwater velocities varying at low tide from 0.39 m/day to as high as 1.33 m/day in different layers of an unconfined aquifer (Harbor Branch Oceanographic Institute 1995). Robinson (1996) and others report seepage rates as high as six to eight times higher on an ebb tide. A wide range of *in situ* velocities can occur under varying tidal conditions and with differing aquifer heterogeneities. Therefore, groundwater velocity ranges should be evaluated at each study location.

Additional factors that need to be considered in the collection of SGD from seepage meters are summarized below:

- Mechanical and physical factors affecting seepage meter performance (Libelo and MacIntyre 1994; Harvey et al. 2000; Shinn et al. 2002)
- Meter-to-bag connectors and types of bags used (Lee 1977; Shaw and Prepas 1990; Belanger and Montgomery 1992)
- Pre-filled volume in the bag (Lee 1977; Shaw and Prepas 1989; Belanger and Montgomery 1992)
- Method of bag deployment and retrieval (Lee 1977; Belanger and Montgomery 1992; Harvey et al. 2000).
- Gas produced from decomposition within the sediment (Harvey et al. 2000).

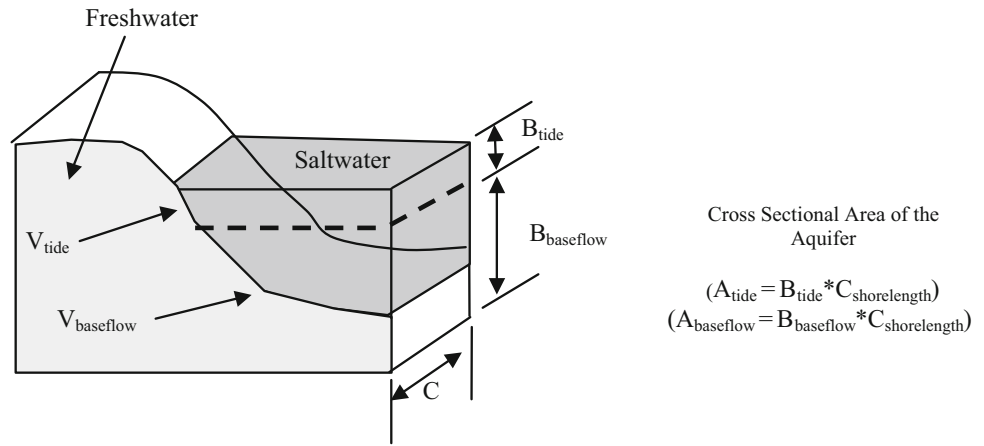
Determination of Seepage Rates

Traditional methods of calculating SGD use one groundwater velocity value derived from hydraulic gradients and hydraulic conductivity estimates. Applying one velocity component to a particular solution can be misleading, as can the approach of using a simplified model to describe a complex system. If measured groundwater velocities and/or gradients are available, this information should be used. Total discharge for the aquifer (Eq. 1) summarizes the products of the two velocities (tidal, baseflow/interflow) and their respective cross-sectional areas (Fig. 3). V_{tide} is defined as the discharge from the seepage face both during low tide and high tide. V_{baseflow} is the groundwater that is discharging regardless of the tide.

Empirical Solution Methods

Traditionally, seepage has primarily been estimated by applying Darcy's equation. This approach estimates discharge based on the hydraulic gradient between two points. The discharge is reported in units of length³/time (L^3/T). Darcy (1856) defined the specific discharge shown below:

Submarine Groundwater Discharge, Fig. 3 Generalized cross-sectional groundwater velocities model for tidally-controlled aquifer



$$Q_{\text{total}} = (V_{\text{tide}} * A_{\text{tide}}) + (V_{\text{baseflow}} * A_{\text{baseflow}}) \quad [\text{Eq. 1}]$$

$$q = \left(\frac{Q}{A} \right), \quad (1)$$

where q is the specific discharge (length/time); Q is the total discharge or recharge per unit width (length³/time); A is the cross-sectional area of the horizontal flow (length²).

By knowing the head loss across the medium and the distance over which it occurred the Darcy equation can be expressed in a differential form:

$$q = -K \frac{dh}{dl}. \quad (2)$$

Setting Eqs. 1 and 2 equal to each other, and solving for Q (Eq. 3) yields the total discharge from the aquifer to the surface water body:

$$Q = -KA \frac{dh}{dl}, \quad (3)$$

where, Q is the total discharge (length³/time); A is the cross-sectional area of the horizontal flow (length²); K is the horizontal hydraulic conductivity (length/time); dh/dl is the horizontal hydraulic gradient (unitless).

Seepage calculations using Darcy's Law are limited because the solution assumes steady-state conditions and does not consider transient conditions, such as tidal action, rainfall, or onshore surface water levels.

To estimate the SGD for a portion of a coastline, the total discharge of freshwater per length of coast would be multiplied by the length of coastline of interest.

$$Q_{\text{shore}} = -KA \frac{(h_1 - h_2)}{(l_1 - l_2)} L, \quad (4)$$

where, K is the horizontal hydraulic conductivity (length/time); L is the length of coastline (length); h_1 is the hydraulic head at point 1 (canal or aquifer water level); h_2 is the head at point 2 (generally the ocean); l_1 is the horizontal spatial location of point 1; l_2 is the horizontal spatial location of point 2.

If the SGD is occurring within a region of the aquifer that is sufficiently close to the shore and if the density of the groundwater varies, the density must be considered. The above equation can be modified to consider the fluid characteristics (Freeze and Cherry 1979).

$$Q_{\text{shore}} = - \left(\frac{k \rho g}{\mu} \right) A \frac{(h_1 - h_2)}{(l_1 - l_2)} L, \quad (5)$$

where, k is the permeability (L/T²); ρ is the density of the fluid (W/L³); g is the gravitational constant (L/T²); μ is the dynamic viscosity (M/LT).

Estimation of actual flow velocities in the aquifer requires that the porosity be considered. Equation 3 is divided by porosity to yield the solute velocity of the flow system (Freeze and Cherry 1979) as represented in Eq. 6:

$$V = \frac{q}{n} = \left(\frac{K}{n} \right) \left(\frac{dh}{dl} \right) \quad (6)$$

where V is the average linear groundwater velocity (length/time); Q should be small "of is the specific discharge (length/time)"; n is the porosity.

This approach, which assumes one seepage rate for a length of coast, is valid when the flow is laminar and the groundwater is not experiencing significant diffusion. When seepage rates differ significantly along the shore it is appropriate to determine the discharge rates for each coastal segment (Cherkauer 1998).

Numerical Solutions

The problems discussed earlier in applying Darcy's equation to SGD also exist in the application of numerical calculations for both non-density and density dependent models. The primary considerations with numerical simulations are as follows:

1. Accuracy of the representation of the aquifer sediment layering, density of the fluids, and the vertical and horizontal hydraulic conductivity.
2. There is limited availability of coastal discharge data around the world. Many numerical modelers/simulators do not check the coastal discharge rates during the simulation process; it is considered an afterthought for the model simulations. The true discharge and spatial locations may or may not be correct. Recent seismic work in support of a modeling effort (Merchant et al. 1996) off the North Carolina coast showed that paleofluvial channels and large-scale collapses constrain the possible discharge locations offshore discharge points.
3. Many numerical models are not designed to simulate density dependent flows, so modelers substitute simulated no-flow boundary conditions for the density difference in a stairstep fashion within the model domain (Fig. 4) or use equivalent freshwater heads. Other models account for density but require data that is frequently unavailable and so may still predict an incorrect flux across the boundary condition. As with other numerical simulations, the validity of the application is a function of the modeler's experience.

Additional Modeling Considerations

Locating the groundwater/oceanic water interface is critical because the morphology, flow patterns (seepage), and solute transport vary significantly with location site (i.e., barrier islands, atolls). Urish and Ozbilgin (1989) state that vertical position of the interface can vary as much as 1.5 m vertically and 61 m horizontally, and is a function of sediment permeability, effective diffusivity, wave amplitude, and wave period (Rasmussen 1998). Background data collection is essential to

understanding the existing porewater salinity profiles at each location.

Additional long-term considerations are rising sea levels and resulting increases in groundwater levels. Nuttle and Portnoy (1992) indicate that relative sea level is rising at 30 cm per 100 years along the east coast of the United States. As a result, groundwater levels are being raised on a regional scale in low-lying areas. With the groundwater level closer to ground surface, decreased aquifer storage will increase runoff and reduce the amount of available recharge to the aquifer, subsequently decreasing SGD to coastal areas.

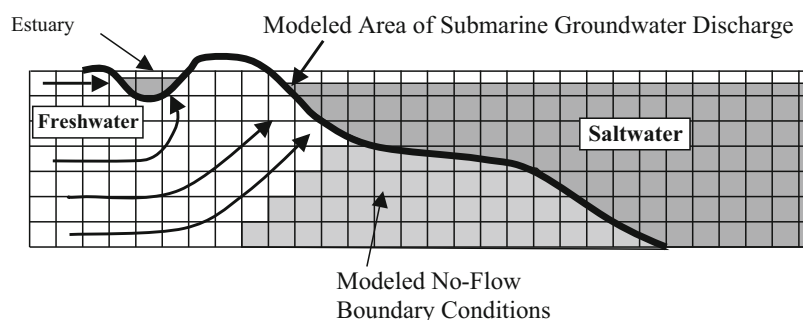
Empirical Solutions Versus Numerical Solutions

A comparison of solution methods for SGD. The South Florida Water Management District (SFWMD) has collected detailed measurements of surface water discharges and associated chemical surface water concentrations over a 9-year period for Palm Beach County, Florida (Finkl and Krupa 2000). Detailed groundwater levels and groundwater chemical concentrations in this area have been collected by the United States Geological Survey and SFWMD over 10 years. The SFWMD used a regional model, known as the South Florida Water Management Model (SFWMM), to simulate the entire southern peninsula of Florida (SFWMD 1993a, b). Shine et al. (1989) utilized a countywide numerical groundwater model (based on MODFLOW) to also estimate groundwater discharges to the Palm Beach County coast. Groundwater levels and surface water data were used in the models for calibration and verification.

Using these models, SGD values were calculated for the groundwater discharged from Palm Beach County to the offshore environment. For comparison, calculations based on Darcy's law were completed for the same length of coastline.

As seen in the above table, large variations in SGD estimation can occur in calculated and numerical model solutions (Table 1). The SFWMM had the lowest groundwater discharge value; this is likely because the model grid size (2 miles $\times$ 2 miles) does not allow the SGD process to be seen and measured. Two of the three solutions yielded SGD that equaled or exceeded the surface water inflows to the ocean. Corresponding groundwater and surface water

Submarine Groundwater Discharge, Fig. 4 Typical finite difference grid used in modeling coastal boundary conditions in non-density dependent model



Submarine Groundwater Discharge, Table 1 Comparison of three solutions of submarine groundwater discharge compared to actual surface water discharges and nutrient fluxes to offshore Palm Beach County, Florida

Discharge type	Water (million-meters ³ /year)	Phosphorous flux (P) (metric tons)	Nitrogen flux (N) (metric tons)
Surface Water Discharge Surface water (actual flow)	1,661	196 ^a	2,471 ^a
Groundwater Discharge Darcian flow solution ^b	2,211	414	5,727
USGS MODFLOW Application ^c	1,659	196	2,469
SFWMM SW/GW model ^d	50	6	75

^aSFWMD 1993b, average annual discharge for 9 years * average annual load for 9 years

^bDarcy's flow assumptions; $i^1 = 0.001$; $i^2 = 0.006$; $k^1 = 3.04$ m/day; $k^2 = 30.48$ m/day; coast length = 72.46 km; $b^1 = 18.2$ m; $b^2 = 45.7$ m; Load numbers total phosphorous (P) = 0.188 mg/L, Total Nitrogen (N) = 1.67 mg/L

^cShine et al. 1989, Average Annual

^dSFWMD 1993a, Average Annual Discharge for 9 years

concentrations for phosphorous and nitrogen were then calculated using the SGD fluxes to yield nutrient loading to the nearshore. It has been estimated that, on a worldwide basis, the total contributions of freshwater and nutrients to the ocean by SGD are roughly equivalent to the total contributions from riverine (in this case canal) discharges (Johannes and Hearn 1985). Based on the estimated results and statements mentioned above, SGD is nearly equal to the surface water discharges in Palm Beach County, Florida.

Summary

Governmental agencies and private organizations need SGD information to better address water degradation issues and to create numerical models. Accurate determination of SGD is essential in calculating mass balances and in the determination of nutrient fluxes to the coastal waters. Experience has shown that quantifying this input is difficult because of uncertainties in the direct measurement of groundwater flux and in the simulation of groundwater fluxes.

Accurate knowledge of SGD is important because it can be an unseen hazard and can be used to assess environmental problems in coastal environments. The contribution of SGD to the coastal hydrologic regime is occasionally recognized in association with crescendo events and concurrent marine algal blooms that degrade water quality, bottom habitats, and coral reef ecology (Finkl and Krupa 2003). However, the more common situation is that SGD laden with nutrients from agrourban activities on adjacent coastal plains causes environmental degradation so gradually that the cause and effect can escape public attention.

Cross-References

- Hydrology of the Coastal Zone
- Shoreface
- Water Quality

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Submerged Coasts

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Definition

A submerged coast is defined as a coast resulting from the relative submergence of a landmass either through eustatic sea-level rise and/or crustal subsidence against subaerially produced forms and structures. The term carries no implication as to whether it is the land or the sea that has moved. Based on seafloor mapping, the study of Quaternary sedimentary sequences, and, the recognition of erosional and depositional features on continental shelves, submerged coasts are increasingly recognized. In siliciclastic-dominated shelves, drowned landforms including economic mineral deposits of sand and gravel, placer deposits such as diamonds, gold, cassiterite, and heavy mineral sands may be present, while in carbonate-dominated shelves, drowned coral reef terraces, karst landforms including karstification surfaces may be present.

IGCP Contributions

Three International Geoscience Program (IGCP) projects from 1996 to 2010 involving participants from over 50 countries have made important contributions to the recognition of submerged coasts. They include:

1. IGCP project number 396 “Continental shelves in the Quaternary” from 1996 to 2000 (Yim et al. 2002a).
2. IGCP project number 419 “Subaerially exposed continental shelves” from 2001 to 2005 (Yim et al. 2008a).
3. IGCP project number 529 “Continental shelves, sea levels and environments” from 2006 to 2010 (Catto et al. 2009).

Continental shelves where submerged coasts occur in the greatest abundance are geographic regions ignored by the international ocean drilling community. A rare exception is Expedition 310 of the Integrated Ocean Drilling Program which is aimed at studying the last deglacial sea-level rise in the South Pacific. However, this location is not even a continental shelf.

Recognition and Dating

Submerged coasts are indicated by the occurrence of drowned subaerial depositional and erosional features (Table 1). Among the earliest drowned subaerial features recognized are the submerged forests off the English coast (Lyell 1850). On continental shelves of the world, about 70% of the total area was identified by Emery (1968) to be covered by relic sediments including previously deposited subaerial, lacustrine, and paludal sediments. For sea-level changes during the past 20,000 years, marine zonation, erosional indicators, depositional indicators, archaeological remains, and historical data provide evidence for submergence (Pirazzoli 1996).

The siliciclastic-dominated continental shelf in the northern South China Sea off Hong Kong has been studied in detail because of opportunities provided by coastal development including seafloor excavation and the availability of high quality offshore boreholes. Research investigations have revealed a sedimentary sequence dating back to about 0.5 million years comprising of five interglacial-glacial cycles (Yim 1994). The subaerial exposure of pre-Holocene interglacial marine deposits during glacial low sea-level stands had resulted in acid-sulphate soil development to form paleosols (Yim and Tovey 1995). The affected marine deposits are characterized by a lower moisture content and a higher density (Bahr et al. 2005) and was found by magnetic susceptibility profiling of continuous boreholes to be caused by iron enrichment (Yim et al. 2004). Soil microfabric analysis assisted by scanning electron microscopy trace-element mapping had resulted in their identification as paleo-desiccated crust (Tovey and Yim 2002).

Submerged Coasts, Table 1 Selected depositional and erosional features associated with submerged coasts

Depositional features	Erosional features
Terrestrial deposits (alluvium, colluvium, aeolianite, paleosols, peat and other plant remains, etc.)	Notches
Karst, e.g., speleothems	Honey-combed
Coral reef	Rock surfaces
Beachrock	Shore platforms
Archaeological remains	Karst, e.g., caverns

A review of submerged coasts in coral reefs in response to Quaternary sea-level changes is provided by Camoin and Webster (2015). The study of long drill cores from locations including fringing reefs and atolls revealed episodes of active coral reef growth during interglacial periods either following or during marine transgressions while glacial periods are represented by erosion surfaces formed by their subaerial exposure including the development of karst features and pedogenic iron enrichment.

For dating submerged coasts, the radiocarbon method and the U/Th method are widely used. The former was first applied to stable crustal regions for the reconstruction of late Quaternary sea-level history (Godwin et al. 1958). In Barbados, the drilling and dating of coral reefs containing *Acropora palmata* indicated that sea level was at -121 ± 5 m around 17,000 yr. BP (^{14}C) (Fairbanks 1989). Subsequent dating of the deepest *A. palmata* using the U/Th method gave an age of ca. 19,000 yr. BP corresponding to a sea level of -118 m (Bard et al. 1990a). Far away from plate boundaries in Tahiti, a large sea-level jump was identified shortly before 13,800 yr. BP with the U/Th method while radiocarbon ages on the same samples were significantly younger (Bard et al. 1996). Dating of inner-shelf sequences off Hong Kong (Yim et al. 1990; Yim 1999) also indicated a young age bias of pre-Holocene radiocarbon dates exceeding about 8200 yr. BP (^{14}C), while some of the shells with finite radiocarbon dates may instead be of last interglacial age. Pre-Holocene materials on the shelf are likely to yield unreliable radiocarbon dates compared with Holocene counterparts because of possible pedogenic alteration during low sea-level stand(s). For a comparison of radiocarbon and U/Th dates, see Bard et al. (1990b). Because there is a maximum difference of about 3500 years for a date of ca. 20,000 yr. BP (^{14}C), the last glacial maximum (LGM) dated to 18,000 yr. BP (^{14}C) may have occurred 21,500 sidereal years ago.

On the northern South China Sea continental shelf, the reconstruction of sea-level changes based on the use of radiocarbon dating alone was found to be problematic prior to the 8200 calendar yr. BP cold event (Yim et al. 2006). This is attributed to the uncertainty of the radiocarbon dates exceeding this age limit due to large variations of atmospheric radiocarbon and/or sample contamination caused by subaerial exposure during the last glacial period.

Optically stimulated thermoluminescence dating and moisture content distribution in a seafloor drill core from a siliciclastic-dominated continental shelf was successfully used for dating submerged coasts (Yim et al. 2002b). Three samples from a terrestrial sedimentary unit in an inner-shelf borehole from the new Hong Kong International Airport site yielded oxygen-isotope stage (OIS) 8 ages. In another study of Quaternary superficial deposits from two shallow bays in Hong Kong (Yim et al. 2008b), last glacial maximum ages

were obtained from the youngest terrestrial unit, while only minimum ages can be obtained from the older glacial terrestrial units. In both these studies, moisture content profiling of the borehole samples was found to be important to provide indications of the episodes of subaerial exposure of the terrestrial deposits prior to their burial by younger marine deposits of interglacial age(s).

Sea-Level Minimum During the Last Glacial Maximum

The global average depth of the shelf break at -130 m below present mean sea level (Shepard 1973) provides a median value of sea-level lowering resulting from continental ice growth. Since this value is based on the compilation of hydrographic charts, it is probably free from bias (Bloom 1983). Additional support for the sea-level minimum at -130 m is the change of about 10 m in global sea level through a 0.1‰ change in the oxygen-isotopic record (Shackleton and Opdyke 1973) and the closeness to the median value sea-level position during the LGM identified on many shelves.

The radiocarbon dating of oolitic and biogenic carbonates associated with drowned beach barriers is a means of identifying the sea-level minimum during the LGM. On the Bengal Shelf, five samples of these materials from depths ranging from 125 to 133 m below the present sea level yielded dates ranging from 16,500 to 24,900 yr. BP (^{14}C) (Wiedicke et al. 1999).

Examples of Submerged Coasts

In North America, subaerial features of LGM to late glacial maximum age drowned by the Holocene transgression were reviewed by Bloom (1983). Terrestrial materials and landforms including mastodon and mammoth fossils, Indian artifacts, moraines, river channels, etc. were reported.

Clear-cut cases of coastal subsidence resulting from earthquakes have been documented. One such example is from near Haikou on Hainan Island in southern China where ruins of ancient villages, including a cemetery with engraved tombstones of 1604, were identified up to 10 m below present sea level (Y. Chen in Anonymous 1983). A coastal area exceeding 10 km by 1 km was affected by subsidence during a Ming Dynasty earthquake recorded in historical documents. The earthquake was estimated to be of magnitude scale 8 and occurred on July 13, 1605.

A long history of submerged coasts has been found on “stable” shelves. Off Hong Kong, the study of offshore borehole sequences revealed a succession of paleo-desiccated crusts formed by the subaerial exposure of marine deposits during Quaternary low sea-level stands dating back to OIS 2, 6, 8, and 10 (Yim and Tovey 1995). This sequence is in agreement with the five interglacial-glacial cycles identified in the Vostok ice core in Antarctica (Petit et al. 1999) confirming that the sea-level changes found were eustatically controlled. Figure 1 shows an example of an early Holocene submerged coast dated at ca. 8000 yr. BP (^{14}C) when sea level was at ca. -18 m below present mean sea level.

Submerged Coasts, Fig. 1 An early Holocene submerged coast when sea level was at ca. -18 m below present mean sea level. The exposure was available through seabed excavation for the foundation of the West Dam, High Island Reservoir, Hong Kong SAR. The Holocene marine deposits (*darker*) rest unconformably on a fining-upward sequence of Pleistocene alluvial deposits (*paler*). Wood and shells from near the base of the Holocene marine deposits yielded dates of ca. 8000 yr. BP (^{14}C) (Photograph courtesy of R.J. Frost)



An example of a long record of submerged coasts occurring in coral reef is provided by the 210-m long drill core into Ribbon Reef on the outer Great Barrier Reef (Webster and Davies 2003). Based on petrographic, isotopic, and seismic characteristics, the core was divided into three sections: (1) a main reef section from 0 to 96 m composed of six interglacial reef units, (2) a section from 96 to 158 m enclosing two thin reef units, and (3) a basal section from 158 to 210 m composed of nonreefal debris. The cyclic nature of submarine coasts were caused by interglacial marine transgressions triggering coral reef growth followed by their subaerial exposure and karstification during glacial periods.

Micropaleontological studies of continental-shelf cores are useful in confirming submerged coasts when they penetrate the formerly exposed soil surface. A pollen sequence identified on the east Queensland shelf shows a transition from terrestrial vegetation (grasses and woodland) to salt marsh and mangroves (Grindrod et al. 1999). This sequence is the reverse of mangrove successions recorded for regressive coasts and is compatible with a drowning coast. Similarly, a diatom sequence in a shelf core off Hong Kong with evidence for five marine transgressions showed a record of diatom preservation consistent with aging and pedogenesis during glacial period(s) (Yim and Li 2000). The tests of diatoms were progressively destroyed through groundwater dissolution during the low sea-level stands with the lowest abundance found in deposits of OIS 11, followed by OIS 9 and OIS 7. Deposits of OIS 5 were similar in diatom abundance to OIS 1 probably because of the “young” age.

Underwater mapping of rocky coasts using SCUBA is a possible way of identifying submerged coasts of late Quaternary age. Off Marseille submerged coasts were shown by the Grotte Cosquer archeological site at a depth of about 60 m below present sea level and a paleo-coast at a depth of about 90 m below present sea level dated at 13,800 yr. BP (^{14}C) (Collina-Girard 1996). Underwater mapping off Corsica and the West Indies have revealed discontinuities at common depths of -11, -17, -25, -35, -45, -55, and -100 m below present mean sea level attributed to late Quaternary submerged coasts. Because notches at similar depths have also been found in the French West Indies (Collina-Girard 2002), these notches are suggested to have been formed by the same lower eustatic sea levels during the late Quaternary.

The sedimentary record of paleoenvironments and sea-level change through the last glacial cycle in the Gulf of Carpentaria in northern Australia was studied (Reeves et al. 2008). The record included open shallow marine, marginal marine, estuarine, lacustrine, and subaerial exposure including confirmation of the existence of the paleo-Lake Carpentaria. During OIS 6, the Gulf was largely subaerially exposed and was traversed by meandering rivers.

The value of archeological remains for estimating the former elevation of submerged coasts is most promising within the Mediterranean Sea because of the relatively small tidal range (Auriemma and Solinas 2009). A range of manmade features including fishponds, piers and quays, coastal quarries, wells, settlements, and beached wrecks may be used. Out of these, fishponds, piers and quays were found to be the most reliable due to their coastal functions.

Future Work

In spite of the advances made on submerged coasts by a “small” community of continental shelf researchers, this geographical region is still neglected in comparison to the deep oceans because of the absence of an internationally funded drilling program. Additionally, the exclusive economic zone prescribed by the United Nations Convention on the Law of the Sea is a sea zone of nations including continental shelves stretching from the present coast out to 200 nautical miles (nmi) may have disputed territorial claims by multiple countries. The high cost needed for offshore work including marine surveying and drilling is another obstacle for developing nations but this may be overcome by collaborative studies with developed countries. A further obstacle is the disturbance of the seafloor of continental shelves caused by human activities including trawling, dredging, mining, installations of pipelines and cable routes, etc. is also making it more challenging for future studies on submerged coasts.

Summary

Submerged coasts which demarcate former coastlines are formed by Quaternary sea-level changes and/or crustal subsidence. They occur in the shallow ocean usually less than 200 m in water depth. Their studies have been much neglected in the past because of the absence of a funded international drilling program like that operating on the deeper ocean floor.

Submerged coasts may be divided into siliciclastic-dominated and carbonate-dominated types. In this account, the recognition and dating of submerged coasts with selected examples are provided. Studies on the position of former submerged coasts may be an indicator of ocean volume changes.

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Submerging Coasts

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A coast is submerging when the relative sea level rises above it. Submergence may be caused by sea-level rise, by land subsidence, or by the two.

Between about 20 and 6 kyr ago, when the melting of the Northern Hemisphere continental ice caps was completed, land-ice melting caused the global sea level to rise some 120 m, at an average rate of 8 mm/year, but with peaks reaching 40–50 mm/year during certain periods (Bard et al. 1996). This rise caused the rapid submergence of huge continental shelf areas. Since about 6 kyr ago, the global sea level has remained almost stable in a high, interglacial position. Global sea-level rise for the last

Paolo A. Pirazzoli: deceased.

Submerging Coasts, Fig. 1

What is visible is not a fringing coral reef at sea level, but the result of a seismic subsidence that occurred in 1981 during earthquake at Mavrolimni, in the Gulf of Corinth, Greece. Before the earthquake, Mavrolimni (literally Black Lake) was a lagoon, up to 10 m deep, connected to the sea by an opening to the north. It was delimited seawards by a sand-and-gravel barrier, a few meters wide, which was used as a mole for fishing boats. In 1981, the area subsided about 0.8 m and the top of the barrier was brought down just to sea level (Stiros and Pirazzoli 1998). (Photo D554, August 1992)



century is estimated to be of decimetric order. For the next century, climatic models that take into account increasing greenhouse effects, have predicted scenarios of sea-level rise between 0.09 and 0.88 m, with a central value of 0.48 m (IPCC 2001).

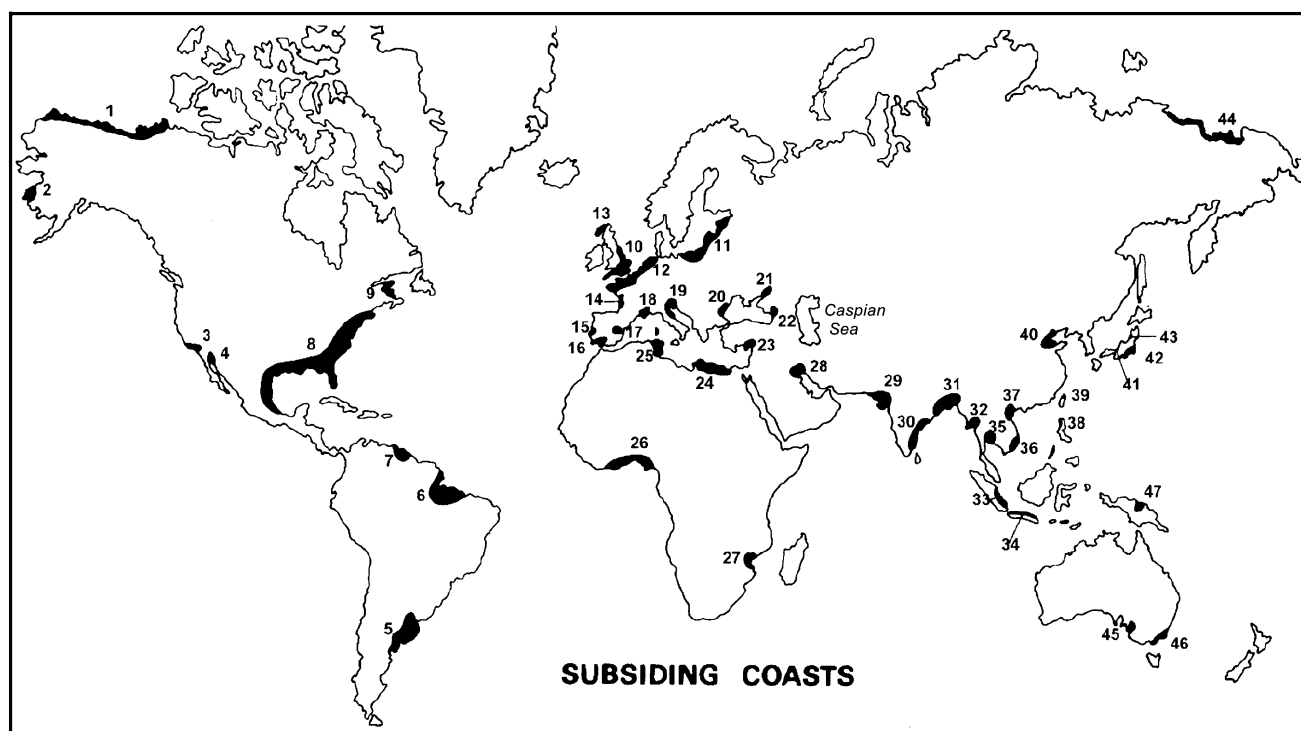
Land subsidence may result from several (neo)tectonic processes. Its rate can be much variable in space and time. Since the peak of the last glacial times (i.e., during the last 20 kyr), in deep-sea areas hydro-isostatic displacements have been of the order of 40 m (about one-third of the relative sea-level rise, that is, an average rate of 2 mm/year). This made the ocean to floor subside. When approaching coastal areas, subsidence rates lessened, depending on the water shallowness and the coastal topography. During the last 6 kyr such isostatic subsidence decreased exponentially. Near delta formations, however, sediment compaction and sedimento-isostasy have often been more significant, with rates of subsidence on the order of one to a few millimeters per year reaching the maximum values near the delta depocenters. During the last centuries, and especially in the twentieth century, human-induced land subsidence was caused in many coastal areas by acceleration of compaction due to drainage, oil and gas extraction, or groundwater exploitation. Though the total amount of human-induced subsidence rarely exceeds a few meters, it was often obtained within a limited time range, generally a few decades, thus reaching during these periods dangerously high sinking rates.

As a result of the above processes, most of the world coastal regions have been submerging rapidly between about 20 and 6 kyr ago, due to the rapid postglacial eustatic

rise. During the Holocene, most deltaic sequences began to accumulate between 8.5 and 5.5 kyr ago, with a modal age of about 7.5–7.0 kyr (Stanley and Warne 1994). It is, therefore, only after that time, that is, after the end of the post-glacial transgression, that land sinking may have started to be active in delta areas. Similar sinking, for recent sediment compaction, must have occurred, though at slower rates, also in most estuarine and lagoonal areas, and even in coastal plains. However, as long as the fluvial sediment input or the longshore drift could compensate for the land sinking, submergence phenomena could not start. It is especially during the last century that human action has been most active, not only in accelerating land sinking rates, but also in dredging from river beds or constructing breakwaters that cut off a longshore sediment supply.

Shallow coastal areas may also be submerged by marine erosion, in particular at the time of strong storm surges. In seismically active areas, coastal submergence may occur at the time of major earthquakes (Fig. 1). One of the most impressive last century events of this kind was probably the great earthquake of March 27, 1964 in Alaska (magnitude ≥ 8.4 , with an epicenter in the Prince William Sound area) when vertical crustal movements affected a region of at least 200,000 km², with a wide coastal zone of subsidence reaching a maximum of 2.2 m (Plafker 1965). In the Rann of Kutch, on the border between Pakistan and India, the 1819 earthquake resulted in a very wide area subsiding beneath the sea (Bird 1993).

In volcanic areas, vertical movements can be very fast and lead to local submergence phenomena. This is the case for many calderas, produced by explosion, collapse, or even



Submerging Coasts, Fig. 2 Sectors of the world's coastline that have been subsiding in recent decades, as shown by evidence of tectonic depression, increasing marine flooding, geomorphological and ecological indicators, geodetic surveys, and groups of tide gauges recording a rise of mean sea level greater than 2 mm/year (assumed to be the present rate of global sea-level rise) over the past three decades. A distinction is made between submerging coasts (where sea level has risen relative to the land) and subsiding coasts (where there has been land subsidence). On the coasts of the Caspian sea level has risen by more than 2 m since 1977 as a result of an increase in water volume, possibly accompanied by some associated subsidence of the bordering land. (From Bird 1993, © Geostudies, reproduced with permission of John Wiley & Sons Limited). *Key to the map:* 1, Mackenzie delta and northern Alaska; 2, Yukon delta, Alaska; 3, Long Beach area, southern California; 4, Colorado River delta, head of Gulf of California; 5, Gulf of La Plata, Argentina; 6, Amazon delta; 7, Orinoco delta; 8, Gulf and Atlantic coast, Mexico and United States; 9, Maritime Provinces, Canada; 10, southern and

eastern England; 11, the southern Baltic from Estonia to Poland; 12, North Germany, the Netherlands, Belgium and northern France; 13, Hebrides, Scotland; 14, Loire estuary and the Vendée, western France; 15, Lisbon region, Portugal; 16, Guadalquivir delta, Spain; 17, Ebro delta, Spain; 18, Rhône delta, France; 19, northern Adriatic from Rimini to Venice and Grado; 20, Danube delta, Rumania; 21, Eastern Sea of Azov; 22, Poti Swamp, southeastern Black Sea coast; 23, Southeast Turkey; 24, Nile delta to Libya; 25, Northeast Tunisia; 26, Nigerian coast, especially the Niger delta; 27, Zambezi delta; 28, Tigris-Euphrates delta; 29, Rann of Kutch; 30, Southeastern India; 31, Ganges-Brahmaputra delta; 32, Irrawaddy delta; 33, Eastern Sumatra; 34, Northern Java deltaic coast; 35, Bangkok coastal region; 36, Mekong delta; 37, Red River delta, northern Vietnam; 38, Manila, Philippines; 39, northern Taiwan; 40, Hwang-ho delta; 41, Maizuru, Japan; 42, Head of Tokyo Bay; 43, Niigata, Japan; 44, East Siberian coastal lowlands; 45, Port Adelaide region; 46, Corner Inlet region; 47, Sepik delta

erosion. Relatively slower movements, probably produced by thermo-isostatic processes, may continue for long periods, that is, for centuries or even millennia in the case of dormant volcanoes, or even for million years for extinct volcanoes capped by oceanic atolls.

If the greenhouse-induced warming predicted by climatic models will be confirmed, a significant relative sea-level rise can be expected during the next centuries. According to IPCC (2001), if greenhouse gas concentrations were stabilized (even at present levels) sea level would nonetheless continue to rise for hundreds of years. It can therefore be expected that submerging coasts, presently confined mainly to sectors where the land has been subsiding (Fig. 2), will become more extensive.

Cross-References

- [Changing Sea Levels](#)
- [Coastal Subsidence](#)
- [Isostasy](#)
- [Sedimentary Basins](#)
- [Tectonics and Neotectonics](#)

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Surf Modeling

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The purpose of surf modeling is to realistically predict and quantify surf characteristics such as breaker heights, breaker types, breaker periods, and surf zone widths. Many surf models also provide other useful surf zone information such as wave setup and wave-generated longshore currents. Increasing computer capabilities and their availability, as well as research advances, have enabled determination of surf characteristics to evolve from use of relatively simple calculations, nomographs, or tabular methods, that are largely based on empirical relationships, to use of numerical computer models. These models solve mathematical equations that are based on the physics of surf zone hydrodynamics (see *Surf Zone Processes*). But, only relatively recently have computer surf models been developed and applied. In the surf zone, wave breaking, turbulence, non-linear processes, and other complex phenomena combine to make modeling challenging. Here, the most important aspects of surf modeling for research and practical applications are described.

Modeled Parameters

Owing to their potential adverse effects, breaker heights and breaker types are of most interest. A breaker's height is the distance between its trough (lowest point) and crest (highest point). Higher waves break further offshore where depths are deeper so that breaker heights vary across a surf zone. Because incident waves (see *Waves*) have varying heights with time, breaker heights also vary with time. The intervals between individual breakers, breaker periods, are the same as the periods of the incident waves and, thus, are more easily predicted than breaker heights and types.

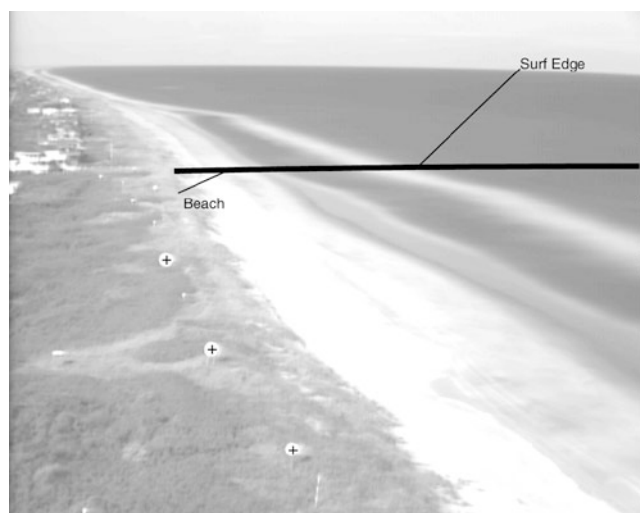
Breakers are categorized as spilling, plunging, or surging (e.g., Wiegell 1964; Dean and Dalrymple 1984; Fredsøe and Deigaard 1992; Smith 2000). A breaker type, collapsing, is

sometimes defined as an intermediate category between plunging and surging. A spilling breaker is characterized by an unstable crest that cascades down the shoreward face of the wave as a turbulent layer. For typical waves and bottom slopes of most beaches, spilling breakers occur most often. A plunging breaker is characterized by an unstable crest that curls over the shoreward face of the wave and plunges into the water below. Plunging breakers are of most concern for surf zone activities due to the forces of their crashing motion. A collapsing breaker is one for which the crest remains unbroken while the lower part of the shoreward face of the wave steepens and falls to produce a turbulent region. A surging breaker has a crest that remains essentially unbroken as it moves up a beach. For typical waves and bottom slopes of most beaches, surging breakers seldom occur. Most surf models lump collapsing breakers with surging breakers.

Because it represents the distance across possibly hazardous surf, the width of the surf zone is a key parameter. At any time, there is a distribution of wave breaking locations associated with the distribution of incident wave heights (see *Waves*). Higher waves break further offshore than lower waves. A practical definition is that surf zone width is the distance from the shoreline to where a trained conscientious observer sees the limit of depth-induced breaking over a reasonable time period, such as 10 min. This definition is preferable to defining the surf zone as the distance between the beach and the outermost breaker since the location of the outermost breaker varies with time. An analysis of over 600 video images of the surf zone shows that this practical definition results in surf zone widths that are most highly correlated to a location where 10% of the waves are breaking (Earle 1999). Surf zone widths generally vary inversely with water levels that are caused, at most locations, mainly by astronomical tides (see *Tides*). For beaches without offshore bars, higher water levels cause waves to break on the steeper nearshore part of the beach resulting in more narrow surf zone widths. For beaches with offshore bars, higher water levels may allow waves to pass over the bar, or bars, without breaking while lower water levels may contribute to depth-induced wave breaking near the bar, or bars. An example of a 10 min average video image for Duck, North Carolina, shows wave breaking over an offshore bar and closer to the beach (Fig. 1). For this example, the surf zone width along the indicated traverse was 44 m.

Longshore currents that move essentially parallel to the beach, also called littoral currents, are generated by waves breaking at angles relative to the beach. These currents are important for sediment transport and beach erosion applications as well as operations in the surf zone.

Rip currents are hazardous intermittent currents that cannot be predicted reliably by present surf models. Rip currents are strong currents that are confined to a narrow jet that flows seaward from the surf zone in a direction perpendicular to the



Surf Modeling, Fig. 1 Ten minute average video image showing a surf zone with wave breaking over an offshore bar and at the shoreline. Surf zone width, 44 m, from the shoreline (labeled Beach) at the time of this image past the main outer region of wave breaking (labeled Surf Edge) was determined from this image along the indicated traverse

beach. Several methods by which rip currents are generated have been proposed, but successful modeling requires further research advances.

Other parameters may also be modeled, in some cases as part of the calculations to predict the parameters of main interest. However, most other parameters are generally considered more important for research than for practical applications. A possible exception is wave setup which is relatively simple to estimate and which somewhat affects the total water depth that largely determines breaker heights. Wave setup is a wave-induced increase in mean water level near a beach. Other parameters include wave runup, swash, water motions at periods longer than incident wave periods (e.g., infragravity waves), and cross-shore currents including undertow.

Estimation Methods

Without using numerical models, estimates of where waves break and breaker heights can be obtained using simple procedures. The relationship that a wave breaks when its height (crest to trough) exceeds 0.78 multiplied by water depth has long been used in engineering practice as a good first estimate. In the ocean, many component waves, each with different wave heights, periods, and directions, combine to produce a wave spectrum. The root-mean-square wave height that is computed from a spectrum of waves is approximately given by 0.42 multiplied by water depth in a surf zone, where essentially all waves are breaking (e.g., Smith 2000). These types of breaking criteria have often been combined with

equations for wave shoaling and wave refraction (see *Waves and Wave Refraction Diagram*), assuming locally parallel depth contours, to develop engineering methods that yield useful estimates of breaker heights (e.g., Wiegel 1964; Dean and Dalrymple 1984; Smith 2000).

Several surf “similarity” parameters that typically depend on breaker height, local bottom slope, and wave period are used to categorize breakers as spilling, plunging, or surging (e.g., Guza and Inman 1975; Wright and Short 1983; Smith 2000). The various formulations provide reasonably similar results. Use of these parameters shows that spilling breakers are most common and that surging breakers are relatively rare.

Mainly for coastal engineering applications, estimation methods have also been developed for other types of surf zone information such as wave setup, runup, and longshore currents (e.g., Wiegel 1964; Dean and Dalrymple 1984; Fredsøe and Deigaard 1992; Coastal and Hydraulics Laboratory 1996, 1980–2000; Smith 2000).

Numerical Modeling

To characterize surf realistically, however, numerical models are needed. A recent research approach is numerically solving the hydrodynamic equations of fluid flow that govern shallow water waves in the time domain throughout a near-shore region. Boussinesq type models are probably the most common models of this type. Fredsøe and Deigaard (1992) and Svendsen and Putrevu (1996) provide overviews. Equations for variable nearshore bathymetry can be solved for water elevations and currents. Some models solve the fully nonlinear equations of motion that are usually expressed using a velocity potential. Time domain models are mathematically complex and computationally intensive. But, they can consider both cross-shore and longshore depth variations. These models have the potential to better predict wave-induced currents, including rip currents, than simpler one-dimensional models. Because there are several unresolved research issues, such as considering wave breaking, time domain models are not used yet for applied surf modeling.

Where a wave breaks is controlled largely by local depths in the immediate vicinity of breaking. Thus, one-dimensional surf models that consider local depth variations from outside of the surf zone to the shoreline are quite successful in determining breaker characteristics even though longshore depth variations are not considered. One-dimensional models are based typically on the energy transport equation given by

$$\frac{\partial(E_w C_g \cos \theta)}{\partial x} = -\langle \epsilon_b \rangle$$

where E_w is total wave energy, C_g is wave group velocity, θ is wave direction relative to a normal to the beach, x is a distance

from outside the surf zone to the land-water interface, and $\langle \varepsilon_b \rangle$ is the average rate of dissipation of wave energy per unit area of sea surface due to wave breaking. Dissipation due to bottom friction is small compared to that caused by wave breaking. E_w is given by

$$E_w = \frac{1}{8} \rho g H_{\text{rms}}^2,$$

where ρ is water density, g is the acceleration due to gravity, and H_{rms} is root-mean-square wave height.

Using various methods for calculating the dissipation rate, the energy transport equation is solved by numerical methods beginning outside the surf zone and progressing to the shoreline. This approach models energy and associated wave statistics rather than time varying characteristics of individual waves. A model that has been modified for several applications (e.g., Earle 1999) was developed by Thornton and Guza (1983). Because most breakers are spilling breakers that resemble bores, equations for the energy dissipation from bores (e.g., Le Mehaute 1962) are often employed. Other equations for wave height decreases caused by wave breaking can be used and provide generally similar results (e.g., Dally 1990). Either wave-by-wave calculations (e.g., Dally 1990; Dally 1992) or probability calculations (e.g., Thornton and Guza 1983) can be used to estimate the average dissipation. For wave-by-wave methods, wave heights are usually selected so that they represent a Rayleigh distribution (Longuet-Higgins 1952) outside of the surf zone, but joint wave height and period probability distributions (e.g., Longuet-Higgins 1983) may also be used. For probability methods that follow Thornton and Guza (1983), the average dissipation is obtained by integrating, over all wave heights, a bore dissipation function multiplied by the probability that a wave of a given height is or has broken.

Because it involves relatively few calculations and wave breaking probabilities can be used to estimate probabilities of different breaker types, a probability approach is attractive. Surf "similarity" parameters allow categorizing breakers as spilling, plunging, or surging. Given beach slope and wave period, values of breaker heights that delimit surging from plunging waves and plunging from spilling waves can be obtained. Using these limits, a wave breaking probability distribution (e.g., Thornton and Guza 1983) can be integrated to provide breaker type probabilities that may vary across the surf zone.

This surf modeling description and the provided equations are reasonably general. Various formulations for breaker dissipation, wave breaking probability distributions, and breaker types have been used. Given an incident directional wave spectrum, $E(f, \theta)$, where f is wave frequency and θ is wave direction, the numerical approach involves calculating E_w , root-mean-square wave height, appropriate wave frequencies

such as the frequency of maximum energy and the average frequency, and a suitable single wave direction that are contained in the complete equations. Wave direction is usually calculated so that concentration of wave energy in this direction provides the correct longshore momentum flux for later longshore current calculations. The energy transport equation is then solved numerically from outside the surf zone to the shoreline. Numerical integrations are also used for some calculations such as the average wave energy dissipation rate and wave breaking percentages. Fredsøe and Deigaard (1992) and Svendsen and Putrevu (1996) further describe several modeling approaches.

While modeling of only breaker characteristics may be performed in this manner, information about wave-generated longshore currents is often desired. Depth-averaged longshore currents are usually obtained using radiation stress theory (e.g., Longuet-Higgins 1970a; Longuet-Higgins 1970b) or related equations. Doing this involves solving the longshore momentum equation given by

$$\tau_y^h + \frac{\partial}{\partial x} \left(\mu h \frac{\partial v}{\partial x} \right) - \langle \tau_y^b \rangle + \langle \tau_y^w \rangle = 0,$$

where V is the depth averaged longshore current, the first term is the longshore driving stress exerted by the waves, the second term represents horizontal mixing across the surf zone, the third term is the average longshore stress due to bottom friction (also a function of v), and the last term is the average wind-generated longshore stress. Including wind effects in this simple manner adds little to the computations. The parameter, μ , is a horizontal eddy viscosity and h is local water depth. The driving related to wave breaking is given by

$$\tau_y^h = \langle \varepsilon_b \rangle \frac{\sin \theta}{c},$$

where c is wave phase speed.

Dissipation due to wave breaking is a key input for longshore current calculations. Thus, most surf models that calculate breaker information also calculate longshore currents. Breaker and dissipation calculations are first completed across the surf zone. The longshore momentum equation is then solved numerically for depth-averaged longshore currents. Formulation of the bottom stress involves the longshore current, the wave orbital velocities, and a bottom drag coefficient. Formulation of the wind stress involves the wind speed, wind direction relative to the beach, and a sea surface drag coefficient.

Water depths (bathymetry) along a traverse extending from seaward of the surf zone to the shoreline are critical to accurate surf modeling. Solving both the energy transport equation and the longshore momentum equation for one-dimensional models involves the local water depth, h , at each

computational step along the traverse. In the energy transport equation, the wave group velocity and the average rate of dissipation of wave energy depend on water depth. In the longshore momentum equation, the longshore driving stress exerted by the waves and the horizontal mixing term depend on water depth. The bottom friction term also includes depth effects because it is a function of depth dependent wave orbital velocities. Because of the small spatial scales of cross-shore depth variations in surf zones, computational steps (calculation intervals) are typically on the order of 1 m. For two-dimensional or three-dimensional time domain models, there are similar depth requirements.

Lack of recent and accurate depth information at a beach of interest is often a limitation. Recent depth data are desirable because nearshore depths may change rapidly and substantially as a result of wave, current, and tide action. Accurate depth data are needed because wave breaking is strongly related to the water depth.

Equilibrium beach profiles (Dean 1977) that have been used widely in coastal processes and engineering studies provide a method to estimate nearshore depths when actual data are not available. An equilibrium beach profile is given by

$$h(x) = A x^{2/3},$$

where h is water depth, x is distance from the mean sea level position of the shoreline, and A is a scale parameter that depends on sediment size. Equilibrium beach profiles were derived from an analysis of over 500 measured beach profiles and consider that beaches with more coarse sediments are steeper than beaches with finer sediments. An infinite slope theoretically exists at the shoreline, but surf models halt their calculations in extremely shallow water before this behavior causes problems. Also, equilibrium beach profiles do not consider the presence of offshore bars indicating the value of actual depth data when it is available.

Breaker heights and types, as well as surf zone widths, can be modeled fairly well by one-dimensional surf models, but longshore currents are often somewhat in error mainly for beaches with shallow offshore bars. At such beaches, modeled longshore current maxima are generally near the bar, but measured maxima are usually further shoreward. Paradoxically, approximately correct energy dissipation, that provides reasonable breaker characteristics and that drives longshore currents, provides relatively poorer longshore currents. Various improvements have been investigated including linear and nonlinear bottom stress equations, turbulent energy production by breakers, longshore pressure gradients caused by wave setup, and wave rollers representing spilling breakers. Turbulent energy and roller approaches delay driving of currents shoreward of breaker locations. Incorporating variable bottom friction to consider

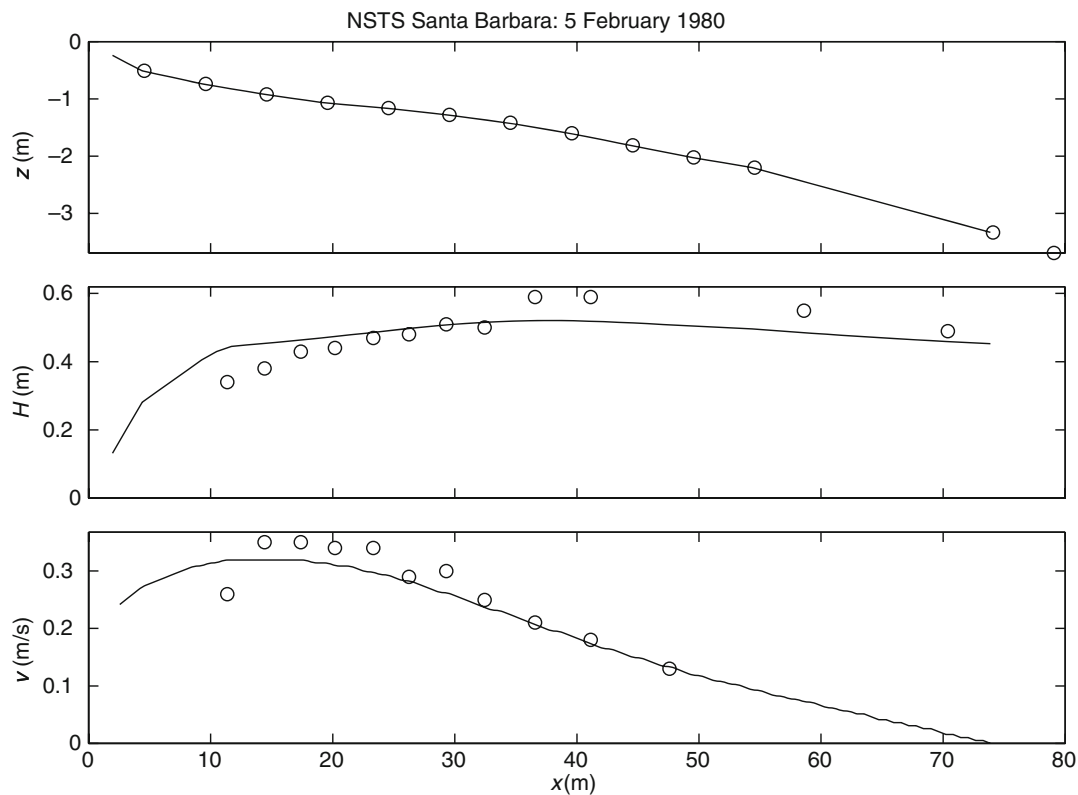
varying sediment sizes across the surf zone and to compensate for missing vertical turbulent effects shows promise. An example output of the US Navy's operational one-dimensional surf model with this modification depicts both breaker height and longshore current variations across the surf zone (Fig. 2).

Because wave breaking is locally depth controlled and one-dimensional models provide suitable wave energy dissipation rates, research is investigating the use of one-dimensional surf models at multiple longshore locations to calculate the dissipation that subsequently drives two-dimensional current models. These more mathematically complicated models then consider longshore variations in both forcing, such as wave setup, and bathymetry (e.g., Putrevu et al. 1995; Svendsen et al. 1998). As earlier noted, time domain solutions of the hydrodynamic equations for wave motion are also being developed.

Modeling Applications

Unlike earlier estimation methods, surf models can provide many types of information across the surf zone. One-dimensional models also can be run at different nearby locations to examine longshore surf variability caused by either incident wave variability or bathymetry changes. Thus, new applications are developing in addition to uses for basic information such as breaker heights. Coastal engineering applications can employ more detailed information such as consideration of different breaker types. Design of systems that operate in the surf zone, and planning of many types of activities, can be better accomplished with quantitative modeled information.

Information about surf is needed for coastal processes studies, such as beach erosion and sediment transport, and engineering analyses, such as design of coastal structures. The potentially adverse effects of surf on activities in the surf zone and the availability of surf models has resulted in increased use of these models to forecast surf conditions. The US Army Corps of Engineers developed a one-dimensional surf model for engineering applications (Larson and Kraus 1991). The US Navy routinely operates a one-dimensional surf model to make surf forecasts for amphibious operations (Earle 1999). Because various military systems require surf information for their designs, this model has also been used to develop surf statistics based on statistics of incident waves (see *Wave Climate* and *Wave Environments*) at locations of interest (Earle 1999). Surf models have also been driven with real-time measured wave data to provide surf estimates without placing instruments in the hazardous surf zone (Nichols and Tungett 1998) and have been linked with other models that provide incident wave conditions (Allard et al. 1999).



Surf Modeling, Fig. 2 Example of a one-dimensional surf model output showing depth (top panel), root-mean-square breaker height (middle panel), and depth-averaged longshore current (bottom panel) across the surf zone. Measured data are marked by circles

Summary

A variety of surf models, primarily one-dimensional models, have been developed and applied. Accuracies of particular models are not reviewed here, but such information can be obtained from the references. Models that have been applied or operated routinely are sufficiently accurate for their intended applications. Research using time domain models shows that these models have significant promise particularly for improved calculations of currents. Most importantly, surf modeling provides quantitative surf information that considers the highly location specific nature of surf.

Cross-References

- [Surf Zone Processes](#)
- [Tides](#)
- [Wave Climate](#)
- [Wave Environments](#)
- [Wave Refraction Diagrams](#)
- [Waves](#)

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Surf Zone Processes

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The surf zone can be defined as that relatively narrow strip of a body of water that borders the land, and which contains waves that are breaking due to the shallow water depth. However, because the tide level, incident waves, and local wind speed, and direction continually change, the width and character of the surf zone vary incessantly. Therefore, in a discussion of surf zone processes, the region of interest is actually the “nearshore” zone, herein defined as that region that is directly or indirectly affected by depth-induced wave breaking. Finally, a subregion called the “swash” zone is commonly delineated at the boundary between land and water, as that area which is alternately wetted and dried by wave uprush and backrush. These zones are indicated in Fig. 1.

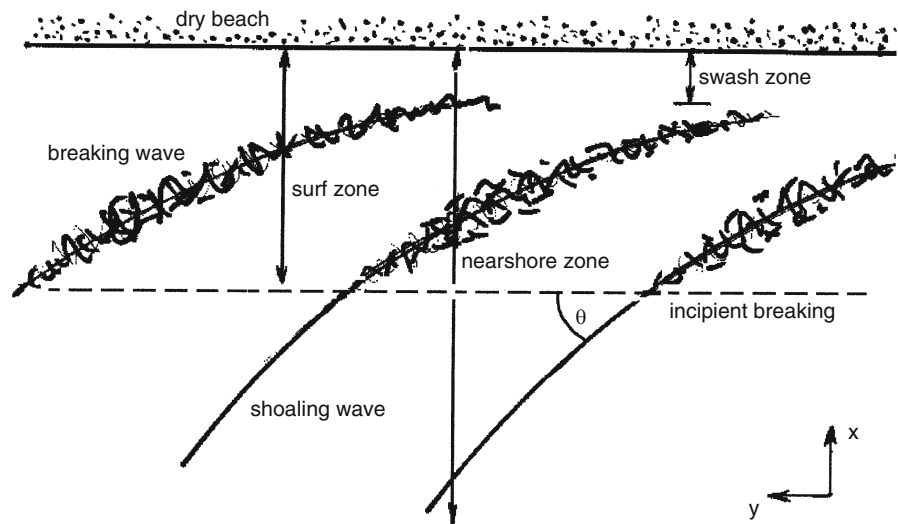
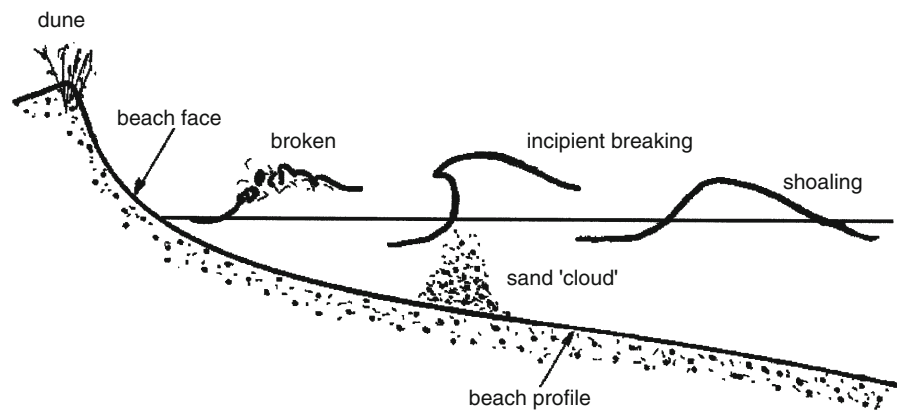
The depth-induced breaking of waves drives a progression of intertwined processes that ultimately leads to changes in the morphology of the beach itself, caused by erosion and accretion of sediment. These processes include the creation of the breaking wave roller, maintenance of residual turbulence in the water column, the creation of setup in the mean water level, generation of currents (cross-shore, longshore, and rip currents) and low-frequency motions, entrainment and suspension of sediment, and finally the transport of sediment.

Wave Transformation in the Nearshore

The phenomena that are most important to wave transformation in the nearshore are (1) refraction, (2) shoaling, and (3) decay due to depth-limited breaking. Refraction (analogous to the refraction of light) is the bending of the waves by variations in bathymetry, and generally tends to align the crests of the waves with the bathymetric contours. Therefore along a straight, smooth coastline, waves that approach from an oblique angle in deepwater may be almost shoreparallel by the onset of breaking in shallow water, as depicted in Fig. 1. In this situation, refraction also serves to suppress the wave height. Along an irregular coastline, however, refraction will tend to focus wave energy on headlands and more subtle protruding bathymetric features, thereby locally increasing the wave height. In order to conserve the total amount of wave energy in the system, the wave height is correspondingly reduced in the embayments adjacent to the protruding feature.

As waves refract from deep to shallow water, the process of shoaling is also at work. In rudimentary explanation, as the waves move into shallow water they slow down, but consequently in order to maintain the total amount of energy flux, the wave energy (proportional to the square of the height) must increase. In the situation of a straight, smooth coastline, it is clear that the amplification of wave height due to shoaling typically exceeds the suppression due to refraction, as an increase in wave height prior to breaking is obvious to an observer on the beach.

When the water depth becomes too shallow to sustain the height of the growing wave, the wave becomes unstable and breaking ensues, as characterized in Fig. 2. This point of incipient breaking is commonly estimated to be when the wave height reaches roughly 80% of the water depth. Incipient breaking is also dependent upon the wave period and the local bottom slope, for which many empirically derived formulas have been offered (e.g., see Weggel 1972). In addition, the wind affects incipient breaking with a following wind causing waves to spill sooner, thereby widening the surf zone. An offshore wind delays breaking, thereby compressing the surf zone, and causes waves to plunge (Douglass 1990).

Surf Zone Processes,**Fig. 1** Definition sketch of the nearshore zone, showing overhead view of wave refraction, shoaling, and breaking**Surf Zone Processes,****Fig. 2** Cross-section of the nearshore zone, depicting wave transformation

As the water depth continues to decrease, breaking becomes fully developed and the wave height continues to decrease. The rate of decay of wave height depends predominantly on the bed slope. For slopes around 1/30, the incipient condition of 80% of the water depth continues to hold true. For milder slopes, however, this value gradually decreases to as little as 35%, whereas for steep beaches the wave height might maintain a value equal to the local depth. More sophisticated models that describe breaking decay are available (e.g., Battjes and Stive 1985; Dally et al. 1985).

Because the beach face is typically steep (Fig. 2), as the waves approach the mean water-level shoreline they collapse to form the characteristic strong up rush and downrush of the swash zone. This is the case especially during high tide. At low tide, however, the water level may intersect the beach at a much flatter segment of the profile, causing the waves to continue their decay as gently spilling breakers. In this situation, the swash zone may be extremely small, or may not exist at all.

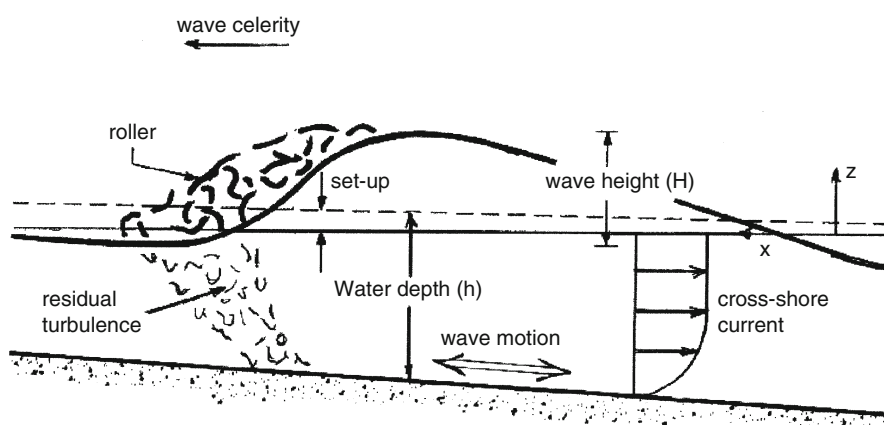
The Breaking Wave Roller

In watching the surf zone the most obvious feature is the “rollers,” which characterize wave breaking. At breaking’s onset, a portion of the energy that was previously transported in the organized wave motion is converted into a region of highly turbulent, aerated water that moves at roughly the wave celerity, as shown in Fig. 3. Energy is dissipated in the form of heat in a shear layer that exists between the roller and the underlying organized flow, as well as inside the roller itself. In addition to its role in the energy budget of the surf zone, the roller is also a significant contributor to the mass and momentum balances (Svendsen 1984).

Due to its highly turbulent, aerated nature, very few direct measurements of fluid motion in the roller have been obtained (see Mocke et al. 2000), and so detailed knowledge is lacking. However, quantitative models for macro-properties of the roller exist, basically providing the evolution of the size of the roller as it crosses the surf zone (Dally and Brown 1995).

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Fig. 3 Hydrodynamic features of the surf zone (after Dally 2000)



Residual Turbulence

As the roller passes by a particular spot in the surf zone, it leaves behind a layer of residual turbulence that expands downwards through the underlying water column as shown in Fig. 3. This turbulence has important links to other surf zone processes, namely the suspension of sediment and the mixing of surf zone currents. The additional “stirring” by the residual turbulence establishes the vertical structure of surf zone currents, with stronger turbulence resulting in currents that are more uniform over depth. Residual turbulence contributes both directly and indirectly to lateral mixing of the currents as well, and is also responsible for maintaining concentrations of suspended sediment in the upper water column that far exceed those under non-breaking conditions.

The behavior of the residual turbulence left in the water column by the passing rollers has been found to mimic that of a free wake created by unidirectional flow past an obstacle. By applying principles from the basic theory for turbulent wakes, the temporal and spatial structure of the residual turbulence has been modeled to some degree (e.g., Sakai et al. 1982; George et al. 1994). The wave-period-averaged turbulence intensity and the vertical mixing in the wake have been quantified in terms of macro-properties of the roller itself (Dally 2000).

Radiation Stress

Although the water particle motion associated with ocean waves is oscillatory, if the momentum flux associated with each wave is mathematically integrated over the water column and then time-averaged over a wave period, one finds that the waves exert a mean residual stress on the water column. This stress, called the “Radiation Stress” (Longuet-Higgins and Stewart 1964), is actually a stress tensor with three components. In a Cartesian coordinate system with x and y directions in the horizontal plane and directed vertically

upwards (as in Figs. 1, 2, and 3), two components, S and S_{yy} , act like normal pressure in the x and y directions, respectively, whereas the third, S_{xy} , acts like a tangential shear stress. In the shallow water of the nearshore, they are each functions of the wave height (H) and local wave direction (θ), and are approximated by

$$S_{xx} \cong \frac{\gamma H^2}{8} \left(\frac{1}{2} + \cos^2 \theta \right), \quad (1a)$$

$$S_{yy} \cong \frac{\gamma H^2}{8} \left(\frac{1}{2} + \sin^2 \theta \right), \quad (1b)$$

$$S_{xy} \cong \frac{\gamma H^2}{8} \cos \theta \sin \theta, \quad (1c)$$

where γ is the density of water. Spatial gradients in these stress components slightly deform the mean water level in the nearshore, as well as drive surf zone currents.

Set-Down and Set-Up

With a coordinate system oriented such that the x -axis is directed onshore, cross-shore gradients in S due to shoaling of waves (i.e., an increasing wave height) actually depress the mean water level seaward of the surf zone. This “set-down” begins as waves enter transitional water and begin to feel the bottom, and gradually draws down the mean water level until reaching the point of incipient breaking. At this point the maximum set-down is observed to be roughly 2–4% of the height of the breaking wave (see Bowen et al. 1968).

As breaking ensues and the roller develops, the mean water level first flattens for a short distance because there is a momentary balance between the decrease in S and the increase in momentum flux in the roller as it grows in size. However, farther into the surf zone as both the wave and roller shrink, a “set-up” in the mean water level is created (see

Fig. 3), which reaches a maximum value at the shoreline. Although dependent on the bottom slope and the wave period, this maximum setup is roughly 12–18% of the breaker height. With a large swash zone, however, the maximum setup is difficult to define and measure, due to the intermittent presence of water.

The local wave-induced set-up is superimposed upon any large-scale deviation in mean water level forced by the winds. For example, offshore-directed winds apply a surface shear stress that tends to suppress the water level in the nearshore, whereas onshore-directed winds will push water up against the shore and hold it there, effectively elevating the mean water level as a “storm surge.”

Cross-Shore Currents

Although the water motion associated with a non-breaking wave is nearly balanced between forward and backward displacement, there exists a residual flux of water in the direction of wave propagation, which occurs mostly in the wave crest. In breaking waves, this flux is augmented by additional onshore flux of water in the aerated roller. Due to the presence of the shoreline, and with longshore-uniform conditions, this mean landward discharge in the upper part of the water column, must be locally compensated by a mean, offshore-directed “return-flow” in the lower water column, depicted in Fig. 3. The return-flow current is often called “undertow,” but is not to be confused with the rip current discussed below. Although the strength of the return-flow is only roughly 10% of the wave celerity, it is this weak, yet persistent current that is primarily responsible for the net offshore transport of sediment that occurs during storms (Dally and Dean 1984; Dally and Brown 1995).

The shear stress applied to the water surface by the wind also contributes forcing that affects the strength and structure of cross-shore current in the surf zone. An onshore-directed stress pushes additional surface water toward the beach which, again because of the presence of the shoreline boundary, must be balanced by an increased return-flow current in the lower water column. Consequently the onshore wind that often accompanies a storm further augments the wave-driven offshore current. With an offshore-directed wind the opposite occurs. That is (neglecting the wave-induced current for the moment), surface water pushed offshore by the wind stress is supplied by water drawn onshore in the lower water column.

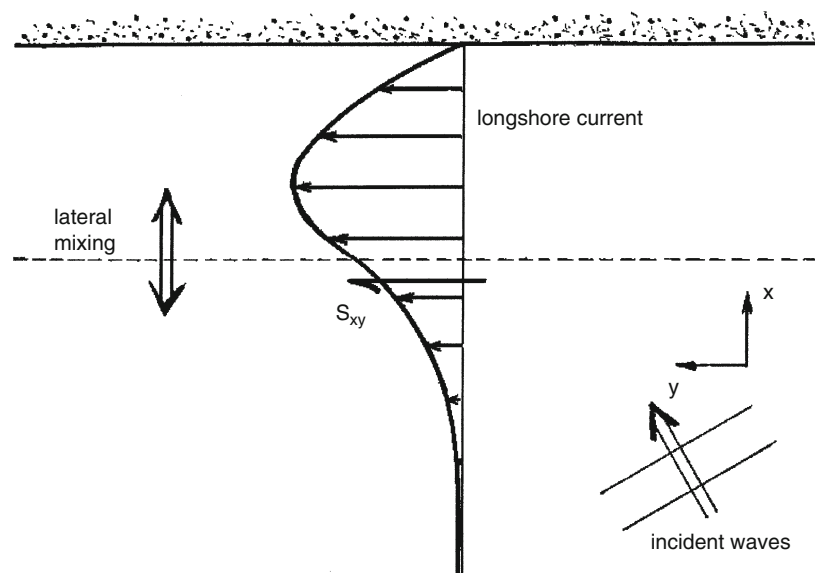
Longshore Currents

The mean flux of water along the beach is called the “longshore” current, and is also forced by both waves and local winds. For obliquely incident waves, the cross-shore gradient in S_{xy} (Eq. 1c) due to breaking creates a residual shearing thrust that is directed down the beach as depicted in Fig. 4. In the outer surf zone this thrust is locally offset by an opposite-directed shear associated with the creation of the roller. However, as the roller subsequently starts to shrink, its thrust switches direction, and both thrusts now act in harmony to drive water down the beach (Osiecki and Dally 1996). Consequently, the location where the longshore current is greatest is usually somewhat landward of the break point. The mean longshore thrust of the waves is balanced by the shear stress that the sand bed exerts on the water.

Seaward of breaking, there is no wave-induced forcing of the longshore current because the increase in S_{xy} associated with wave shoaling is exactly compensated by the decrease

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Fig. 4 The longshore current, driven by the S_{xy} component of Radiation Stress for obliquely incident waves. Lateral mixing creates the “tail” of the current



due to wave refraction. However, lateral mixing processes driven by both turbulence and residual convective acceleration enable the current inside the surf zone to drag the outside water along with it, creating a tail on the longshore current that diminishes in the offshore direction (Svendsen and Putrevu 1994). Outside the breaker line, this lateral mixing stress is balanced by the mean bed stress of the current.

Depending upon its speed and direction, the wind can also be a significant contributor to forcing of the longshore current. In fact, it is often the primary driver during the initial stage of development of a storm. The alongshore component of the wind stress vector, acting on the water surface, can either act in harmony with the wave-driven forcing, or oppose it. Although a rare event, strong opposing winds can even arrest the wave-driven longshore current.

At the base of the beach face, the behavior of the longshore current is controlled by the character of the swash zone. For very flat profiles with little swash, the current speed approaches zero as depicted in Fig. 4. For large swash zones though, the mean longshore current observed at the toe of the swash can be significant.

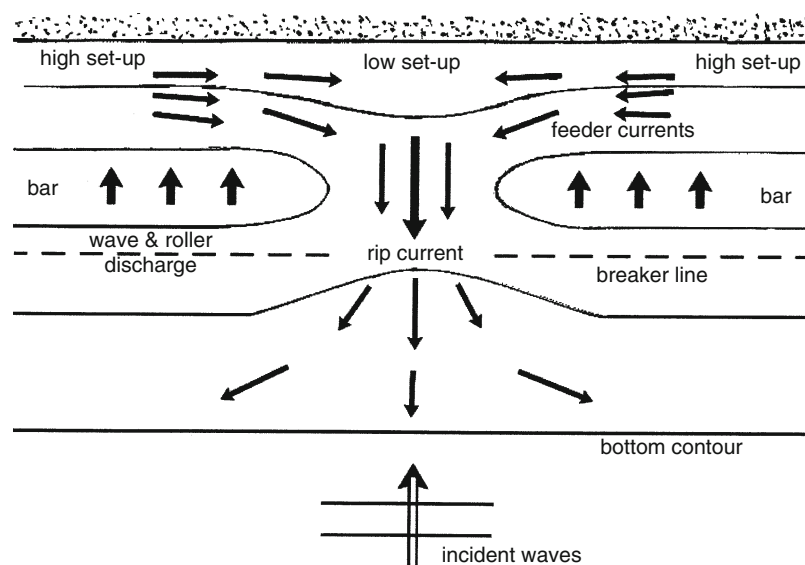
Rip Currents and Nearshore Circulation

In situations where either the incident wave field or the underlying bathymetry is not uniform along the beach, complicated circulation patterns can be established. These include the infamous “rip current” that is so dangerous to bathers. Not to be confused with the mild cross-shore current (undertow) described above, rip currents are very strong, yet locally confined, offshore-directed flows that can reach speeds more

than 1.0 m/s, making them nearly impossible to swim against. One likely generation mechanism for a rip current is the presence of a low spot in a longshore bar, through which the majority of water carried landward over the bar by the waves along the adjacent beaches is returned. As shown in Fig. 5, the rip current is fed by the mild, mean onshore transport of water that occurs in the wave crest and roller as discussed above, which accumulates to form longshore-directed feeder currents that search out the narrow, low section in the bathymetry to form the seaward-directed rip. Setup is reduced in the rip channel but elevated on either side, which serves to drive the circulation cell. If the bathymetry is rhythmic in the longshore direction, rip currents can form at regularly spaced intervals, and can be present as long as the special bathymetry persists – sometimes even for several days (see Noda 1972). Another possible origin of regularly spaced rip currents is the interaction of waves approaching a beach from different directions (Dalrymple 1975). The waves tend to enhance each other at regular intervals along the beach (antinodal zones), with zones of suppression in between (nodal zones). Where the waves reinforce, the setup is greater than where they negate. This results in structure in the mean water surface that drives water from the antinodes towards the nodes, which is again turned out to sea in the form of a rip. Additional complex nearshore circulation patterns are often generated adjacent to coastal structures such as groins, jetties, piers, and breakwaters. Basically, the structure perturbs what might otherwise be a longshore-uniform wave climate, creating a region of reduced setup in its wave shadow, to which water is driven from remote regions of higher setup. The shadow can be the obvious result of wave blocking by the structure, as well as more subtle energy losses incurred due to “rubbing” of the waves against the structure.

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Fig. 5 Rip currents generated at a low spot in the crest of a longshore bar



Low-Frequency Motions

Often the incident waves arrive at the beach in distinct groups, especially if they were generated by a distant storm and local wind conditions are calm. This groupiness manifests itself as a low-frequency modulation of the wave height, where the wave height grows over several successive waves and then decreases over several more. Each group can last up to several minutes, sometimes with a notable period of calm in-between groups. The setup phenomenon described above responds directly to this rhythmic unsteadiness in wave height, and the gradually varying mean water level that results is referred to as “surf beat.” Surf beat can become particularly pronounced if the bottom slope is slight and the waves are large (i.e., a wide surf zone).

Because surf beat is a low-frequency motion ($f < 0.01$ Hz), some of this energy can be trapped against the coastline by refraction processes or can be amplified, particularly by reflection from headlands and shore-perpendicular structures. These long-period, “edge” waves actually propagate along the beach and, although measurable, are usually undetected by the naked eye (e.g., see Guza and Davis 1974).

Groupiness in the waves also causes the mean forcing in the nearshore to vary in time, and consequently the cross-shore and longshore currents are unsteady. One additional low-frequency motion sometimes found in the surf zone is the “shear wave,” which is actually a periodic instability generated by strong longshore currents. This phenomenon is detectable only with a current meter, as it does not directly influence the mean water-level elevation (Oltman-Shay et al. 1989).

Sand Entrainment and Suspension

Although researchers continue to struggle with the extremely complicated problem of quantitatively modeling sediment entrainment and suspension in the surf zone, much is known qualitatively. In general terms, sediment is first entrained by the near-bed fluid motion, and then carried up into the water column by small-scale turbulence or largerscale vortices. The most germane parameters are the size and gradation of the sediment (typically sand), its settling velocity in still water, and the strength and character of the water motion, which is usually quantified in terms of wave height and period, and current strength. Small sand grains with slow settling velocities are easily entrained and suspended in large quantities, whereas large, heavy particles understandably hug the bed. Larger waves suspend more sediment due to their stronger oscillatory velocities, whereas short-period waves support suspension by maintaining higher mean turbulence levels.

Outside the surf zone, the entraining water motion is comprised of oscillatory wave motion, mean currents, and low-frequency motions. Once entrained, sediment is mixed/suspended higher into the water column by the turbulence generated at the bed as the waves and currents interact with bedforms such as ripples and mega-ripples.

Inside the surf zone, entrainment is enhanced by stronger wave motion, particularly the more pronounced flow that occurs under the crest of nonlinear, breaking waves. Suspension into the upper water column is greatly augmented by the residual turbulence left by the roller. Finally, the extreme flow conditions and large vortices generated by plunging breakers result in dramatic entrainment and suspension events.

At any particular location in the nearshore, the vast majority of entrainment and suspension occurs as the wave crest passes, creating a “cloud” of sediment (depicted in Fig. 2). Outside the surf zone this cloud may be quite small, both in size and in concentration, especially if the entrainment is associated with sand ripples that are present on the bed. Near the break point, large, long-period waves can create what is known as “sheet-flow” conditions, in which the bed is scraped flat and, although there is strong entrainment, there is limited suspension due to a lack of turbulence. In both of these situations, the cloud of sand settles back to the bed before the next wave crest arrives. However, in the surf zone, and particularly for plunging breakers, the sediment cloud may take several wave periods to settle back to the bed, and may in fact be resuspended by a subsequent wave. It is for this reason that sediment concentrations inside the surf zone are generally several times greater than those outside (e.g., see Kana 1979; Nielsen 1992).

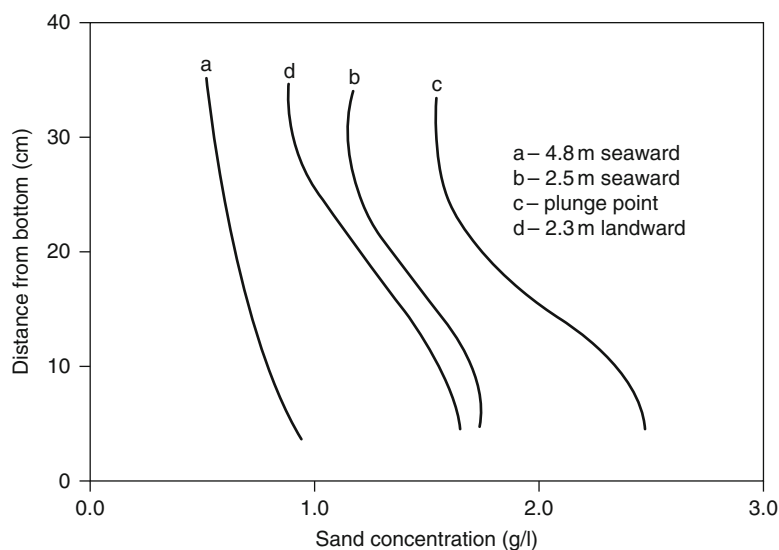
Farther into the surf zone, because the wave energy decreases due to breaking and the oscillatory motion abates, ripples will once again appear if the surf zone currents are weak. If the currents are strong, “mega-ripples” may be created, resembling those found in rivers. Finally, in the swash zone where wave collapse and runup occurs, sheet flow conditions return and the bed is smooth. In this region, sand entrainment and suspension is characterized as a slurry, rather than as a cloud.

Due to the randomness of the incoming waves, the creation of the sediment clouds is an intermittent process. In fact even in laboratory wave channels where so-called “regular” waves can be created, the details of the entrainment and suspension events vary from wave to wave because the underlying hydrodynamic processes are highly nonlinear. Consequently, time-series concentration measurements (typically made using either sonic or infrared instruments) are often averaged over many wave periods so that the basic structure can be identified.

To illuminate the concepts introduced above, Fig. 6 presents idealizations of time-averaged measurements of

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Fig. 6 Idealized profiles of mean concentration of sand suspended by plunging breakers. Concentrations at the plunge point can be several times greater than elsewhere



suspended sand concentration observed in the outer surf zone in a large wave channel for regular, plunging breakers (see Barkaszi and Dally 1992). Note that both the shape of the vertical profile and the overall magnitude of concentration change considerably as one moves from outside the break point (location “a”), to the plunge point (location “c”) and into the surf zone (location “d”). However, as is typically the case the mean concentration decreases with elevation above the bed, as intuition would dictate.

Sediment Transport and Beach Evolution

In the discussion of sand transport in the surf zone, and ultimately the changes in the beach that result from spatial gradients in transport, the water column can be split into two layers: (1) a “near-bed” layer, in which both the oscillatory wave motion and the mean current contribute to the displacement of sediment particles, and (2) a “fully suspended” layer where, effectively, only the mean current contributes to the net displacement (Dally and Dean 1984). The thickness of the near-bed layer, d_f , is simply the distance that a sand grain falls in one wave period, given by $d_f = T \cdot \omega$, where T is the wave period and ω is the fall velocity in still-water. Of course on a natural beach, the incident waves have differing periods and the sand grains are not uniform in fall velocity, so the boundary between the near-bed and fully suspended regimes is not distinct. Also, when the water depth is less than the value of d_f , the nearbed regime extends over the entire water column, as is the case in the swash zone, for example.

The utility in defining the two regimes is that in the fully suspended layer, the average sediment transport rate is well approximated simply by multiplying the suspended sand concentration by the local current velocity. However, in the nearbed layer, because the sediment is in suspension for less

than one wave period, the wave-induced oscillatory velocity does contribute to the net displacement, on average (Dean 1973). Neglecting the mean current for the moment, and with suspension triggered as the wave crest passes, the sand in the lower half of the near-bed layer will experience a net shoreward displacement, whereas that in the upper half experiences a net seaward displacement, as shown in Fig. 7. Because generally the concentration in a sand cloud decreases with elevation above the bed, it is apparent that in the nearbed layer the total net transport of sand is directed onshore. Reintroducing the mean current modifies this scenario of course, depending on its strength and direction.

Cross-Shore Transport and Beach Profile Evolution

With this two-layer framework, the details of the sand transport processes that cause beach erosion and accretion can be described. At times when waves directly approach the beach ($\theta = 0$), the mean current in the surf zone and nearshore is comprised of only the offshore-directed undertow. Consequently, mean sediment transport in the fully suspended layer is directed solely offshore. If relatively little sand is being suspended above the nearbed layer, however, the total net mean transport will be onshore, and the beach face will accrete as it receives sand moved from the nearshore. When a storm develops, however, (1) the wave period shortens, thereby reducing the thickness of the nearbed layer, and (2) the residual turbulence levels increase, thereby carrying greater quantities of sediment into the fully suspended layer. Both of these characteristics enhance the offshore-directed transport to the point where it dominates the diminishing onshore transport, and the beach and inner surf zone consequently erode. The material is deposited in a longshore bar that develops near the point of incipient breaking where the entrainment processes weaken. When calmer conditions return and the mild onshore/nearbed transport once again

Surf Zone Processes,

Fig. 7 Sediment transport regions in the nearshore. Sand suspended in the near-bed layer returns to the bottom within one wave period, whereas that in the fully suspended layer does not (after Dean 1973; Dally and Dean 1984)

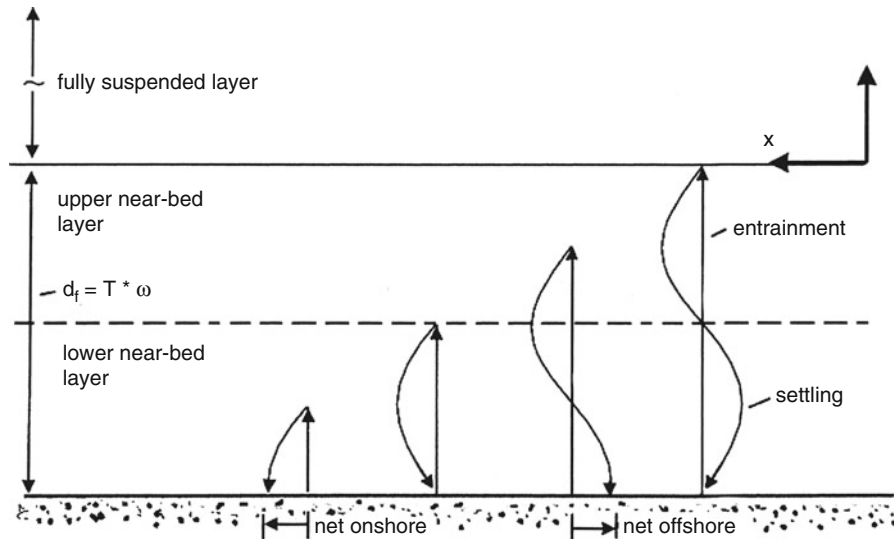
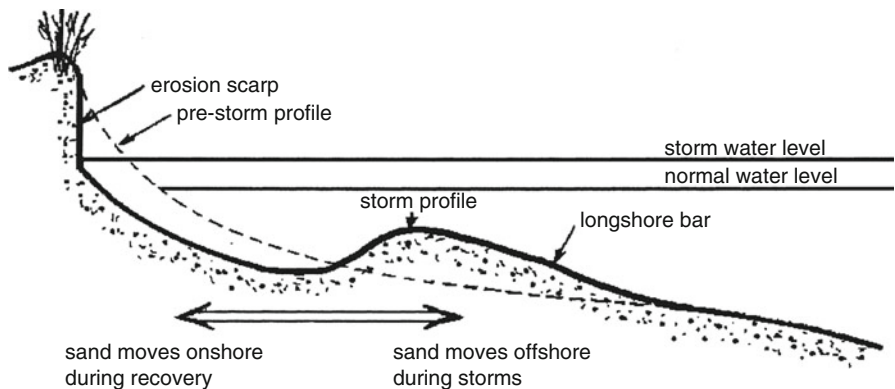
**Surf Zone Processes,**

Fig. 8 The cycle of storm-induced beach erosion (which creates a longshore bar), and subsequent recovery (which pushes the bar material back onshore)



dominates, this bar will gradually migrate onshore and supply sand to the swash zone, thereby rebuilding the beach face. This well-known sequence of beach erosion and recovery is illustrated in Fig. 8.

From the discussion above it follows that if wave and water level conditions could be held constant, the beach would evolve until a state of “dynamic equilibrium” was attained, in which the onshore transport in the lower portion of the nearbed layer was balanced by the offshore transport in the upper portion and the fully suspended layer – a balance that must be maintained everywhere across the beach profile. Such a state is easily attained in a laboratory wave channel, but seldom approached in nature.

Longshore Transport and Coastline Change

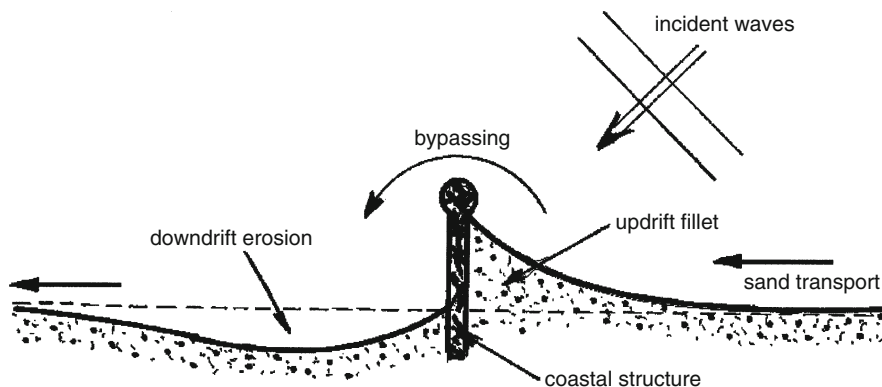
If waves approach the surf zone obliquely, both the longshore component of the oscillatory wave motion and the longshore current serve to carry sand along the beach. The flux of sand in the longshore direction in the fully suspended layer is again well approximated by the product of the mean concentration and the longshore current speed. In the nearbed

layer, although sand initially suspended in the top of the layer may experience net upcoast displacement due to the oscillatory motion, the mean flux for the entire layer is downcoast, and is further enhanced by the longshore current. In the swash zone the uprush can occur at a strongly oblique angle, whereas the downrush typically does not. Consequently, the net longshore transport in the swash can be significant, and in fact is often the predominant zone of longshore transport in the nearshore (see Bodge and Dean 1987).

If there is no longshore variation in the flux of sediment along the beach, then the beach profile and coastline remain stable. However, if the longshore transport is blocked by a natural immobile feature (e.g., a headland or reef/rock outcrop) or by a man-made structure (e.g., a groin or jetty), sand will accumulate on the updrift side and erode from the down-drift, as shown in Fig. 9. Also, on a seemingly unperturbed beach, longshore variations in offshore bathymetry cause variations in the nearshore wave climate, and consequently nonuniformity in the transport along the beach. This can result in erosion “hot spots,” as well as otherwise mysterious zones of accretion. Finally, longshore variations in sand

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Fig. 9 Shoreline change due to interruption of longshore transport of sand by a structure. The updrift fillet grows and the downdrift pocket erodes until the fillet is saturated and bypassing begins



characteristics can lead to longshore gradients in transport and ultimately to the creation of a meandering coastline.

Additional suggested readings on this subject may be found in Battjes (1975), Dean and Dalrymple (1984), Fredsoe and Deigaard (1992), Komar (1998), Nadaoka and Kondoh (1982), and Whitford and Thornton (1996).

Cross-References

- Bars
- Beach Erosion
- Beach Processes
- Coastal Currents
- Cross-Shore Sediment Transport
- Erosion Processes
- Longshore Sediment Transport
- Surf Modeling
- Wave Refraction Diagrams
- Waves

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Surfing

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The origins of surfing are believed to date back to the ancient Polynesians, well before Captain James Cook became the first westerner to document surfboard riding, ca. 1777 (see Lueras 1984). Simply hand-carved planks of wood, the first surfboards were large (up to 5 m in length), heavy, and difficult to control. Subsequent evolution during the 1920s to 1950s included hollowing the board, as well as shaping from balsa. In the early 1960s, the advent of synthetic foam and fiberglass ushered in the modern era of surfing, and forever changed the nature of the sport by making surfboards shorter, lighter, faster, and more maneuverable. The modern surfboard is typically made of hightech, composite materials with high strength-to-weight ratios. A custom-made board starts as a “blank,” consisting of a foam core with one or more wooden stringers running its length to provide strength. After cutting to rough dimensions, the blank is shaped by hand, and the finished core is then covered with fiber-glass, using either polyester or epoxy resin. One or more fins, known as “skegs,” are incorporated near the tail of the board to provide control and to enhance stability. Although a limited amount of laboratory and computer-simulation modeling of surfboards has been attempted (e.g., see Hornung and Killen 1976), the evolution of board shape and design has been mostly by trial-and-error. Generally, long boards are used in gently spilling breakers, and shorter boards used in steep-faced, plunging waves.

“Stand-up” surfing is the most prevalent type of surfing, although body-boarding, where the surfer lies prone on a much shorter board and uses flippers to aid in wave catching, has gained significant popularity. Knee-boarding is a more rare form. Finally, in body-surfing no board is needed at all, with the surfer’s own body used as a planing surface. This is arguably the “purest” form of surfing, practiced even by sea lions and bottlenose dolphins in the wild.

Catching a Wave

In catching a wave, the surfer first takes up a position near the seaward edge of the surf zone, that is, at the break point of the largest incoming waves. As a suitable wave approaches, the surfer pivots to face towards shore and begins to paddle in order to gain forward momentum. With the face of the wave steepening, the surfer is lifted and the board tips forward, consequently accelerating in the direction of wave motion. The surfer then jumps to a semi-crouched position and slides down the wave face, thereby momentarily matching the speed of the wave. After “dropping in,” the surfer executes a “bottom turn” in order to establish a path more tangential to the wave. From that moment on the surfer tries to maintain a position somewhere in the area immediately in front of the critical breaking region (the “pocket”), and so the ride is essentially a delicate reconciliation between the speed of the point of incipient breaking (the “break rate”), and the speed of the surfer (the “board speed”). If the break rate becomes greater than the maximum board speed that can be sustained by the surfer, the break point overtakes the surfer and the wave “closes out.” If the break rate is less than the maximum sustainable board speed, the surfer has time to “carve,” that is, maneuver up-and-down the face of the wave, as well as attempt other acrobatics. Maximum sustainable board speeds have been documented in excess of 18 m/s (40 mph) (Dally 2001a).

Analysis of Natural Surfing Breaks

What makes a particular site good for surfing? One of the first comprehensive studies of basic surfing mechanics, as well as the analysis of specific surfing breaks, was conducted by Walker (1974) for several famous breaks in Hawaii. Bathymetric surveys were performed, and estimates of surfer speed and “peel angle” (viewed from overhead, the angle between the wave crest and the path of the break point) were collected. The reason for these enduring, high-quality breaks was found to be the interplay of certain wave conditions – for example swell waves of sufficient height arriving from a specific direction – with the underlying reef bathymetry via the processes of wave refraction and shoaling. Recent surveys of other well-known surfing breaks around the Pacific Ocean confirms this, and detailed wave modeling demonstrates the importance of the interaction of bathymetric features of different scales in creating exceptional surfing breaks (Mead and Black 2001a, b). In addition to bathymetric features, the interaction of waves and coastal navigation structures such as jetties has also created some well-known surfing breaks (Dally 1990) such as at Sebastian Inlet, Florida (Fig. 1). Finally, stochastic modeling has been utilized in attempts to quantify the “surfability” of a site, that is, the proportion of time in which good surfing waves are available (Dally 1990, 2001b).

Surfing, Fig. 1 Surfer enjoying “First Peak” at Sebastian Inlet, Florida, which is a perennial break created by wave reflection from the northern jetty. (Photo by Gibber, courtesy of Eastern Surf Magazine)



Engineering of Man-Made Surfing Breaks

For beaches that do not have the benefit of naturally induced, highquality surfing breaks, the long-contemplated idea of enhancing waves for surfing using man-made structures has only recently been undertaken both in Australia (Pattiaratchi 1997; Black et al. 2001) and the United States. These artificial surfing breaks are submerged, moundlike structures, designed to both focus wave energy, thereby increasing the breaker height, as well as reduce the break rate by adopting an orientation that is strongly oblique to the crests of the incoming waves. Such reefs have been constructed out of concrete rubble units, as well as sand-filled geosynthetic bags.

Cross-References

- [Geotextile Applications](#)
- [Lifesaving and Beach Safety](#)
- [Rating Beaches](#)
- [Surf Zone Processes](#)
- [Wave Refraction Diagrams](#)
- [Waves](#)

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Swash Zone Dynamics

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Definition

The swash zone is loosely defined as the region in which the beach face is intermittently exposed to the atmosphere,

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possibly over both long (minutes) and short (seconds) time scales (Elfrink and Baldock 2002). Alternatively, the local swash zone can be defined as the region of wave run-up and run-down around the mean water level, i.e., excluding setup (Holman 1986).

Introduction

The swash zone forms the land-ocean boundary at the landward edge of the surf zone, where waves runup the beach face (Figs. 1 and 2). It is perhaps the region of the ocean most actively used by recreational beach users and, being very visible, is the region of the littoral zone most associated with beach erosion and the impacts of climate change. The landward edge of the swash zone is highly variable in terms of geomorphology, and may terminate in dunes, cliffs, marshes, ephemeral estuaries, and a wide variety of sand, gravel, rock, or coral barriers. This influences the exchange of sediment between the land and ocean, which ultimately forms the coastline.

In terms of coastal processes and coastal protection, a large part of the littoral sediment transport occurs in the swash zone, both cross-shore and longshore, which influences beach morphology, and beach erosion and beach recovery during and after storms. Wave runup is an important factor in the design of coastal protection and also generates hazards for beach users, and is the dominant process leading to the erosion of coastal dunes. Swash hydrodynamics also influence the ecology of the intertidal zone and groundwater levels in sub-aerial littoral beaches and low lying islands, which is often critical for freshwater water supply on islands and atolls (Nielsen 1999).

Characteristics of the Swash Zone

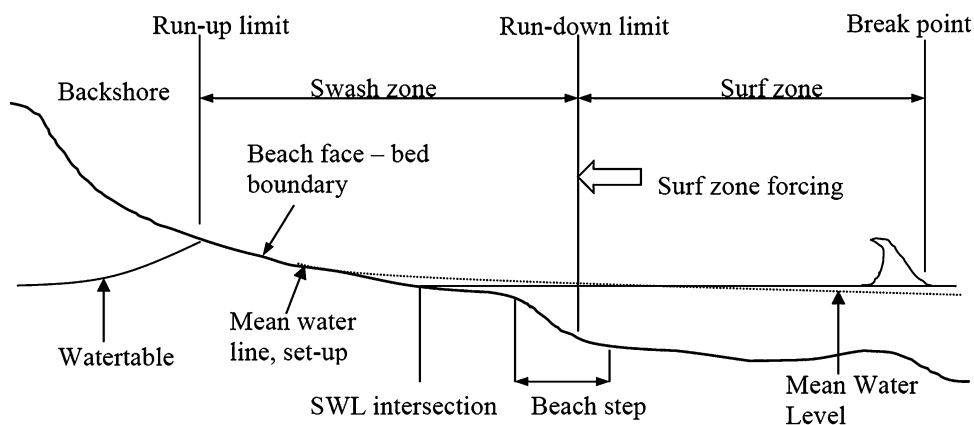
The hydrodynamic processes in the swash zone are characterized by very different types of fluid flow compared to the

open ocean and surf zone, as illustrated in Fig. 2, where the strongly orbital motion of waves is transformed into flow parallel to the bed, usually in thin shallow sheets. In terms of fluid mechanics, the key difference is the occurrence of supercritical flow in the swash zone, where the Froude number, $\frac{V}{\sqrt{gd}}$, is greater than 1, which has important implications for the nature of the flow. Other important distinctions are that friction becomes more important in controlling aspects of the shallow flow in the swash zone than in the surf zone, and that turbulence and sediment transport in the swash zone is generated locally in the swash zone and advected into the swash zone from the surf zone. A key feature of both the hydrodynamics and sediment dynamics in the swash zone is intermittency, where the extent and degree of inundation of the swash zone varies over different timescales, from orders of seconds and tidal periods to years and decades. This is a challenge for coastal scientists, both in terms of measurement and modeling of the physical processes. For the purposes of this summary, the swash zone will be considered as the region of the beach face exposed to the atmosphere over wind, swell and infragravity wave durations, i.e., seconds to a few minutes.

The characteristics of the swash zone hydrodynamics and sediment transport are governed by the inner surf zone and the underlying beach, with feedback of course between the morphology and hydrodynamic processes. The beach slope is a controlling parameter (Guza et al. 1984). On dissipative beaches, with wide surf zones, most of the wind wave and swell energy is dissipated seaward of the swash zone. Therefore, swash processes are dominated by those due to long, or infragravity, waves, which are frequently nonbreaking standing waves (Fig. 2a). On intermediate and reflective beaches, short wave energy reaches the beach face in the form of bores or shore-breaks, which collapse at the beach, initiating a runup motion characterized by a thin sheet of water with a rapidly propagating wave tip which is analogous to a dam-break flow over a dry bed (Fig. 2b). This sheet of water is slowed by gravity and friction until the flow reverses and forms another shallow flow seaward, the backwash.

Swash Zone Dynamics,

Fig. 1 Definition sketch for the nearshore littoral zone (swash zone width exaggerated) (After Elfrink and Baldock (2002))



Swash Zone Dynamics, Fig. 2

(a) Seven Mile Beach, NSW, Australia, a dissipative beach. Photo shows conditions after a swash rundown, with only small bores reaching the swash zone (Photo: Dr. Hannah Power, University of Newcastle, NSW, Australia). (b) Avoca Beach, NSW, Australia, a reflective beach. Photo shows the inner surf zone and a bore reaching the swash zone in the background and a swash uprush reaching the top of a beach berm in the foreground



On coarse grained sand and gravel beaches, a significant volume of the uprush and some of the backwash may percolate into the beach, reducing the volume of water in the surface backwash flow. These two distinct types of swash zone make modeling hydrodynamic processes difficult, since parametric models rely on similarity of processes,

and therefore phase resolving models of the whole surf zone, or at least the inner surf zone, are required if the details of the hydrodynamics are required. Fortunately, some processes are modeled very well by parametric models, perhaps more accurately than phase-resolving models, particularly wave runup.

Wave Runup and Overtopping

Wave runup is perhaps the most important aspect of swash zone flows. While the motion of the water volume as a whole may be considered as runup, conventionally wave runup refers to the landward limit of the swash motion on the beach face, usually defined vertically above the ocean level. The runup and rundown of the shoreline is referred to as the swash excursion or oscillation. A typical plot of the shoreline motion is shown in Fig. 3.

Remarkably, the maximum runup is still most reliably described by a simple empirical parametric formula, which is perhaps one of the oldest regularly in use, proposed by Hunt (1959). Despite numerous variations, the underlying scaling still holds over a very large range of wave conditions, both in the laboratory and field.

The scaling for runup proposed by (Hunt 1959) is:

$$R_{max} = \xi H_0 = \tan \beta \sqrt{H_0 L_0} \quad (1)$$

where ξ is the surf similarity parameter, or Iribarren number ($\xi = \frac{\tan \beta}{\sqrt{H_0/L_0}}$) and H_0 , L_0 , and β are offshore wave height, wave length, and the swash zone beach face slope, respectively. This is strictly only applicable for monochromatic waves and the maximum runup. However, the same formulation has been widely used to describe random waverunup, using appropriate statistical parameters to describe the wave

conditions. This parameter is usually the runup elevation exceeded by 2% of the waves, $R_{2\%}$. The most widely used derivative of Hunt's formula is perhaps due to Stockdon et al. (2006), which also allows for a contribution from infragravity waves:

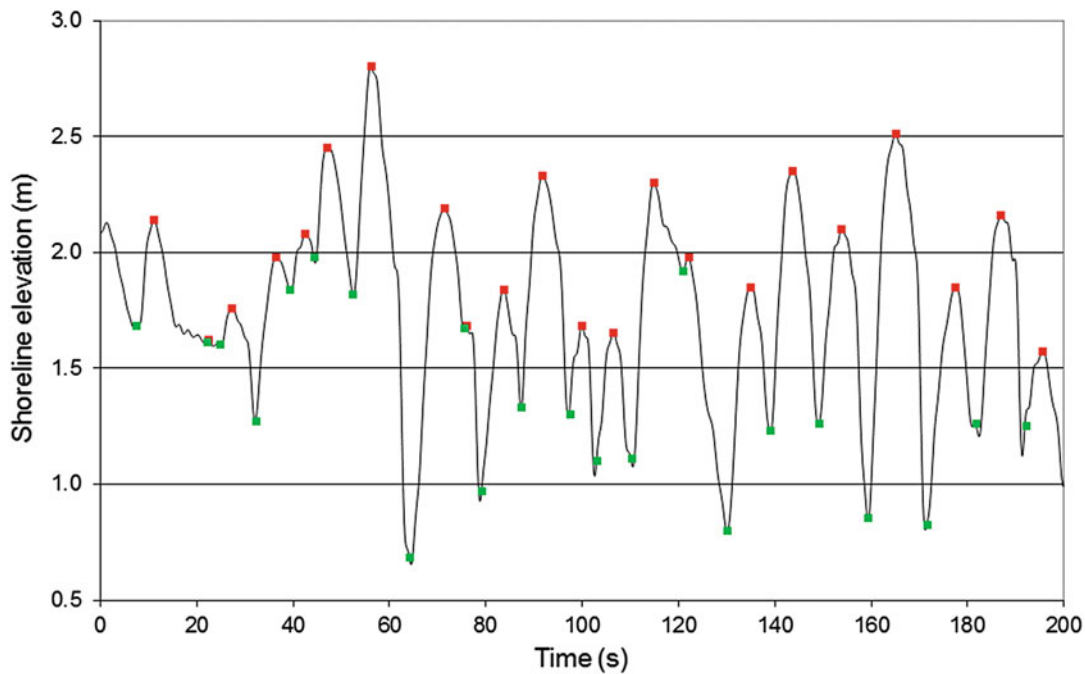
$$R_{2\%} = 1.1 \left(\frac{\sqrt{H_s L_p (0.563 \beta^2 + 0.004)}}{2} + 0.35 \beta \sqrt{H_s L_p} \right) \quad \xi_p \geq 0.3 \quad (2)$$

where the term $0.35 \beta \sqrt{H_s L_p}$ represents the setup, or

$$R_2 = 0.043 \sqrt{H_s L_p} \quad \xi_p < 0.3 \quad (3)$$

and H_s is the deep water significant wave height, and L_p is the wavelength corresponding to the deep water peak wave period, T_p . Atkinson et al. (2017) compared a number of recent runup formulations to measurements from a range of beaches and concluded that the models generally predict runup with errors of order $\pm 25\%$.

Swash-swash interactions occur through the overtaking of a swash uprush by the following bore or during the collision of the backwash flow with the next uprush. The magnitude, or vertical excursion, of the swash oscillations, from rundown position to runup, is strongly influenced by interaction between wave uprush and backwash, with the period of the incident waves also controlling the period of the swash



Swash Zone Dynamics, Fig. 3 Typical pattern of bore-driven swash oscillations. Data show shoreline elevation versus time at Avoca Beach, NSW. Red and green squares indicate maxima and minima of individual

swash events (Data courtesy of Dr. Michael Hughes, NSW Office of Environment and Heritage)

oscillations at swell and wind wave frequencies (Holman 1986). Hence, given a finite time for the uprush and backwash to occur, there is a finite magnitude for a swash oscillation at a given frequency on a given beach slope if the motion is solely controlled by gravity. This leads to swash saturation, where an increase in incident wave height does not increase the magnitude of the swash oscillations. This can be parameterized for individual events, or through a spectral representation.

For nonbreaking monochromatic waves, this limit is given by (Miché 1951):

$$\varepsilon_s = \frac{a_s \omega^2}{g \beta^2} = 1 \quad (4)$$

where a_s is the vertical amplitude of the shoreline motion, ω is the angular wave frequency ($2\pi f$ – where f is the wave frequency), g the gravitational acceleration, and β the beach slope. This assumes a saturated surf zone and, based on the limiting amplitude for monochromatic unbroken standing waves, $\varepsilon_s \approx 1$. For swash initiated by breaking wave bores, Baldock and Holmes (1999) derived a theoretical value for $\varepsilon_s \approx 2.5$, where saturation is controlled by swash-swash interaction as opposed to surf zone saturation. Spectra of the shoreline oscillations also indicate saturation at higher frequencies, with a typical roll-off in the energy density that is proportional to f^{-4} , which is also evident from Eq. 4. Huntley et al. (1977) proposed a uniform spectral form for saturated swash spectra, but variations occur due to different surf zone conditions (Fig. 4).

When the runup exceeds the elevation of the crest of a structure, beach berm or dunes, wave overtopping or wave overwash occurs. This process is very important in building

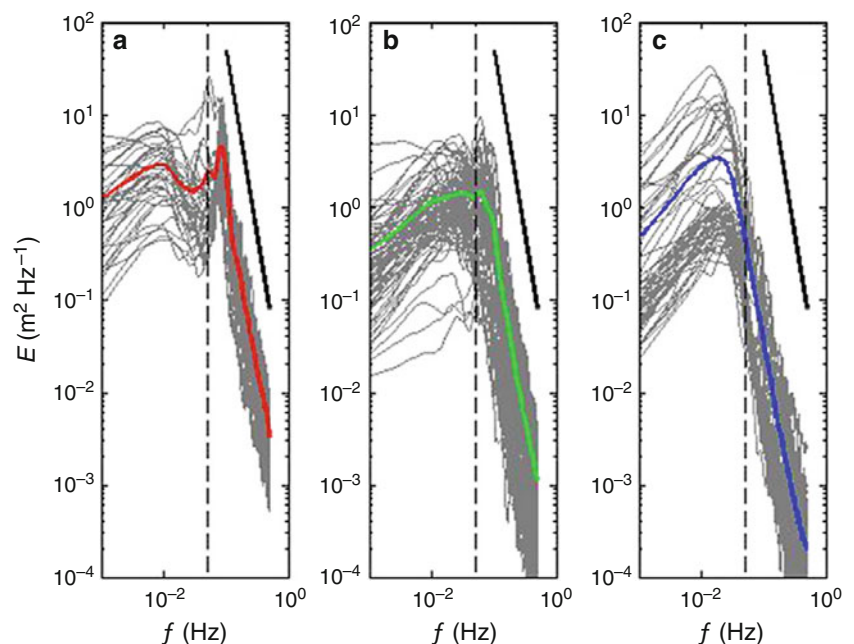
beach berms higher, in association with the spring-neap tidal cycle, but also leads to coastal flooding and inundation of the backshore region. The geomorphology of barrier islands and gravel barriers is strongly dependent on swash overtopping, and breaching of these systems by landward transport of sediment during the overtopping can lead to rapid and potentially catastrophic failure of protective coastal barriers (Fig. 5). The response of coastlines to sea level rise is also be influenced by swash overtopping and sediment over wash, which increases recession in comparison to the classical Bruun Rule (Rosati et al. 2013). A combination of parametric modeling and numerical techniques is required to model these scenario (Roelvink et al. 2009).

Swash Zone Hydrodynamics

Swash zone flows have several features that differ from those in the surf zone, but this is dependent on the dominant wave conditions in the inner surf zone as noted above. Many key characteristics again depend on the Iribarren number (Roos and Battjes 1977). For infragravity standing long waves, the variation in flow depth and flow velocity at a point is relatively symmetrical during uprush and backwash. Short wave bores generate more asymmetrical flows, with the maximum velocity occurring at the start of inundation, with a rapid rise to the maximum depth, with an almost linear deceleration to flow reversal, and a correspondingly similar uniform acceleration in the backwash, at least until the flow becomes very shallow, when friction retards the flow significantly. A key aspect of these flows is diverging flow, which means the swash lens thins rapidly.

Swash Zone Dynamics,

Fig. 4 Swash energy spectra from (a) reflective, (b) intermediate, and (c) dissipative beach-states. The gray lines show individual spectra, the colored lines show the average spectrum for the beach-state, the black line represents an f^{-4} energy roll-off, and the vertical dashed line demarcates the short-wave and long-wave frequency bands (From Hughes et al. (2014))





Swash Zone Dynamics, Fig. 5 Washover fan deposited on Ocracoke Island, North Carolina, during Hurricane Isabel, September 2003 (From Donnelly et al. (2006))

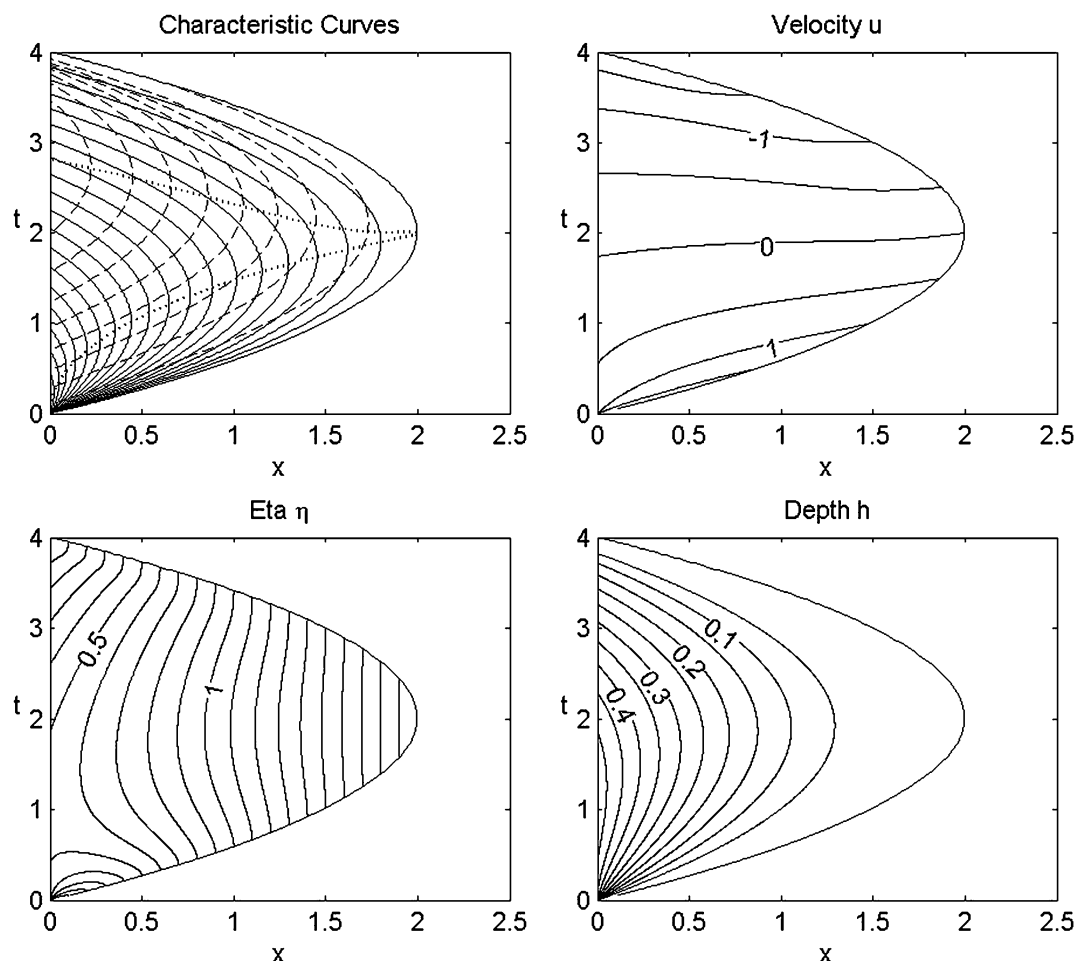
Runup and backwash velocities in the field reach 2–5 m/s, which are generally larger than those in the surf zone. The runup durations are typically shorter than the backwash duration, and the backwash depths are shallower than during the uprush, and therefore the velocity moments tend to be skewed offshore, which has important implications for the sediment dynamics (Raubenheimer et al. 1995). The asymmetry is however affected by the mass and momentum advected into the swash zone, which depends on the flow in the inner surf zone. Self-similar solutions for different boundary conditions are presented by Guard and Baldock (2007), following the work of Peregrine and Williams (2001), Fig. 6. These indicate the fundamental nature of the hydrodynamics, which comprise of a near parabolic motion of the shoreline (due to gravity being the dominant process) and a saw-tooth shaped variation in velocity with time, which decreases at a near linear rate from the peak velocity, which occurs as the shoreline passes a given location. The water surface slope dips seaward for nearly the

whole swash cycle, i.e., the total fluid acceleration is offshore throughout the swash cycle.

This results in a key difference between the surf zone and swash zone bed boundary conditions, namely that there is generally little phase lag between velocity and the bed shear stress in the swash zone, i.e., maxima in bed shear stress occur close to the instants of maxima in velocity. Close to the time of flow reversal, the flow near the bed does however reverse prior to the flow higher in the water column, due to the adverse pressure gradient during the uprush (Kikkert et al. 2012). The boundary layer is thinnest at the seaward edge of the swash zone during uprush, and grows following the flow up the beach. The boundary layer largely vanishes at flow reversal, and again grows from the bed as the flow recedes. Accounting for such processes in sediment transport models remains to be tackled.

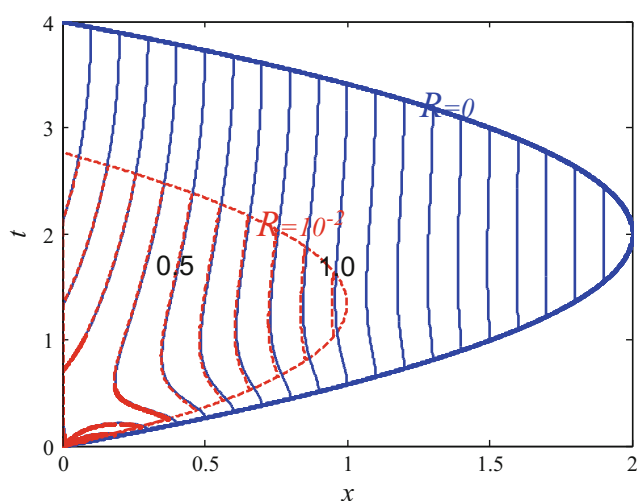
There are several sources of turbulence in the swash zone. In the runup, turbulence is advected from the inner surf zone, which combines with further generation of turbulence at the bed. The boundary layer is evolving, generally increasing in thickness and may become depth limited. During the backwash turbulence, generation occurs mainly at the bed, with swash-swash interactions generating further turbulence as the next wave arrives. Overall, the high turbulence near the bed leads to high bed shear stresses and the potential for high concentrations of suspended sediment transport (Puleo et al. 2000).

Friction plays a large role in controlling the shoreline motion, with friction factors based on conventional fluid mechanics principles typically in the range $0.02 < f < 0.1$. Friction effects are strongest when the flow is shallowest, at the swash tip, reducing runup excursions, and late in the backwash, so the shoreline recedes more slowly. Simple models, comprising of a ballistic motion plus friction, describe the shoreline motion reasonably well, although details are missing (Hughes 1995; Puleo and Holland 2001). Inclusion of friction effects in the internal flow requires numerical modeling at present, or the use of integral models which can avoid the uncertainty in the treatment of friction at the shoreline (Archetti and Brocchini 2002). However, recent results suggest that the effects of friction on the internal flow are small compared to the effects at the swash tip. For example, Fig. 7 shows a method of characteristics solution for the swash flow with and without friction. The numerical results indicate that the two solutions are similar when water is present, which is interesting given the significant effect of friction on the location of the shoreline. The reason is the supercritical nature of the flow, which is a particular feature of the swash zone. This means that the large change in the shoreline position due to friction does not significantly affect the flow seaward of the shoreline until the flow becomes subcritical, which does not occur until late in the uprush. Similarly, the supercritical nature of the backwash flow means that the changes to the shoreline position further



Swash Zone Dynamics, Fig. 6 Forward (solid) and backward (dashed) characteristic curves (a), and contours of flow velocity (b), surface elevation (c), and depth (d) for swash initiated by a near uniform

bore in the nondimensional coordinates of Peregrine and Williams (2001). Dotted lines in panel (a) show locus of $u = c$ (critical flow) for uprush and backwash (From Guard and Baldock (2007))



Swash Zone Dynamics, Fig. 7 Contours of surface elevation with two different resistance coefficients, $R = 0$ (blue lines) and $R = 0.01$ (red dashed lines) in the nondimensional coordinates of Peregrine and Williams (2001) (From Deng et al. (2016))

landward cannot significantly affect flow further seaward. Thus, the supercritical nature of the swash flow means that the significant changes in shoreline position do not significantly affect the flow in the interior of the swash lens.

Sediment Transport Mechanics

Cross-shore sediment transport in the swash zone generally occurs as a combination of bed load under sheet flow conditions, with a flat bed, plus an additional component of suspended load, generated locally and advected into the swash zone by surf zone bores. For the bed load, the Meyer-Peter and Muller (1948) formulation, or derivatives, generally perform well with calibration, i.e., determining the transport coefficient and friction factor remain problematic. In these models, the transport is typically a function of the velocity cubed. The relative balance between bed load, which is generated locally, and suspended load depends on the sediment grain size, and



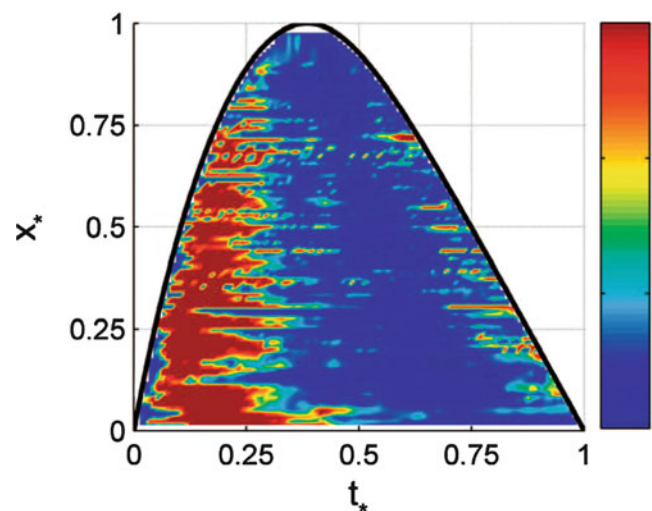
Swash Zone Dynamics, Fig. 8 A turbulent bore containing entrained suspended sediment just prior to reaching the swash zone. The sediment is then advected into the swash zone during the runup. The end of a

supercritical backwash flow is visible in the right of the image. The posts are 1 m apart and the orange stringlines are horizontal

also on the quantity of sediment advected into the swash zone from the inner surf zone (Fig. 8). This can be considerable, and affects the distribution of suspended load across the swash zone. While the basic sediment transport equations still apply, model-data comparisons are lacking, particularly close to the bed where suspended sediment concentrations are largest.

In simple u^3 -type sediment transport models the asymmetry of the flow tends to export sediment from the swash zone. In reality, this is balanced by the suspended sediment imported and advected into the swash zone from the inner surf (Hughes et al. 1997), the distribution of which is illustrated in Fig. 9. Measurement of suspended sediment in the backwash remains a challenge, and is difficult to separate from bed-load. The quantity of suspended load entering the swash zone affects the net deposition pattern, and hence zones of erosion or accretion, as illustrated by Pritchard and Hogg (2005). Data is still lacking in the field to reliably quantify this process, which is complicated by sediment suspended by swash-swash interactions and the influence of turbulence in the inner surf zone, which is significant (Butt et al. 2004). The infiltration and exfiltration of water into the beach during swash flows is an important contributor to groundwater processes, particularly on coarse grain beaches, and also influences sediment transport (Turner and Masselink 1998).

The mechanics of longshore sediment transport in the swash zone are well-understood in principle, but data and modeling is limited, despite the clear presence of a significant contribution to the total longshore transport in the littoral zone. In the swash zone, the flow direction during both runup and rundown has a longshore component under oblique waves (the usual case). The boundary between the surf and



Swash Zone Dynamics, Fig. 9 Measured normalized suspended sediment concentration c indicated by colors mapped onto the normalized x^*-t^* plane representing the swash excursion and duration. Color bar indicates normalized concentration from high (hot) to low (cold) (From Hughes et al. (2007))

swash zone is also an active region of longshore transport, particularly on steep cobble or coral beaches, where sediment may also move offshore into the inner surf zone, then along-shore, and then back onshore as the wave conditions or tide change (Kench et al. 2017). While the relative importance of longshore transport in the swash zone compared to the surf zone is greater during milder wave conditions than during storms, longshore sediment transport in the swash zone may account for up to 50% of the total longshore transport (Kamphuis 1991). Longshore transport in the swash zone is



Swash Zone Dynamics, Fig. 10 A LIDAR system mounted above the swash zone, together with ultrasonic distance point sensors. Photo: Dr. Chris Blenkinsopp, University of Bath, UK

also relatively more important on steep beaches and where small oblique waves break frequently at the shoreline, e.g., in estuary mouths and landward of lagoons behind fringing and barrier reefs.

Morphodynamics

The evolution of the beach face in the swash zone is controlled by the sediment fluxes across the boundaries, both longshore and cross-shore. Thus, the morphology is influenced by the presence of dunes, cliffs, and hard structures at the landward extent of the beach, by conditions in the inner surf zone, notably bars, rip cells, beach steps, and by lateral boundaries such as groynes, breakwaters, and estuary mouths. While the general processes are again well understood, and similar to those applied for littoral transport in the surf zone, the details of dune erosion, barrier degradation, and progradation and the short- and long-term balance of sediment deposition and erosion remain a challenge (Masselink and Puleo 2006). For example, rates of deposition and erosion vary significantly on wave-by-wave time scales, wave group time-scales, and tidal time-scales, plus seasonal and annual changes due to variations in wave and wind climate. Further,

while the active swash zone is usually relatively plane (excluding dunes), feedback between morphology and hydrodynamics does lead to more complex morphology such as beach cusps. Varying tides also lead to the formation of beach ridges and berms at different elevations, which complicate the overall topography. Dunes act as significant sources and sinks of sediment that control swash zone morphodynamics and that of the whole beach, and are essential in developing the sediment budget for the whole beach. A further aspect is the influence of vegetation, which can provide important stabilizing mechanisms for dunes. The upper beach and swash zone is where the impacts of sea level rise will be most visible, with loss of the upper beach if the coastline cannot recede landward.

Measurement Techniques

Research on swash zone processes is both assisted and hindered by the nature of the swash. The beach face is generally accessible for deployment of instruments, particularly with the aid of large tides, it is close to the shore and associated infrastructure, and much wave energy has been dissipated in the surf zone. However, flows are very shallow and also

intermittent, making measurements over the depth difficult and time-averaging complex. In terms of in situ instruments, the shoreline motion can be measured by runup wires or point instruments such as pressure sensors or ultrasonic sensors, flow velocity with electromagnetic or acoustic instruments, and morphology changes with standard survey techniques or ultrasonic sensors. To tackle larger scales and to avoid in situ measurements and to sample over longer time-scales, remote sensing by video has adopted techniques first developed for the surf zone by Aagaard and Holm (1989) to determine friction factors (Puleo and Holland 2001), runup (Stockdon et al. 2006) and internal kinematics (Power et al. 2011). Shore-mounted LIDAR is becoming a promising tool for high-resolution and long-term monitoring (Blenkinsopp et al. 2010), Fig. 10. However, both these techniques suffer from loss of resolution in the thinning backwash, and do not provide flow velocity with any degree of reliability. Sediment transport measurements using sediment traps (Horn and Mason 1994; Masselink and Hughes 1998) still provide the most reliable measure of total load transport rates in the field, supplemented by sediment transport rates derived from morphological measurements through sediment continuity, which can be particularly useful during over wash events. Both techniques require predominantly cross-shore transport.

Summary

The swash zone forms the land-ocean boundary, where waves runup the beach face, and is the region of the littoral zone most associated with beach erosion and the impacts of climate change. The geomorphology is very variable, and may include dunes, cliffs, marshes, ephemeral estuaries, and a wide variety of sand, gravel, rock, or coral barriers. Littoral sediment transport is high, in both cross-shore and longshore directions, which has a major influence on beach erosion and beach recovery during and after storms. Recent research provides methods to estimate wave runup and sediment transport with a fair degree of accuracy, although predicting the morphological response remains a major challenge. Advances in instrumentation, particularly remote sensing, are providing high-resolution data sets over large plan areas, and will assist in improving both hydrodynamic and morphological models for the swash zone and upper beach face.

Cross-References

- [Accretion and Erosion Waves on Beaches](#)
- [Beach and Nearshore Instrumentation](#)
- [Beach Erosion](#)
- [Beach Processes](#)
- [Coastal Erosion](#)

- [Coasts, Coastlines, Shores, and Shorelines](#)
- [Cross-Shore Sediment Transport](#)
- [Drift and Swash Alignments](#)
- [Erosion Processes](#)
- [Longshore Sediment Transport](#)
- [Nearshore Sediment Transport Measurement](#)
- [Nearshore Wave Measurement](#)
- [Remote Sensing of Coastal Environments](#)
- [Shore Protection Structures](#)
- [Surf Zone Processes](#)
- [Wave-Dominated Coasts](#)
- [Waves](#)

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Synthetic Aperture Radar Systems

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A synthetic aperture radar (SAR) is a remote sensing imaging system whose primary output product is a two-dimensional

Keith Raney: deceased.

mapping of the radar brightness of a scene. Radar brightness is an expression of the scene's reflectivity in response to oblique illumination by microwave electromagnetic emissions. By definition, a SAR must be mounted on a moving platform, such as an aircraft or a satellite, and its illumination is directed to the side (and downward) to the surface. Its images are formed by scanning the area in two dimensions: range and azimuth. Range scanning is essentially at the speed of light, as radar pulses are transmitted and their reflections (backscatter) are received and recorded. Azimuth scanning is accomplished by the forward motion of the radar. SAR imaging "works" because the speed of light is very much greater than the along-track velocity of the radar. As the radar progresses along its flight path, the reflected signals from each transmission are collected and stored in memory. The heart of the system is the processor, which derives output image products from the stored data. The name "synthetic aperture" reflects the fact that signal processing algorithms replicate the signal collecting functions of a real aperture antenna array, thus "synthesizing" the effects of a physical antenna from a virtual data array.

One can visualize the key stages in the system through an optical comparison—a simple camera. A SAR's data memory is analogous to a record of the light field from a scene that may impinge on the front surface of a camera's lens, and the processor imitates (at least mathematically) the actions of the lens as it focuses the field and directs the refracted rays to the camera's film. As in a camera exposing black-and-white film, larger reflectivity (greater radar brightness) is expressed as lighter tones of gray tending toward white in a radar image. Lower reflectivity (less radar brightness) is expressed as tones of darker gray tending toward black. Colors may be artificially imposed on radar images to portray added dimensions, such as differences in reflectivity between two or more images, or differences in reflectivity as a function of wavelength or polarization.

A Brief History

The original SAR idea is due to Carl Wiley in 1951. It was first reduced to practice by the University of Illinois, and by the Willow Run Laboratories of the University of Michigan. Several airborne systems were developed in the 1960s, although the technique was largely classified at the time. The first satellite SAR was embarked on Seasat (1978). Important earth-observing satellite SARs since then include Japan's J-ERS, the European Space Agency's ERS-1 and ERS-2, and Canada RADARSAT (1995), which was still operational in June 2004.

Signal processing has been the pacing development in the history of SAR. Early SAR processors were analog optical contraptions, excited by lasers, and depended on relatively conventional photographic film for both the memory medium and the output image format. The first digital processors were

developed for Seasat. In the fall of 1978, typically it would take more than 40 h to process one 50 km by 50 km scene, approximately one quarter of the nominal 100 km square Seasat SAR frame. Today, primary SAR image processing has progressed to the point that it is nearly transparent to most users of the data.

Basic Concepts and Parameters

Resolution is the minimum separation between two small and similar side-by-side reflectors such that they appear discretely in the image to be two, rather than merged into one. The distinguishing SAR performance attribute is the *azimuth resolution* realized through the synthetic aperture technique. The primary advantage of the SAR approach is that the radar-processor combination using data collected by a relatively small antenna can replicate the resolution that could otherwise be delivered only by an impossibly large antenna. For example, the Seasat SAR in its finest (or highest) resolution mode had about 6 m azimuth resolution. If a real aperture imaging radar were to have flown in place of Seasat, its antenna would had to have been about 16 km long (!) to achieve the same results. Azimuth resolution is proportional to range and inversely proportional to antenna length for a real aperture radar, but it is approximately equal to one half of the antenna length for a SAR. Seasat's SAR antenna was only 10 m long, not 16 km. Quite a difference, but not atypical of satellite systems that operate at 1,000 km range or so.

As microwave systems, most SARs employ relatively long wavelengths in contrast to optical or infrared systems. The wavelengths used by earth-observing satellite SARs include L-band (23-cm), S-band (10-cm), C-band (6-cm), and X-band (3-cm). RADARSAT is at C-band. At these wavelengths, roughness is the primary geophysical characteristic of the illuminated surface that determines the power of the back-scattered signal, the radar brightness portrayed in SAR imagery. Surface roughness must be judged relative to the radar's wavelength, conditioned by the incident angle of the illumination at the surface, and other less important attributes. A surface appears rough to the radar as the topographic relief becomes comparable to the illuminating wavelength. A given surface always appears to become smoother as the angle of incidence is increased.

SAR backscatter may be enhanced considerably by favorable multi-reflection geometries. For example, stands of mangroves usually appear as bright areas, because the radar illumination bounces off both the water's surface and tree trunks, sending a relatively strong reflection back to the radar. The resulting image contrast is strongest for mangroves in calm (smooth) water.

Synthetic aperture formation requires that the data over the effective azimuth processing field must be phase-coherent. This leads to significant advantages for interferometric

measurements, for example, which have demonstrated the ability to measure systematic surface displacements on the order of centimeters.

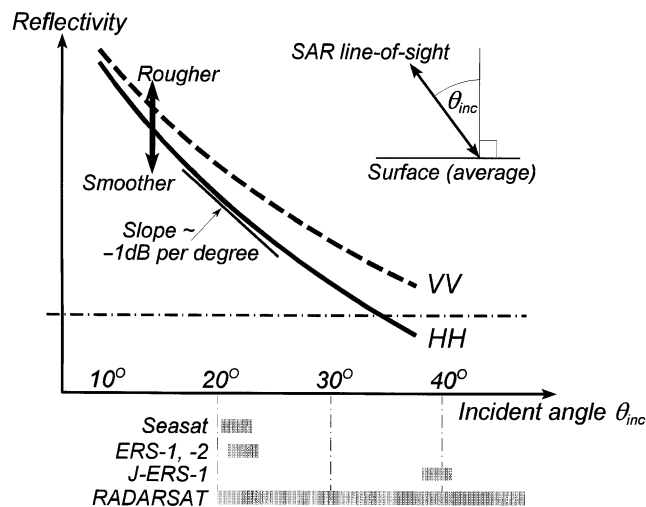
There is an inherent and not always pleasant consequence to phase coherence. For reflections from many distributed reflectors such as the ocean's surface, the random phases of the individual scattering facets cause mutual interference, leading to brightness modulation of the imaged field. This is known as speckle, whose variance is proportional to the mean radar brightness of the neighboring image area. Speckle may be reduced by averaging, which implies a signal-processing tradeoff with image resolution. "Looks" describes the number of statistically independent sub-images used to form the averaged image. The number of looks, divided by the product of the resolution parameters (in range and azimuth), is a fundamental SAR parameter, which is proportional to the Nyquist information capacity of the system. Usually, SAR range resolution is equal to the inverse range bandwidth (scaled to ground range), and azimuth resolution is approximated by half of the azimuth aperture length multiplied by the number of statistically independent looks. For example, the Seasat SAR used four looks in the processor, thus compromising its azimuth resolution to about 25 m, while reducing the speckle noise variance to a reasonable level (for many applications). The standard mode of RADARSAT, and the conventional modes of ERS image products all have comparable image quality, at least as rated by this norm. A significant increase in this parameter would imply a significant increase in SAR system cost.

All radar observations of scene reflectivity are in competition with additive radar noise. The additive noise is described by an equivalent reflectivity number, known as the noise equivalent reflectivity. Unlike speckle, the effect of additive noise is suppressed for higher relative levels of radar brightness, which can be due to either increased reflectivity or larger effective radar power.

Swath width, the span of the imaged area measured on the surface orthogonally to the satellite path, is the last of the top-level SAR image parameters. Swath widths typically are about 100 km for nominal 25-m 4-look image products. Higher resolution modes induce smaller swaths (e.g., 40 km). The extra-wide (~500-km) swaths available from RADARSAT's ScanSAR mode are of particular interest to the oceanographic community, and can be interpreted to depict the detailed structure of nearshore wind fields. The trade-off implied by a ScanSAR image product is coarser resolution, which for a well-designed system can be offset somewhat by more looks, thus preserving image quality in spite of increased swath.

The SAR Seen Sea

For a distributed scene such as the sea, radar brightness is an expression of radar reflectivity, which in general is a function

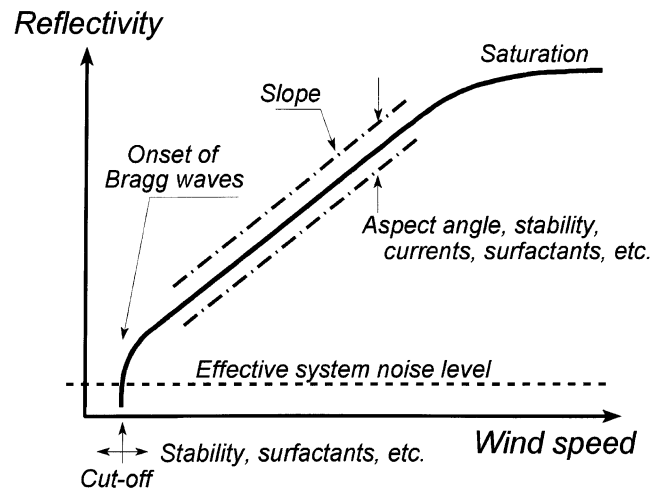


Synthetic Aperture Radar Systems, Fig. 1 Radar reflectivity (backscattered power) tends to decrease with increasing incident angle, and is larger for rougher surfaces. The span of incident angles of several satellite SAR systems is indicated below the horizontal axis

of instrumentation and of geophysical parameters. The instrumentation factors include polarization and wavelength of the radar illumination, and the incident angle at which the scene is observed. Geophysical parameters include surface roughness and conductivity, and environmental factors such as water temperature, air-sea temperature difference, wind speed, and surfactants, all of which have an impact on the ocean's roughness at microwave scales. The radar reflectivity of the ocean tends to decrease with increasing incident angle, as shown in Fig. 1.

Polarization of an emitted electromagnetic field is determined by the geometry of the antenna. The observed reflectivity depends on the polarization state of the receive antenna as well as of the transmitted field, where vertical (V) or horizontal (H) linear polarizations are the most commonly used for SAR. Most systems use the same antenna for both transmission and reception, leading to like-polarized reflectivity estimates. Ocean reflectivity is usually larger for VV polarization than for HH polarization.

For a wide range of illumination incidence, Bragg scattering is the dominant source of radar backscatter from the ocean. Bragg scattering is due to a systematic phase fit between the illuminating half-wavelength projected onto the sea surface, and a matching periodic ocean roughness component. For the wavelengths and incident angles of typical ocean remote sensing radars, Bragg waves range from long capillaries (~ 2 cm) to short gravity waves (~ 50 cm). "Roughness" in the microwave ocean remote sensing context implies a sensible population of these special Bragg surface wave components. These waves, which have relatively short lifetimes, are locally generated, and are modulated and advected by longer and more energetic seas. It follows that the larger



Synthetic Aperture Radar Systems, Fig. 2 Between the lower and upper limits, the ocean's reflectivity (relative power) increases in proportion to local wind speed. The slope and level of the proportionality depend on radar and environmental parameters. The lower limit is relatively abrupt; backscatter arises only above the minimum wind speed sufficient to excite Bragg waves matched to the radar. The upper saturation limit is more gradual

wave structure of the ocean is observable by microwave systems primarily through patterns expressed in the shorter Bragg waves. This is known as the two-scale model of microwave ocean scattering.

Roughness at Bragg wavelengths is set up primarily by local wind stress. For a given incident angle and radar wavelength, there is a first order power law dependence between radar reflectivity and wind speed, as suggested in Fig. 2. The slope of the quasi-linear part of the response is a function of wavelength, among other parameters. Within this quasi-linear region, a SAR's radar brightness can be inverted to estimate wind speed, realizing in effect a high-resolution scatterometer.

Within limits, the power law is robust. The upper limit is approached as the radar reflectivity tends to saturate with increasing wind speed. The saturation wind speed is above 25 m/s for the radar parameters of RADARSAT at its steeper incident angles, for example.

The lower limit is more abrupt, and may be of considerable interest in coastal oceanographic applications. Bragg wave formation, and hence sensible radar brightness, occurs only for wind speeds above the lower cutoff threshold, which typically is about 2.5 m/s. Both the onset of Bragg waves and the level of reflectivity are functions of environmental conditions. If the system additive noise is sufficiently low such that the reflectivity threshold is visible, then relatively small changes in local environmental conditions may lead to large contrasts in radar brightness. For example, a change in relative air-sea temperature of only one degree or so can invert the boundary layer stability, which can be manifested as a

relatively large change in radar reflectivity. These contrasts are often observed in oceanic SAR imagery, and have proven to be invaluable in tracking oil spills, for example. In general, the challenge is to deduce the appropriate geophysical cause from radar brightness patterns seen in the image data.

In oceanographic applications, the scene of interest is in motion, which can have noticeable consequences on oceanic SAR image characteristics. Gravity waves are characterized by the orbital velocities of their elemental surface constituents. These motions are transformed by a SAR into azimuth shifts of the corresponding image features. As a result, an image of azimuthal waves is distorted, an effect known as “velocity bunching.” Directional wave spectra can be deduced from oceanic SAR imagery, although extra effort is required to compensate for the degraded azimuthal wave components. The “wave modes” of ERS-2 and Envisat SAR have been designed with these characteristics in mind.

Summary

Image data from satellite-based SAR systems are in increasing demand. Applications span directional wave spectral estimation, ship detection and tracking, oil spill monitoring, fisheries monitoring, coastal wind analysis, coastal erosion,

and storm tracking. The major limitation at this time is lack of timely data, due primarily to a shortage of operational systems. This limitation should be at least partially rectified in coming years.

Cross-References

- [Mangroves, Ecology](#)
- [Mangroves, Geomorphology](#)
- [Mangroves, Remote Sensing](#)
- [Photogrammetry](#)
- [RADARSAT-2](#)
- [Remote Sensing of Coastal Environments](#)
- [Shoreline and Coastal Terrain Mapping](#)
- [Waves](#)

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Tafone

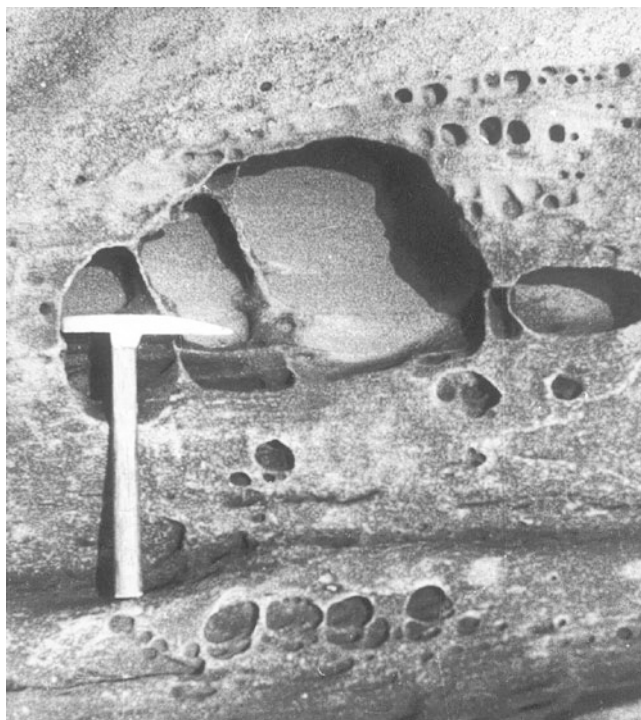
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Originally used to describe unusual weathering features found in Corsica, tafone (or tafoni, plural) has become established as the generic name for a type of cavernous weathering characterized by the existence of hollows or cavities that range in size from a few centimeters to a meter or more in diameter. Depth is variable, the cavities commonly being nearly hemispherical. The shape and orientation are strongly influenced by the rock fabric, causing the hollows to be elongated in the direction of foliation of bedding planes. Coalescence of the cavities produces mushroomlike shapes, natural arches, and other unusual sculptured forms. Eventually the outcrop surface may be destroyed by this process of expansion. When numerous small cavities occur, the resulting spongelike texture is termed honeycomb weathering. Both honeycomb weathering and tafone may occur independently, but often the two forms coexist and seem to originate from a similar process. Tafone occurs commonly in granitic rocks and in sandstone, but has also been observed in many other rocks. The rate of development may be rapid, cavities having formed in stone seawalls built less than a century ago.

Tafoni are abundant in the polar regions of Victoria Land, Antarctica, but they also occur in deserts of Australia, Africa, central Asia, and South America as well as in such humid regions as Hong Kong and the island of Aruba, West Indies. In North America, tafoni occurs in such diverse areas as the deserts of New Mexico, Utah, Arizona, and the coast of Washington state (Fig. 1), and at mid-continent locations in Wisconsin and Illinois (Bryan 1928; Blackwelder 1929; Mustoe 1982). Tafoni are commonly found in outcrops having a hardened surface layer caused by precipitation of iron hydroxides or other compounds derived during weathering of

the interior rock. The cementing action of these oxides produces a resistant rind. Any form of erosion that attacks the outcrop in a localized fashion will produce cavities, since penetration of the protective layer leads to rapid destruction of the weaker interior rock. Differential attack of the surface might result from variations in lithology, the presence of fissures or zones of high porosity that allow water to penetrate, or localized attack by organisms such as lichens. Because the formation of the protective ring requires moisture, this explanation is consistent with the observation that tafone seldom occurs in extremely dry environments, being more common along the margins of deserts than in the arid central regions. In many locations, tafoni occur mainly as coastal features presumably owing to the existence of a favorable microclimate, since even in extremely arid regions sea winds may cause the coastline to be relatively humid. The existence of a protective rind is most favored by an environment where some moisture is present but where periods of dryness allow evaporation to occur at the rock surface.

Tafoni typically form as a result of salt weathering, where evaporation of saline water triggers physical and chemical attack of the rock surface (Evans 1970; Young 1987; Rodriguez-Navarro and Doehne 1999). In general, the distribution of tafone throughout the world correlates well with environments where salt crystallization occurs. In arid zones, salt weathering is concentrated where moisture is retained along the base of cliffs and undersides of boulders, where shadow weathering (tafone) may often be found. Along the coast, cavities are presumed to form by salt crystallization as wave-splash evaporates. Because salt weathering may attack in a relatively selective fashion, being controlled by variations in moisture or exposure to salt spray, tafoni might be formed even on outcrops where a hardened surface layer is absent, though the development of cavities would be enhanced when such a rind is present. Early explanations of tafoni invoking erosion by wind action, temperature fluctuation, or frost wedging have largely been abandoned owing to lack of substantiating evidence.



Tafone, Fig. 1 Tafone in arkosic sandstone, Larrabee State Park, near Bellingham, Washington USA. (Photo, George Mustoe)

Cross-References

- [Bioerosion](#)
- [Cliffs, Lithology versus Erosion Rates](#)
- [Coastal Climate](#)
- [Coastal Hoodoos](#)
- [Coastal Wind Effects](#)
- [Desert Coasts.](#)
- [Honeycomb Weathering](#)
- [Notches](#)
- [Shore Platforms](#)
- [Weathering in the Coastal Zone](#)

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Tectonics and Neotectonics

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Introduction

Rock deformation caused by the structure of the earth (e.g., folds, faults, joints, cleavage) is often the only kind of tectonic deformation considered by geological manuals, giving the impression that areas devoid of such type of deformation are “tectonically stable.” However, according to the American Geological Institute (1960), “tectonic” is defined as “designating the rock structure and external forms resulting from the deformation of the earth crust.” This definition implies that all processes which modify the external form of the crust, also when they result from forces external to the earth, have to be considered tectonic. This is the case, for example, for unidirectional vertical movements produced by earth surface processes of weathering and erosion (sedimento-isostasy), and also for rise and fall of the solid earth surface, especially in coastal areas, caused by external factors such as climate change (glacio-isostasy, hydro-isostasy) or eventually, at a smaller timescale, by the attraction of the sun and the moon (earth tides).

The term “neotectonics,” ignored by most geological dictionaries or glossaries, was introduced by geodesists, geophysicists, and quaternarists during the last decades and defined in 1978 by the Board of the Neotectonic Commission of the International Association for Quaternary Research (INQUA) as “any active earth movement or deformation of the geodetic reference level, their mechanisms, their geological background (how old it ever may be), their implication for various practical purposes and their future extrapolation.” This definition shows care not to isolate crustal movements from their geodynamic inheritance. Therefore, neotectonics has a wider scope than “active tectonics” (Wallace 1986), which is defined as “tectonic movements that are expected to occur within a future time span of concern to society” and no real boundary back in time. It includes all timescales of movements, from instantaneous (seismic), 10 – 10^2 year (geodetic), 10^2 – 10^4 year (Holocene studies), 10^4 – 10^6 year (Pleistocene studies), up to about 10^7 year, if it is necessary to enable us to understand the origins of recorded movements (Mörner 1980). The difference between former tectonics and neotectonics; is therefore not far from that existing between an extinct fossil and a still living organism. Among the distinctive attributes of neotectonic studies listed by Stewart and Hancock (1994) are: (1) A wide range

Paolo A. Pirazzoli: deceased.

of methodologies and a variety of experts are commonly involved in a comprehensive study of the neotectonic history of a region, whereas paleotectonic structures are often investigated by structural geologists working on their own. (2) The neotectonician has the ability to compare inferences drawn from field observations with geophysical and geodetic data about rates and mechanisms of present-day processes. (3) Neotectonic displacement histories; can be established with greater precision than paleotectonic ones, because Quaternary dating techniques allow relatively short time intervals to be detected. However, the approach by Stewart and Hancock (1994) mainly refers to active fault ruptures which are indeed privileged indicators of neotectonic displacements, but remains limited to boundaries of deformed crustal blocks, which may also remain obscured, without reaching the earth surface, or become obliterated by recent erosion or sedimentation processes.

In the following, various possible processes of tectonic/neotectonic deformation in coastal areas are briefly reviewed, with special attention to vertical displacements which may be related to sea level. Significant examples of tectonic displacements are provided and rates of vertical movements are assessed. Some developments are partly inspired from ideas already discussed by Pirazzoli (1995).

Present-Day and Fossil Coasts

The coastal outline has undergone major changes during the earth history. Some 300 Myr ago, when all continents are believed to have formed a single, huge landmass, Pangea, marine coasts could only exist along the perimeter of such a landmass. This supercontinent started to split apart about 200 Myr ago, first into two parts, Laurasia at the north and Gondwana near the South Pole, separated by a sea, Tethys. Subsequently Laurasia split out into North America, Europe, and North Asia, whereas Gondwana fragmented into South America, Africa, Arabia, Madagascar, India, Australia, New Zealand, and Antarctica. Lastly, a slow drift has brought the various continents or continental fragments to their present state.

The reconstruction of the positions of the continents during the splitting or at the time of their coming together, during their migration, as well as the mechanisms of plate tectonics, make possible a comprehensive view of present-day and fossil coastal areas, with some marks of the latter now perched also near the top of mountain chains.

When the separation of two continents had been caused by the opening of a rift, continental margins are usually characterized by a low seismicity and a weak orogeny; in these structurally passive continental margins, which are devoid of an oceanic trench, the continental and the oceanic crust, which belong to the same plate, are joined. Such margins are frequent around the Atlantic Ocean and

characterized by a well-developed continental shelf and a relative structural-tectonic stability (though they may be affected by important vertical deformation of isostatic origin).

On the other hand, where subduction processes occur, tectonics of continental margins are structurally active and show high seismicity and relief. Such margins, typical of Pacific coasts, correspond to areas where an oceanic plate is underthrusting beneath a continental plate. The continental shelf if present is very narrow, or even absent. Coastal California corresponds to a variant, with a sliding motion between the Pacific oceanic plate and the continental North-American plate along the San Andreas Fault.

In accretion zones, two continental plates are colliding. These active continental margins with strong seismicity and orogeny are generally located inside continents, far away from present-day coasts. However, at the first stages of collision, wide marine and coastal areas can be affected. This is the case in the present time of the Mediterranean region. Lastly, the coasts of epicontinental seas (bordering oceanic depths) form a special case, because the greater part of these shallow seas extending above continental shelves usually disappear during low sea-level periods and are therefore affected by significant hydro-isostatic vertical deformation.

Processes

Thermo-Isostatic and Volcano-Isostatic Tectonics

It is now generally accepted that plate tectonics modify, slowly but continuously, the shape of the oceanic basins. The rock of oceanic plates is formed by lava extrusions along oceanic ridges. After its formation, the oceanic crust moves slowly away and crosses the ocean, to be later destroyed in a subduction trench. When the crust spreads away from the ridges, it cools and thickens, thereby increasing in density. As a result the seafloor subsides isostatically, gradually submerging oceanic islands as they are carried laterally. Over several million years, thermo-isostatic subsidence will be on the order of kilometers. This can be verified easily in any world atlas, where the ocean depth is shown to increase gradually from 2,000–3,000 m above submarine ridges to 5,000–6,000 m or more above oceanic trenches. In tropical waters, the rate of thermo-isostatic subsidence is increased by the load of coral reefs, which have to grow vertically to maintain their sea-level position. The normal vertical evolution of an oceanic island in intraplate areas, when eustatic fluctuations of sea level are disregarded, would therefore be a gradual submergence, accompanied by relatively rapid erosion of exposed rocks by subaerial weathering and wave action. The existence of such general oceanic subsidence was first observed by Darwin, who used it to explain the different types of coral reefs by evolution from

fringing reefs, to barrier reefs, and to atolls. Darwin did not know what the cause of subsidence was, but his theory was correct. We know today that oceanic intraplate subsidence is due to thermo-isostasy.

If an intraplate oceanic island is not subsiding, it is anomalous and requires explanation. Two main kinds of anomalies have been recognized: thermal rejuvenation and volcano-isostasy.

When the oceanic crust approaches a hot spot, which normally corresponds to an asthenospheric bump above a more or less fixed mantle melting anomaly, the normal cooling process is reversed, so that the crust is heated, becomes less dense and thinner, and thus rises (thermal rejuvenation process). Moving away from the hot spot area, cooling and subsidence predominate again.

Near hot spots, lava eruptions often occur. Their load produces on the lithosphere an isostatic depression under the volcanic mass and a peripheral raised rim at some distance (volcano-isostasy). According to empirical observations made around oceanic islands, the radius of the depressed area is generally less than 150 km from the load barycenter, whereas the peripheral rim develops at a distance of between some 150 and 300 or 330 km. When the translation movement of an oceanic plate over a hot spot has produced a line of islands, and two or more thermal rejuvenation spots form an alignment in the direction of plate movement, interaction between isostatically depressed areas and uplifted rims of nearby islands is possible and may produce complicated sequences of vertical deformation, with repeated phases of uplift and sinking.

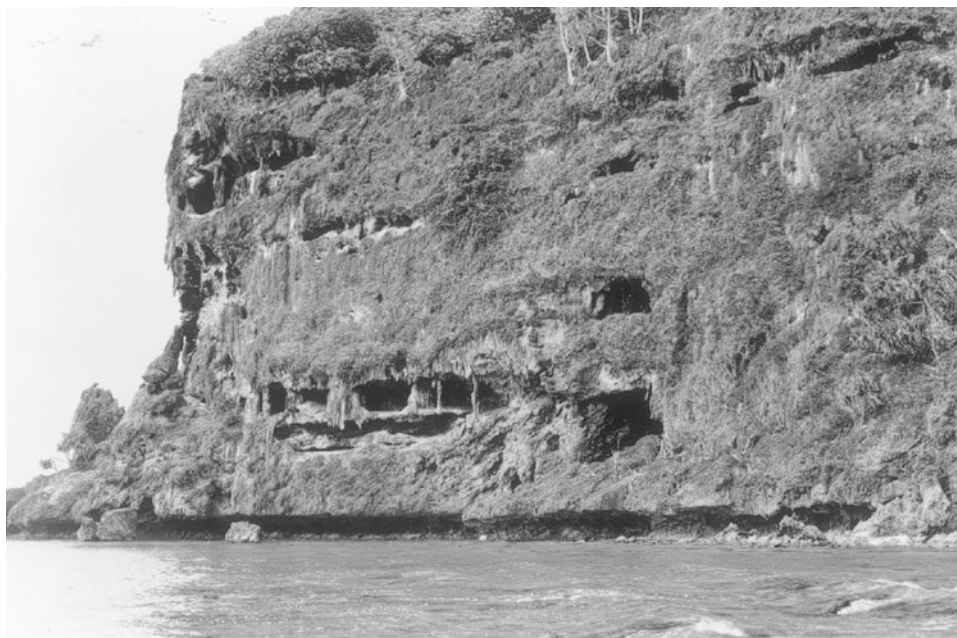
Example of Quaternary thermo-isostatic uplift: Rurutu Island (South Pacific). Rurutu (22°30'S–151°20'W) is

a basaltic island of the Austral–Cook island chain, approximately halfway between the hot spot near the active undersea Mac Donald volcano and the northernmost atoll of the Cook Islands (Palmerston). The volcanic basement of Rurutu has been dated 10.5 ± 2 Ma, whereas the age of the surrounding seafloor lies in the range of 40–90 Ma. The age of the volcanic basement of Rurutu is therefore consistent with the migration rate (10–11 cm/year) of the Austral–Cook lithosphere with respect to the hot spot reference frame and the distance of about 1,350 km from Mac Donald. There is, however, a second hot spot along the Austral–Cook island chain, which generated some Cook Islands (Mauke, Atiu, Aitutaki) during the last 10 Myr. This second hot spot or plume is located near the present position of Rurutu (Dickinson 1998). During the Quaternary, Rurutu entered the uplifting side of the hot spot swell. No new lava extrusion seems to have occurred, but gradual uplift took place, so that today the volcanic basement of Rurutu is surrounded by raised limestone tabular blocks, dated less than 1.85 Ma, reaching about 100 m in altitude. The limestone blocks finish in high vertical cliffs facing the ocean. Variations in the water table associated with Quaternary sea-level changes superimposed on gradual uplift have provoked karstic dissolution in the limestone at several levels (Fig. 1). Dated remnants of the Last Interglacial coastline are now at +8–10 m in elevation; marks of Holocene sea stands have been identified at +1.7, 1.2, and +0.6 m (Pirazzoli and Salvat 1992). The available evidence indicates that thermo-isostatic deformation has uplifted Rurutu Island at the average rate of 0.05–0.10 mm/year since the early Pleistocene and at about 0.17 mm/year in the Holocene.

Example of historical rapid thermo-isostatic deformation: “Temple of Serapis,” Pozzuoli (Italy). This site demonstrates

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Fig. 1 The outer limestone cliff of Rurutu (Austral Islands, South Pacific) appears cut by networks of caves at several levels, remnants of karstic drainage corresponding to higher relative sea-level positions. The lowest line of caves, with the floor at +8–10 m, contains marine material which has been dated from the Last Interglacial. The cliff is also undercut by a slightly elevated deep notch dating from the Late Holocene. (Photo 5743, April 1980)



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Fig. 2 The “Temple of Serapis” in Pozzuoli (Italy). Note the dark band on the columns produced by molluscan borings (the upper limit of which is slightly less than 6 m above the paving of the “Temple,” which is partly flooded by a few centimeters of water). (Photo D388, October 1991). The same floor appears more submerged in photos taken in 1982–83 (Vita-Finzi 1986, p. 10)



repeated rapid up and down displacements in the Phlegraean Fields caldera, near Naples. Burrows of *Lithophaga mollusca* in the three columns still left standing of the “Temple” (a Roman market probably built in the second century BC near the Pozzuoli Harbor), clearly show that the ruins had subsided under sea level and then been gradually uplifted (Fig. 2). Geomorphological, archaeological, historical, and radiometric data suggest a complex relative sea-level history. After the “Temple” was constructed, a subsidence of about 12 m took place. Elevated marine fossils indicate two submersion peaks, between the fifth and the seventh century AD, then again in the thirteenth to fourteenth century AD (Morhange et al. 1999). An uplift of about 7 m followed, culminating in the Monte Nuovo volcanic eruption in AD 1538, and further subsidence. More recently, tide-gauge data from the Pozzuoli Harbor show two brief periods (1970–73 and 1982–84) of rapid uplift reaching a total of 3.2 m between 1968 and 1984, followed by slight renewed subsidence, still going on. Such vertical movements are understood as being of thermoisostatic origin, with uplift corresponding to rock expansion under the action of heat beneath the caldera, and subsidence to contraction on cooling.

Vertical Tectonic Deformation Near Plate Boundaries

When oceanic crust approaches the end of its travel, near a trench or a continental margin, the arrival of its mass in a subduction or a collision zone produces high seismicity and vertical tectonic movements become more complex.

In a subduction zone, the underthrusting side, which will be destroyed by plunging into the earth mantle, has a relatively simple tectonic behavior, differing notably from that of the overthrusting side. Before being submerged, the oceanic

plate is generally subjected to arching phenomena, in order to make possible the change from a horizontal translation to a plunging beneath the overriding plate. This arching implies a wave-like flexuring of the lithosphere, with first a gradual uplift, as in the case of an oceanic island approaching a hot spot, than a gradual subsidence at accelerating rates. Case studies of such arching phenomena have been reported from Christmas Island, 200 km southwest of the Sunda Trench, from the Daito (Borodino) Islands, 150 km east of the Ryukyu Trench, and from the Loyalty Islands, at varying distances west of the New Hebrides Trench.

On the overthrusting side, various local or regional structural geodynamic factors may be superimposed. Here uplift is frequent and may be caused by: (1) piling up above the oceanic plate, on the inner side of the trench, of sediments too light to be subducted, which will raise the overthrusting edge isostatically; (2) elastic rebound phenomena linked to subduction faulting; (3) tilting of lithosphere blocks or other structural tectonic processes; and (4) volcanic activities. Subsidence in structural basins may also be significant.

Where subduction processes are prevented by the occurrence of a transform fault or by continental lithosphere on both sides of a plate boundary, collision or transform processes will occur with vertical movements which can become highly irregular. The best-known example of a transform fault crossing a coastal area is probably that of the San Andreas Fault, which extends south to north over a distance of about 1,300 km between the Gulf of California and the Mendocino Fracture Zone. South of the Mendocino triple junction, strike slip predominates along a broad and braided system of faults with a roughly northwesterly strike; blocks between faults are warped, folded, uplifted, depressed, and rotated.

Farther northwestwards, vertical movements predominate (Crowell 1986). In several areas along the San Andreas Fault evidence of uplift is missing, however, and aseismic uplift has also been reported from southern California.

The superimposed effects of long-term gradual uplift and of eustatic sea-level fluctuations often produce sequences of stepped marine terraces.

Example of stepped marine terraces in the Santa Cruz area, California. Six marine terraces, in which raised shell beds were first described by Darwin, indent a 60 km stretch of coastal topography. These terraces seem to have been raised by repeated slip earthquakes on the San Andreas Fault, with a return time of three to six centuries, uplifting the terraces at average rates between 0.13 and 0.35 mm/year (Lajoie 1986; Valensise and Ward 1991).

Seismic Displacements

In many seismic areas, vertical displacements of land may appear gradual in the long term, but in the short term they can consist of sudden vertical movements, separated by more or less long periods of quiescence or of gradual movement (Fig. 3). The sudden movements usually take place at the time of great-magnitude earthquakes, which are often accompanied by surface faulting or folding and ground deformation. Land displacements occurring spasmodically at the time of an earthquake are called *coseismic*. Gradual displacements, often in opposition to the coseismic ones, may occur during

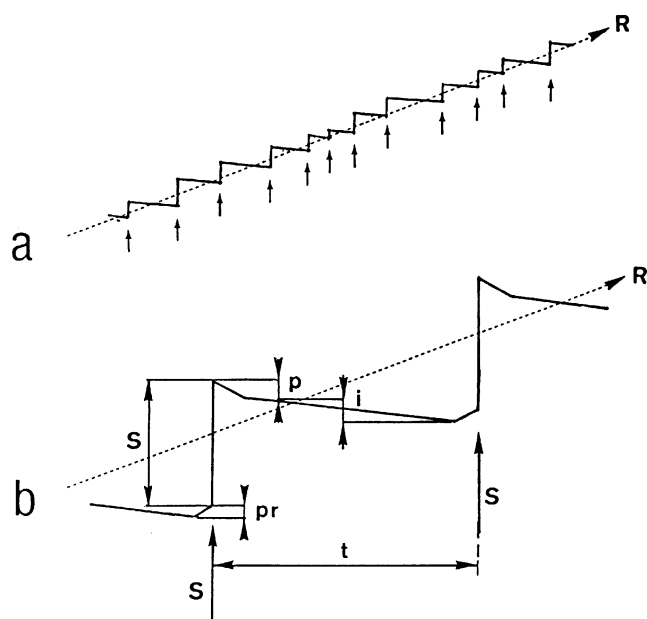
the few years or decades preceding (*preseismic*) or following (*postseismic*) the coseismic event. The duration of the time interval between two coseismic events (*interseismic* period) is not perfectly regular and depends on the variability of the local tectonic stress accumulation. It may tend, however, to be repetitive statistically, with a return time which can vary, according to the seismotectonic area considered, from a few centuries to over ten thousand years. In coastal seismic areas, investigation and dating of former shorelines different from the present ones may be used to determine the age, distribution, and succession of vertical displacements. Of special interest for seismic prediction is the identification of the preseismic movements which may indicate the imminence of a great-magnitude earthquake.

Certain coseismic movements can be impressive. In Crete (Greece), a jerk of coseismic uplift raised the southwestern most part of the island as much as 9 m, probably on July 21, AD 365 (Pirazzoli et al. 1996); in Alaska, the great earthquake of March 27, 1964 (magnitude ≥ 8.4) was accompanied by uplift reaching 10 m on land in Montague Island and more than 15 m offshore (Plafker 1965).

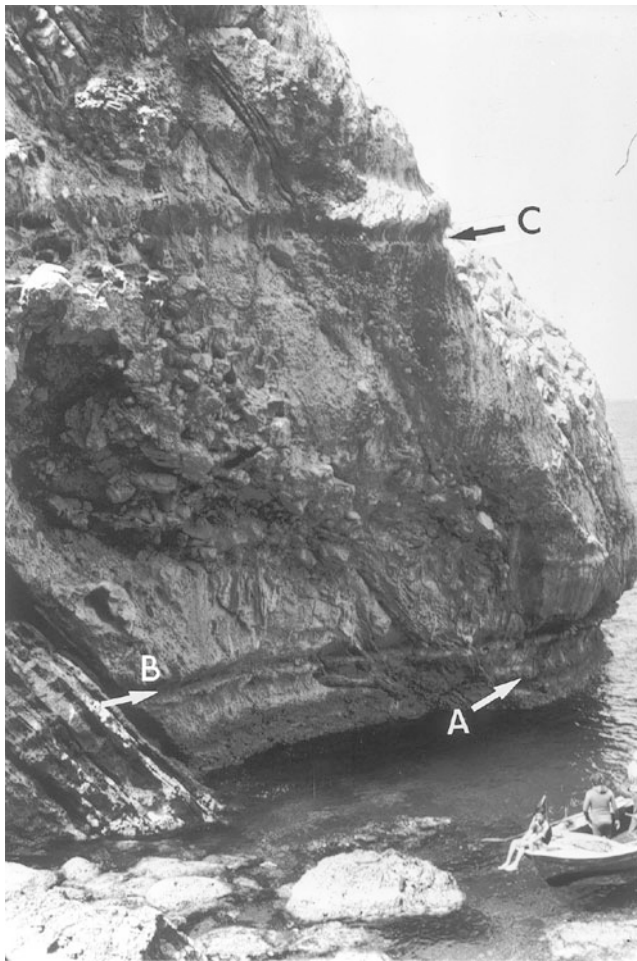
Example of repeated Holocene coseismic deformation: Hatay (Turkey). On the Hatay coasts, clear evidence of uplifted Quaternary coastlines near the Orontes Delta have been recognized up to at least 140 m in elevation. The lower evidence, between 35 and 8 m above present sea level, consists of stepped, elevated, deltaic sedimentary surfaces, suggesting that the Orontes River has built here intermittent delta formations since at least the late mid-Pleistocene. Detailed study of late Holocene marine notches (Fig. 4) and bioconstructed rims made of vermetids, oysters, and calcareous algae, have demonstrated that uplift has been episodic rather than gradual (Pirazzoli et al. 1992). Two rapid land movements, probably of seismic origin, have been identified. The first movement, which occurred about $2,500 \pm 100$ years BP, was the strongest one and caused a local vertical displacement of about 1.7 m. The second movement (an uplift of 0.7–0.8 m) occurred around 1,400 years BP, probably at the time of the great earthquake followed by tsunami waves of May AD 526, which caused devastating damage in Antioch and prevented further use of its harbor Seleucia Pieria. The long-term trend of uplift in this coastal area can be ascribed to episodic reactivation of local fault lines, probably in connection with movements on the East Anatolian Fault system.

Glacio-Isostatic and Hydro-Isostatic Neotectonics

Transfers of ice or water masses between ice sheets and the ocean produce load displacements and therefore isostatic phenomena. The load of an ice sheet deforms the earth's crust. The resulting subsidence beneath the ice makes deeper material flow away, and raises an uplifted rim at a certain distance. When the ice sheet melts, an unloading occurs, resulting in uplift beneath the melted ice; the marginal rim



Tectonics and Neotectonics, Fig. 3 In uplifting seismic regions, the raising trend, though apparently gradual over the long term, may consist in the short term of sudden uplifting movements accompanying great earthquakes, separated by more or less long periods of seismic quiescence and gradual subsidence. *pr* preseismic, *S* coseismic, *p* postseismic, *i* interseismic, *t* return period



Tectonics and Neotectonics, Fig. 4 Two continuous notches cut into the limestone cliff at about (a) 0.7 m and (b) +2.0 m, mark the sea-level positions before two coseismic uplift movements which occurred some 1,500 and 2,500 years ago. (c) A similar undated notch at +12.3 m, may correspond to a Late-Pleistocene shoreline. South of the Orontes Delta, near the Turkish–Syrian boundary. (Photo A739, May 1988)

will consequently tend to subside and move towards the center of the vanishing load. Part of these glacio-isostatic movements are elastic, and thus contemporaneous with loading and unloading. Due to the viscosity of the earth material, however, part of the movement will continue for several thousand years after the loading or unloading has stopped. Although most of the ice in Canada and Fennoscandia had disappeared between about 8,000 and 6,000 years BP, large vertical movements of uplift and subsidence continue in these areas today.

It was during the second half of the nineteenth century, when the postglacial coastal record in Fennoscandia became accurate enough to yield a persuasive description of dome-like uplift, that the concept of glacio-isostasy could be asserted by De Geer, as had been done at about the same time for Lake Bonneville, in the United States, by Gilbert. During deglaciation, meltwater from ice sheets produces a

considerable load on the ocean floor (on the order of 100 t/m^2 for a sea-level rise of 100 m), so that the sea bottom subsides (hydro-isostasy). In the upper part of gently sloping continental shelves, or in shallow seas where the postglacial water depth is less than the global change in sea level, the meltwater load will vary according to the local topography and bathymetry, generally increasing gradually towards the open sea. In this case, the hydro-isostatic constraints will produce a lithospheric flexure with a typical seaward tilting.

Small islands which rise steeply from the deep ocean floor are often considered as potentially good place to measure eustatic changes because, as discussed by Bloom (1971, p. 371) they “will be warped downward by the water load around them, but because the entire deep ocean floor is depressed, the volume of the ocean basin increases and sea level with reference to an island, or to a hypothetical buoy moored in deep water, should not change because of the isostatic deformation.” In equatorial and tropical regions remote from the ice sheets, hydro-isostatic subsidence may be partly compensated by “equatorial ocean siphoning” (Mitrovica and Peltier 1991). This mechanism, driven by the subsidence of those portions of the glacial forebulges which exist over oceanic regions, acts to draw water away from most equatorial and tropical regions, in order that the oceans maintain hydrostatic equilibrium, making some emergence possible. For this reason, slight Holocene emergence is frequent in tropical oceanic islands. This is the case, for example, for most Pacific atolls (Fig. 5) in spite of their long-term trend towards thermo-isostatic subsidence.

Some geophysical models, based on the mathematical analysis of the deformation of a viscoelastic earth produced by surface loads, have been developed during the last two decades. They have been able to mimic, with an accuracy on the order of a few meters, sea-level changes reported from the field in various areas. These models, which assume a simplified earth structure and a melting history for all the continental ice loads which existed at the time of the last glacial maximum, have demonstrated that the rate, direction, and magnitude of crustal movements must have varied from place to place and, therefore, that no region can be considered as tectonically stable. Though they cannot replace field observation, these models have also been useful in providing a first-order approximation of the deglacial and postglacial sea-level history in areas where no field data are available.

Sediment-Isostatic Neotectonics

Weathering and gravitational forces cause continuous transfers of sediments in a one-way direction, from the continents to the oceans. This implies isostatic adjustments, with predominant uplift on the continents and subsidence in nearshore basins affected by rapid sedimentation. At the mouth of great rivers, when the rate of fluvial input overtakes the rate of



Tectonics and Neotectonics, Fig. 5 Late-Holocene *scleractinian* corals have been left emerged in growth position, at the top of a pinnacle in the lagoon of Takapoto Atoll (Tuamotus), by a slight fall in the relative sealevel of probable hydro-isostatic origin (Pirazzoli and Montaggioni 1988). (Photo 5478, March 1980)

sea-level change, deltaic sequences may tend to accumulate, especially on microtidal coasts. Recent sediment accumulation is affected by compaction, which produces subsidence, but is also expected to form on the lithosphere surface an isostatic depression under the load and a marginal rim slightly rising at some distance. However, remaining in an area where general subsidence predominates, such rims may only appear as an area of lesser sinking. Lateral displacement of major delta branches have often occurred during the late Holocene; they imply, after some time, also a migration if their isostatic peripheral bulge.

The purely isostatic component is generally difficult to estimate, because observed rates of lowering include also sediment compaction, possible faulting, and for the past millennium also anthropogenic influences, such as conversion of wetlands to agricultural fields, river channelization,

pumping, and withdrawal of water and diversion of water flow for irrigation. What can be measured is only the total amount of vertical displacement that has occurred during the Holocene delta formation, that is, since the postglacial sea-level rise began to decelerate, generally from about 8,000 to 6,500 years BP (Stanley and Warne 1994). This vertical displacement varies locally, usually reaching a maximum on the outer delta plain close to the present coast.

Structural Faulting and Folding

Faults (and folds, which can develop ahead of blind faults propagating from depth towards the surface) are generally the main object of study for most structural tectonicians. In coastal areas, however, at least for vertical displacements, the most important neotectonic indicator is sea level. With high-tide shoreline marks, sea level leaves evidence of an altimetric datum all along the coast. An active fault may cross a rocky coast, leaving recognizable displaced high-tide shorelines on both sides (Fig. 6). Such an occurrence is precious, because it may enable one to measure very precisely the vertical displacement between the footwall and the hanging wall of the fault, that is, the maximum amount of displacement when we are dealing with a tilted crustal block. In even more favorable cases former high-tide shorelines may be preserved, at least over a certain distance, at various levels on the fault plane; it will then be sufficient to map them along the coast to reconstruct tilting patterns and, if the high-tide shorelines can be dated, vertical displacement rates. Some caution is necessary, however; high coastal cliffs which are abnormally straight or gently curved are commonly suspected of being fault scarps. When uplifted evidence of former high-tide shorelines is missing from the cliff, the footwall of the fault may be underwater, concealed by sediments, or eroded, and this makes estimations of vertical movements hypothetical. But straight high cliffs may also result from differential erosion having removed a weaker seaward rock formation, without any tectonic implication. Many mappable faults may also have been formed by geologically ancient deformations, under tectonic regimes and stress situations long abandoned.

In most cases, active faults do not cut the coastline, but remain at a certain distance from it, on land or offshore; marks of uplift or subsidence indicated by former high-tide shorelines are therefore likely to indicate a lesser vertical displacement than at the fault scarp and a more complete survey is necessary before interpreting the vertical displacement observed.

Example of a paleotectonic coastal scarp: northernmost Chile. Between Arica and Iquique, except locally, an exceptionally high cliff directly drops into the sea, and is still retreating under wave action. In the surroundings of the

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Fig. 6 The surface of Late Pleistocene raised marine terraces has been displaced by a recent fault near Kupang, West Timor, Indonesia. (Photo B266, August 1988)



Cerro Punta Madrid ($18^{\circ}57'S$, $70^{\circ}18'W$) it reaches a height of about 1,000 m. This cliff derives from a large system of north–south oriented, *en échelons*, normal faults which formed at the end of the Oligocene epoch and whose scarps retreated under marine erosion during a Middle to Upper Miocene transgression. South of Iquique, the scarp is an abandoned cliff no longer receding under wave action; at its foot lies a shore platform, sometimes wider than 1 km, which has been uplifted to 50 m above present sea level (Paskoff 1996).

Example of a neotectonic coastal scarp: southwestern Calabria (Italy). Calabria has been affected by strong uplift movements during the Quaternary. At least 12 Pleistocene levels of marine terraces have been identified, reaching as much as 1,350 m in altitude. Near Reggio Calabria, Last Interglacial marine deposits with the guide fossil *Strombus bubonius* at 157 m suggest that the average uplift rate exceeded 1 mm/year during the last 125 ka. Near Palmi, the steepness of the coastal relief (Fig. 7) and the narrowness (only a few hundred meters) of the continental shelf can be ascribed to active faulting, which is leaving abrupt scarps on the gneissic rock formations.

Example of neotectonic coastal folding. Along the southernmost coast of Taiwan (Hengchung Peninsula), the inner edge of the mid-Holocene marine terrace is 10–15 m high, but its elevation gradually increases westwards, deformed by undulation movements. Its altitude ranges from 5 to 36 m along a distance of less than 10 km (Liew and Lin 1987).



Tectonics and Neotectonics, Fig. 7 The active fault scarp near Palmi (Calabria, Italy). (Photo E722, April 1996)

Rates of Vertical Neotectonic Movements

Because of the various potential tectonic processes, rates of vertical movements in coastal areas have varied much in space and time.

Over the long term (10^6 – 10^5 year), structural tectonics is indeed the predominant factor, with rates reaching several millimeters per year reported from tectonically very active areas such as near Ventura, California, where Pleistocene marine terraces have been uplifted at average rates exceeding 8 mm/year (Lajoie 1986), with even much higher rates reported from the land interior (about 10–15 mm/year estimated for the crest of the Ventura anticline during the last few hundred thousand years (Wallace 1986, p. 7)). At Huon Peninsula, Papua New Guinea, uplift of up to 3 mm/year during the last 124 ka has been demonstrated by Chappell (1974). Evidence of well-preserved Quaternary shorelines is, however, rare in areas of rapid uplift, because strong erosion rates most often effaced most ancient marine marks. At the same timescale, volcano-isostatic factors can be estimated to have been on the order of a few millimeters per year for subsidence and one order of magnitude less for arching uplift; thermo-isostatic vertical displacement rates usually remain limited to the order of a fraction of millimeters per year: average subsidence rates have been estimated at 0.2 mm/year during the last 60 Myr in the Marshall Islands and 0.12 mm/year since the Pliocene in Mururoa Atoll (Tuamotus). Glacio- and hydro-isostatic factors cannot be assessed at the timescale of 10^6 – 10^5 year.

During the last 20 kyr, the most important vertical displacements have been of glacio-isostatic origin; near the center of the former Fennoscandian ice sheet, uplift since the last glacial maximum has been estimated at about 800–850 m, that is, at 40–42 mm/year on average (Mörner 1979); since the beginning of the Holocene, uplift has been about 284 m, that is, 28 mm/year on average, decreasing exponentially from over 50 mm/year at that time to about 9 mm/year at the present time (Pirazzoli 1991). Residual present-day uplift is therefore almost 1 m/century. In Canada, with a marine limit close to 300 m in elevation near the eastern part of Hudson Bay (Andrews 1989) maximum uplift values are probably of the same order as in Fennoscandia.

Hydro-isostatic displacements are also significant over the timescale of 10^4 year. They can be estimated on the order of about one-third of the sea-level change, that is, some 40 m in deep-sea areas for an eustatic change of 120 m (i.e., 2 mm/year on average since the last glacial maximum) and less in shallow coastal areas.

At the timescale of 10^4 year, structurally tectonic displacements can be in some cases on the same order as glacio-isostatic ones: on the east coast of Taiwan, marine evidence suggests an average uplift rate of about 8 mm/year since 14 kyr ago near Tulan (Pirazzoli et al. 1993) as well as since 4,000–3,000 kyr ago near Hualien (Konishi et al. 1968) (Fig. 8). Because the amount of glacio- and hydro-isostatic displacements has been decreasing exponentially with time after complete melting of the ice caps, movements of structurally tectonic origin have become predominant and easily recognizable in several areas during the last few thousand

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Fig. 8 In situ coral reefs about 3 kyr old, uplifted up to 25 m above sea level, are visible (arrows) above a cliff at Hualien, east coast of Taiwan. (Photo B805, January 1990)



years. Very rapid rates of vertical displacement of thermo-isostatic origin can be reached in volcanic areas during relatively short periods. The most spectacular example is probably Iwo (Sulphur) Island, a volcano situated 1,200 km south of Tokyo, where over 20 steps of marine terraces occupying nearly the whole area of the island (used as military airstrips by the Japanese during the last world war) have been recognized up to 120 m above sea level. The average uplift rate was here of more than 100 mm/year over the last several centuries, reaching 200 mm/year in some parts of the island (Kaizuka 1992). In the case of the Pozzuoli area mentioned above, average vertical displacement rates have reached even faster peaks, such as the uplift rate of up to 800 mm/year recorded between 1968 and 1984.

Rates of subsidence in delta areas induced by sediment compaction and sedimento-isostasy is often significant over the long term. At the shelf edge of the eastern Mississippi delta, for example, the long-term averaged subsidence rate since the last glacial period has been estimated to at least 1 mm/year (Stanley et al. 1996). For the four largest modern depocenters in the Mediterranean, long-term mean lowering rates have been reported of 4–5 mm/year from the Ebro Delta, 7 mm/year at the mouth of the Grand Rhône Delta, possibly from 1 to 3 mm/year from the Po Delta, and 4–5 mm/year from the Nile Delta (Stanley 1997). The best-known example of land subsidence produced by sediment compaction, accelerated by man-induced drainage during the last centuries, is indeed that of the Netherlands, where the soil level may be as low as 6 m below sea level. More recently, man-induced land subsidence following oil and natural gas extraction or groundwater exploitation has taken place during the last century in many coastal plains of estuarine, delta, or lagoonal areas: the greatest measured land subsidence (a maximum of 4.6 m) is probably that reported from the Tokyo Lowland region, where an area of about 70 km² supporting more than half a million inhabitants, is below mean sea level. Significant amounts of land subsidence have been reported also from the Po Delta (3.2 m), Houston-Galveston (3.0 m from 1906 to 1987), Shanghai (2.7 m), Tianjin (2.5 m), the southwestern part of Taiwan (2.4 m), Taipei (1.9 m), and Bangkok (1.6 m). In several big cities built in coastal plains of developing countries (Jakarta, Hanoi, Haiphong, Rangoon), where detailed measurements are lacking, the probable occurrence of land subsidence is revealed by increasing difficulties in drainage which produce more frequent flooding.

Among the many possible causes of coastal change and evolution, tectonics and neotectonics have been indeed major controlling factors, virtually at all timescales. For the present time, however, a new tectonic cause – man-induced land subsidence – is becoming significant in many densely populated or industrial coastal areas, with short-term impacts

which may exceed those of all other natural causes occurring in the same area.

Cross-References

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- [Faulted Coasts](#)
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- [Physical Models](#)
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Thalassostatic Terraces

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A term derived from Greek roots, the expression “Thalassostatic” obtains its prefix from “thalassic,” having

to do with the sea, and its suffix “static” referring to its equilibrium state. A fluvial terrace is said to be thalassostatic when it is the product of a former sea level that caused flood-stage sediment load to accumulate as an alluvial surface. Its seaward limit is usually determined by a beachridge or beachridge plain.

The actual term itself was coined by Zeuner (1945), who saw that an estuary created by downcutting during a Pleistocene low sea-level state would become aggraded during a following interglacial phase. Even earlier, Ramsay (1931) pointed out that alluvial terraces of this sort would provide clearer evidence of former eustatic levels than strictly marine terraces that are more liable to erosion or other factors. Clayton (in Fairbridge 1968, p. 142) mentions the actualistic evidence of modern aggradation behind dams. This filling will extend only a short distance upstream where the gradient is changed. Zeuner believed that gravel terraces of the lower Thames in England must be solifluction products of glacial-phase stream loading, thus having nothing to do with thalassostatic events.

In the classic work of de Lamoignon (1918) on the Somme Valley in northern France, the concept of eustatic control was first presented, but erroneously extended to the entire fluvial terrace system. The Dutch worker, Brouwer (1956) considered the question of thalassostatic terraces in general and showed that only the lowest sectors were appropriately so-named. Earlier, Pleistocene examples would be largely obliterated by erosion.

In the world's major deltas like the Mississippi and the Rhine, a quasi-stable tectonic fulcrum develops, with subsidence on the seaward side, uplift on the landward side. Thus the fluvial terraces are stacked progressively lower and in stratigraphic sequence downstream. On the Mississippi the fulcrum area is above Baton Rouge, on the Rhine about at Nijmegen. However, the exact fulcrum point shifts up- or downstream depending upon the current eustatic level. A problem develops, however, with these large rivers in their propensity to major flood events, their discharge being greatly amplified by the large area and multiple sources of their drainage basins. Such events are likely to overtop levees in the deltaic regions. These so-called “avulsion” breaks are commonly matched by the drying out of other distributary channels, so that a rather complex stratigraphy develops (Törnquist 1994; Berendsen 1995). Nevertheless, a strongly cyclical pattern is discernible. Where an extensive data source has been provided by detailed geophysical transects, as in the Mississippi delta, it can be seen that layers of Holocene transgressive facies are alternating with regressive events when channels deepen and their outlets shift seawards. In contrast, the transgressive intervals are marked by sandy chenier ridges (sand from longshore sources) that form

distinctive markers in the otherwise muddy or peaty deltaic facies. Compaction of these water-saturated substrates leads to areas of accelerated subsidence, as shown by very variable tide-gauge evidence, that can be misleading for the delta as a whole, which has a very modest subsidence rate. With each transgression the thalassostatic terrace building shifts upstream, but with the renewed downcutting after avulsion an entirely new delta-lobe is likely to evolve (Lowrie and Hamiter 1995).

A pioneering attempt to relate thalassostatic eustasy to the rivers of Borneo (Kalimantan) was made by Smit Sibinga (1953). A detailed glacial-age drainage network extends over much of the Sunda Shelf according to the submarine mapping by Molengraaff and referred to as the “Molengraaff River System” by Umbgrove (1947). With each successive eustatic rise thalassostatic terrace deposits must have lined these former waterways. The eustatic curve is not a sine-wave, however, but is a strongly fluctuating one, so that there are extended intervals of pause or reversal, during which there would have been accelerated thalassostatic sedimentation.

Best known and well dated of these interruptions was the Younger Dryas event which lasted approximately 1,000 year, 11,740–10,740 cal. BP with abrupt transitions, both at its commencement and terminations. Eustatic sea level at that time fell to about 30 m below present mean sea level. Its coastline is particularly well marked by the presence, in the warmer latitudes and along the more stable coasts, of an abrupt line of reefs or distinctive beach deposits. A wave-resistant beachrock facies is a distinctive marker where carbonate sands permit rapid lithification. In some sectors carbonate sands led to the development of littoral dunes, which also become rapidly lithified to form “eolianites.” Today these eolianites form reefs that are well known to fishermen (e.g., “Twelve Fathom Reefs” of Western Australia).

Other sea-level fluctuations were marked in the deep ocean by the “Heinrich events” (at 3–7 kyr intervals) which mark iceberg distribution in the N.W. Atlantic. Heinrich events are marked by climatic extremes, both positive and negative, thus raising the chicken-and-egg problem. Analogous eustatic fluctuations are indicated by the thalassostatic data. Sudden warming of the water around Greenland, such as occurred in AD 1912, accelerates iceberg calving (with disastrous effect on RMS *Titanic*). In this case, however, it was not so much climatic, as tidal, because 1912 marked the largest lunisolar tidal extremes in over 500 years (Wood 2001). Thalassostatic terrace building (and its erosion) thus relates to movement of relative sea level (RSL). The bed of a river on reaching the sea is nicely adjusted to velocity and bed-load, which dictate the

equilibrium water depth at the river mouth, subject to tidal characteristics and seasonal changes such as storminess and monsoonal shifts.

The equilibrium status may be modified by one or more of several processes: (a) Tectonic, which may be sudden as a result of an earth-quake, such as the historic shifts in 1855 and 1931, of Wairapa (Wellington) or Napier in New Zealand or the subsidence of parts of Yokohama in Japan. It may also be secular as a result, for example, of glacioisostatic rebound such as observed today at Stockholm in Sweden (at 5 mm/year) or at Great Whale River in Quebec (at 8 mm/year). (b) Eustatic and/or climatic; these are related but partly independent, although not easily distinguished. When a dominant wind system changes, it affects also RSL as observed, for example, in the Hudson Bay, in the Gulf of California and in the eastern Gulf of Mexico, all being partly related to the 11 year sunspot cycle and the 18.6 year lunar cycle (Fairbridge 1992). It is also observed on a 6-month basis in the Red Sea and Arabian Sea and Bay of Bengal, a consequence of the Asiatic monsoon reversal. Longer cycles, of the order of 45, 90, and more years are observed in the arctic and subarctic latitudes, relating in part to the polar anticyclone and the magnetic pole in its periodic shifts from the longitude of Greenland to that of eastern Siberia, and back again.

For whatever reason, a fall in RSL leads to a shallowing of the debouching river mouths. Two possibilities ensue. If the fluvial discharge is linked, as it usually is, to climate fluctuation, a decreased discharge coinciding with a shallowing of the stream beds, then the result will be a “siltation” of local harbors. Around the Mediterranean in Roman times (first century AD) there was widespread siltation and as a result many Roman docks, harbor facilities, and fish tanks had to be reengineered. Some harbors, as in Turkey and farther afield in India (entrance of the Nerbada River) were abandoned altogether, or the city shifted somewhat downstream (Kayen 1997).

The opposite condition is associated with a systematic rise in RSL. This occurred, for example, in the post-Carolingian (after Charlemagne) global warming that persisted until about AD 1300. River courses were backed up and deepened. Roman-era dock or defensive facilities were drowned. An example of the latter can be seen at Portchester in the south of England.

Cross-References

- [Archaeology and Sea-Level Change](#)
- [Beach Ridges](#)

- [Cheniers](#)
- [Eolian Processes](#)
- [Late Quaternary Marine Transgression](#)
- [Offshore Sand Banks and Linear Sand Ridges](#)

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Thermal Expansion

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Synonyms

Steric changes

Definition

Thermal expansion. The expansion of the ocean water column due to heating of the water. The opposite occurs at cooling of the water.

Steric effects. Expansion or contraction of the oceanic water column due to changes in temperature and/or salinity.

Introduction

Steric effects were realized as secondary factor to glacial eustasy in the science of Holocene sea level changes (Fairbridge 1961). According to Schofield (1980), the steric changes in sea level were driven by changes in salinity, rather than changes in temperature. With the initiation of the IPCC project (IPCC 1988/1990), “thermal expansion” became a fashionable factor in sea level research, and it is nowadays commonly addressed in sea level papers. Often, however, it is used without basic anchoring in physical and oceanographic facts.

Origin of Thermal Expansion

Thermal expansion (as well as thermal contraction) leads its origin to the simple physical fact that matter expands when heated (Tipler and Mosca 2008; TETB 2017). The amount of expansion has a direct relation to the column heated and the distribution of this heating along the column.

Thermal expansion is easily expressed in the expansion of a copper rod when heated. A classical practical demonstration is the 1 cm open joining of train rails to prevent curvature due to thermal expansion at hot climate.

In the oceans, the problem is, of course, much more complicated.

- *Pro primo*, the water depth – i.e., the water column susceptible for expansion – varies from zero at the shore to about 6000 m in the abyssal.
- *Pro secundo*, the heating differs with depth: the surface water (0–100 m or 0–300 m) is much easier to change than the deeper water, and water below 2000 m seems hardly affected at all.
- *Pro tertio*, the quantification of thermal expansion has been quite differently expressed by different authors.
- *Pro quadro*, the vertical expansion and possible horizontal flow have often been treated in ways not anchored in observational facts.

Vertical Expansion of the Ocean Water Column

The quantification of ocean thermal expansion differs quite much. A basic estimates were given by Nakibogul and Lambeck (1991), who found that this effect may rise sea level by 5 cm up to year 2100. According to the

Thermal Expansion, Fig. 1

Relations among water expansion, water column affected, and heating by increased temperature (Modified from Möner 2011). Each variable has its strict frames. On a regional basis, the heating can hardly exceed a couple of degrees. Approaching the coast, the expansion of the water columns progressively decreases, becoming zero at the shore (*black dots*). Therefore, coasts are not affected by thermal expansion of thesea

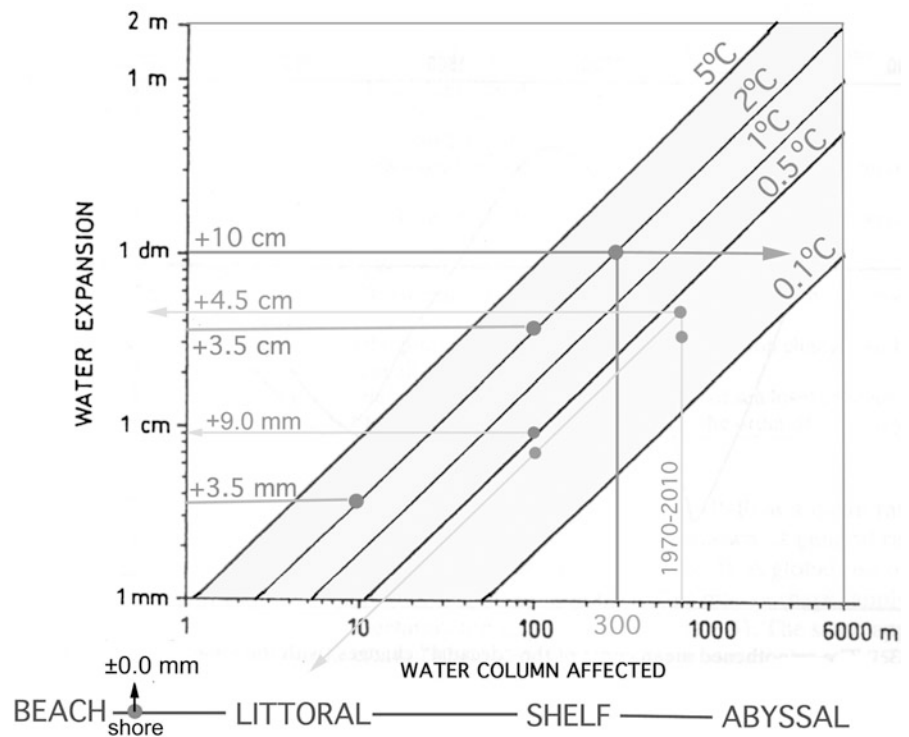
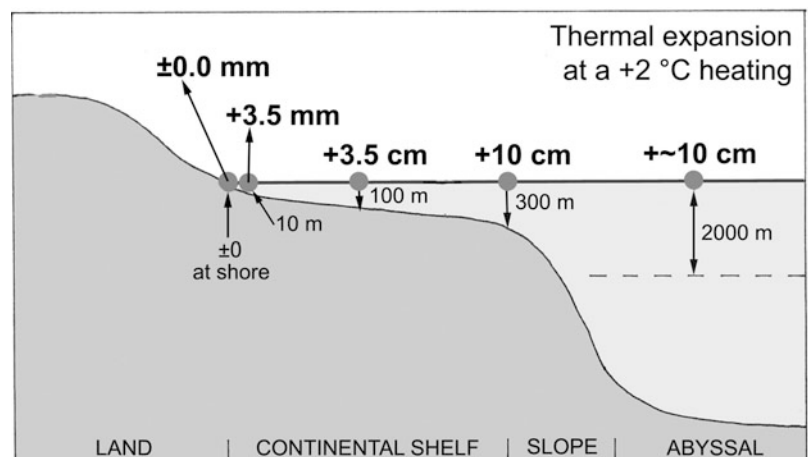
**Thermal Expansion, Fig. 2**

Illustration of the depth dependence on thermal expansion at an assumed $+2^\circ\text{C}$ rise in ocean temperature (which is exceptionally much and only possible for the oceanic surface layer). At the shore, the water column is zero and hence there can be no thermal expansion, and no water is flowing coastward from the rise seaward



IPCC (2013), thermal expansion amounted to 0.8 mm/year for the period 1971–2010. Möner (2011) concluded “an expansion (sea level rise) of more than 5–10 cm seems hardly likely.”

Figure 1 gives the relations among heating, water column, and expansion (Modified from Möner 2011, 2013).

Earlier, the upper 300–700 m of the ocean were claimed to have become heated by 0.3–0.4 $^\circ\text{C}$ in the period 1970–2010, corresponding to a 4.5 cm expansion. New data by NOAA (2017) suggest that the top 100 m has been heated by 0.55 $^\circ\text{C}$ in the period 1955–2015,

corresponding to 9.0 mm expansion. Both values are marked in Fig. 1.

Horizontal Flow of Expanded Water

The dynamic sea level surface has a strong relief with departures from the gravitational geoid surface of several decimeters up to some meters (e.g., a 5 m high over the Gulf Stream, and 30–40 cm evaporation low in the Indian Ocean). Therefore, there are no reasons why a thermal

expansion off the coast would flow towards the coast and rise sea level.

This implies that the effect of thermal heating expansion becomes progressively smaller as the depth (water column) becomes shallower towards the coast, and at the shore the effect is zero, and no water will flow towards the coast because of expansion seaward.

This is illustrated in Fig. 2 with an assumed +2 °C heating of the ocean column (Modified from Mörner 2011 and 2013).

Summary

Thermal expansion is an important factor. It is often exaggerated, however. Heating of the oceans is confined to the upper 700–2000 m. This heating is always cyclic (e.g., the 60-year cycle), never linear over multidecadal and longer periods.

The contribution to sea level changes up to year 2100 is estimated not to exceed 10 cm.

Cross-References

► [Changing Sea Levels](#)

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Tidal Creeks

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Is a tidal creek an estuary? Stamp (1966) in his *Glossary of Geographical Terms* quotes the OED definition of creek as, “a narrow recess or inlet in the coastline of the sea, or the tidal estuary of a river; an armlet of the sea which runs inland in a comparatively narrow channel” The *Dictionary of Geological Terms* published by the American Geological Institute (1976) defines a creek as, “a small inlet, narrow bay, or arm of the sea, longer than it is wide, and narrower and extending farther into the land than a cove.”

Geomorphically, the essential characteristics of a tidal creek are that they are relatively long and narrow, are shallow, and exhibit tidal water level fluctuations and weak tidal currents. They are relatively small-scale landforms with a low hydrodynamic energy environment without significant wave action or strong current action (Fig. 1). Typically along its banks the tidal creek is well vegetated. The friction from the banks and the length of the creek ensure that the tidal wave within the creek is hypo-synchronous (reducing tidal range upstream). Such conditions facilitate sediment deposition, so tidal creeks exhibit active deposition and infilling over a time frame of decades. Tidal creeks may drain to the open coast of an enclosed coastal sea or bay, or may be part of a larger estuary or delta system. They also occur at the distal ends of arms of drowned river valley (ria) harbors where tidal creeks drain sectors of mangrove stands and/or salt marsh. In tropical areas, tidal creeks and distributaries create a dense network of small-scale drainage through the mangrove forests, while a similar reticulated plexus of creeks occurs in temperate salt marsh (Guilcher 1958). Pethick (1984) notes that salt marsh creek systems have a very high ratio of total creek length to drainage area – on the order of 40 km/km². Tidal creeks also occur in sinuous meandering forms draining extensive intertidal mudflats. However, tidal creeks do not tend to occur draining to open coast high-energy sandy beaches.

Often the creeks extend as the headwaters of a large estuary into the surrounding drainage basin, and on occasion carry freshwater discharges and episodic floodwaters from surrounding catchments. Thus the tidal creeks are frequently the mixing zones of fresh and saline tidal water, and exhibit an enhanced vertical salinity structure due to the low-energy conditions. Accordingly, they are also favored zones for flocculation processes, so that deposition of muds is typical and ongoing within tidal creeks. Unfortunately, tidal creeks often

Terry R. Healy: deceased.

Tidal Creeks, Fig. 1 Tidal creek in mangroves showing the narrow shallow channel and low hydrodynamic energy conditions



tend to become a receptacle for human waste and rubbish, and are frequently reclaimed.

Cross-References

- [Estuaries](#)
- [Mangroves, Ecology](#)
- [Mangroves, Geomorphology](#)
- [Muddy Coasts](#)
- [Ria](#)
- [Salt Marsh](#)
- [Tidal Flats](#)
- [Vegetated Coasts](#)

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Tidal Datums

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Introduction

A sea-level datum is a surface constructible statistically from sea-level observations that is used as a reference for

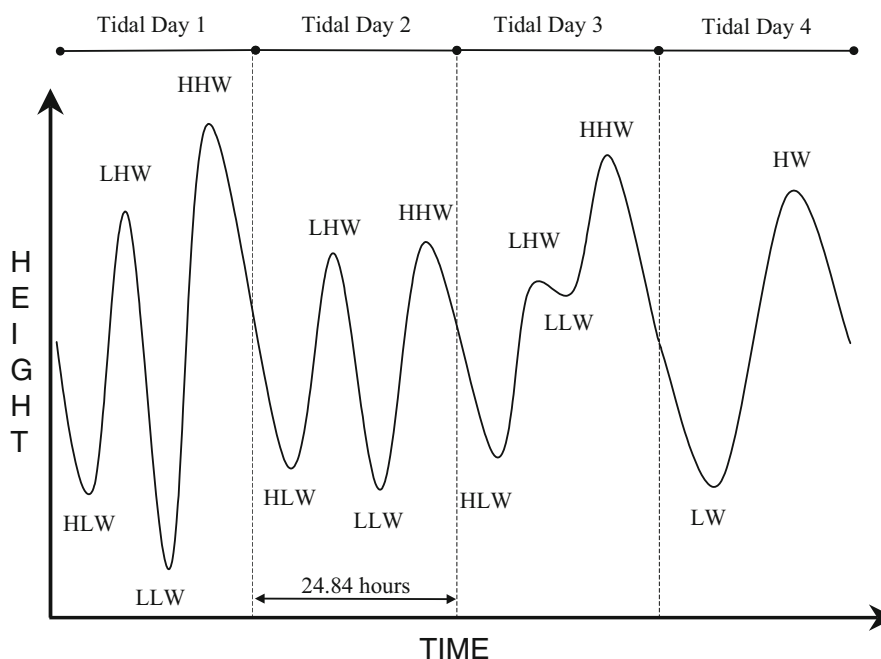
measuring and describing vertical positions near the earth's surface. At official intervals, the elevations of sea-level datums are revised to reestablish their relationship to a new mean water level. A multitude of sea-level datums are in use. Each has its own advantages and disadvantages with no universally superior choice. Preferences depend on location, purpose, and past practices. Focus here will be on tidal datums defined at different phases of the tide; for example, high water and low water (Fig. 1). Discussion of tidal datums illustrates some of the uses, practices, and limitations common to the broader class of vertical datums.

Scope

Details in the operational definitions of tidal datums evolved as concepts matured, needs for precision increased, field methodologies improved, and new legal issues arose. In this discussion of broad tidal datum issues, operational details may be sacrificed to provide an overall understanding of the basics, at least as understood by a coastal morphologist limited in practice to North America. Specific details continue to evolve with advances in measurement technology and changing society needs. Prescriptions of such details are left to official documents.

Historical Developments

The astronomical tide causes sea level to vary on daily, monthly, and longer cycles. These deterministic cycles are superimposed on more random water level fluctuations related primarily to weather. Normally, the reference point for repeated measurements should be stable, that is, anchored to the system containing the objects measured. What factors



Tidal Datums, Fig. 1 Tide phases labeled on a stylized tide curve. When two high-tide phases can be distinguished in the tide curve during a single tidal day (24.82 h), the higher of the two is called higher high water (HHW) and the other lower high water (LHW). When two low-tide phases are discernable in 1 day, the higher of the two lows is called higher low water (HLW) and the other lower low water (LLW). When

only one high and one low-tide phase appear (e.g., Tidal Day 4), the high phase is called high water (HW) and the low is called low water. The record from sites that usually have only a single high and low per day will periodically transition in and out of multiple- cycles-per-day episodes that can have distinctly different phases like Tidal Day 2 or less distinct phases, more like Tidal Day 3 above

explain widespread adoption of a continually fluctuating water surface as a vertical reference? Early communities developed at sea ports where knowledge of the water depths in navigable channels was vital for the economy and defense. Soil productivity [EBH1] was another fundamental concern related to height above water because of altitudinal climate gradients if not more directly through frequencies of flooding. Referencing elevations to water levels was, therefore, of keen interest to early civilizations. Sea level also provided a common reference elevation applicable widely throughout the world. It could be reestablished with little loss of accuracy near coasts and without the need for long-distance surveys or recover of any destructible marker (e.g., benchmark). Fluctuations of the sea surface (and hence the zero reference) were probably small relative to the precision required to make decisions, especially for civilizations developing around the Mediterranean Sea where the range of tidal fluctuations is small compared with a typical range in the ocean. As coastal property ownership disputes increased, navigation clearances tightened, and tidal fluctuations became recognized (though not yet understood), comparison of elevations with respect to different sets of sea- level observations became more problematic. Later recognition of a broad spectrum of sea-level changes required further adjustments to allow tidal datums to persist as widely used references. Continuing advantages of referencing to sea level, instead of a fixed

datum, include: conveniences related to historical precedence, natural differences in human use of property above and below sea level, lateral continuity (at least along coasts), apparent simplicity, lack of geographic bias in definition, and the greater relevance of mean water levels over any truly level surface for most hydrologic and navigation considerations.

Because concerns of navigation and coastal engineering are closely tied to sea level, the reference point in these fields should change with long-term changes in mean sea level. For endeavors in which sea level is irrelevant, a stable nontidal datum may be preferable. For this reason, and to better understand tidal datums themselves, some discussion of nontidal datums is included.

The name of the organization having responsibility for official datum and charting standards within the United States changed with time from the Survey of the Coast, to the Coast Survey, Coast and Geodetic Survey, National Ocean Survey and, now, the National Ocean Service (NOS). The NOS's authoritative references on present and historic coastal boundary determinations are Reed (2000) and Shalowitz (1962, 1964). Kraus and Rosati (1997) summarize procedures for determining shoreline positions for coastal engineering. Prescription for US Army Corps of Engineer surveys of waterways and additional guidance on tidal datums can be found in USACE (2002). All five references can be found on the World Wide Web (www).

Methods of Tidal Datum Determination

Statistically averaging sea-level measurements is the process by which an unsteady surface is fixed to serve as an unambiguous and repeatable datum. There are a number of ways to average sea level, giving rise to a number of different tidal datums. For an unambiguous datum definition, the frequency of the measurements and the time span over which they are to be averaged must be specified. For any tidal datum, the manner of relating the measurements to a clearly defined tidal phase (Fig. 1) must also be specified. Breakdown in any one of these fundamental requirements, causes problems as we shall see.

Astronomical Cycles and the Tidal Response

Astronomical tides are the sea's response to gravitational forces of the moon and sun (and to a lesser extent other celestial bodies) that vary over the earth's surface. These forces also vary with time as relative astronomical positions continually change over a wide spectrum of cycles. Procedures to calculate the local elevation of each of six standard tidal datums (upper portion of Table 1) follow similar official

algorithms regardless of which of the six is selected. Selection would typically be based on location and purpose, for example, whether it involves resolving legal disputes, promoting navigation, assessing environmental resources, or analyzing geophysical phenomena.

The longest cycle conventionally damped by tidal averaging (the Metonic cycle) has a period of about 18.61 years. The NOS selects a specific 19-year time interval, called the *National Tidal Datum Epoch* as the official span over which tide measurements are averaged to obtain the current tidal datum elevations throughout North America.

Gauges

The NOS presently maintains about 189 *primary tide gauges* where hourly water levels have been recorded continuously for more than 19 years. *Tidal benchmarks* are physical monuments maintained along the coast and to which tidal datum elevations have been surveyed (see “► [Geodesy](#)”). Three to five bench marks are tied to NOS tide gauges by precise surveying and monitored to detect possible gauge disturbances.

Tidal Datums, Table 1 A few North American vertical datums used by coastal engineers and scientists

Acronym	Name and definition	Origin, use or advantages
<i>Tidal datums (based on gauge records)</i>		
MHHW	Mean higher high water Mean of all higher high waters ^{a, b, c}	
MHW	Mean high water Mean of all high water heights (i.e., HW, HHW, and LWH, see Fig. 1) ^{a, b, c}	Frequently used boundary separating private from state lands. See text. Landward limit for US Corps of Engineers jurisdiction over navigable waters
MTL	Mean tide level Mean of MHW and MLW datums ^{a, b, c, d}	More descriptively called half-tide level
MSL	Mean sea level Mean of hourly water surface heights ^c	Maintained US-wide by the National Geodetic Survey as “the most practicable and stable datum for general engineering,” Shalowitz (1962)
MLW	Mean low water Mean of all low water heights (i.e., LW, LLW, and HLWs, see Fig. 1) ^{a, c, d}	Use of the synonymous term, mean low tide, is discouraged by NOS
MLLW	Mean lower low water Mean of all lower low water heights ^{a, c, d}	Principal chart datum for all of U.S. since 1981 Used by six states as boundary separating private and state lands
<i>Orthometric vertical datums (based on topographic surveying)</i>		
NGVD 29	National Geodetic Vertical Datum 1929	Earlier standard geodetic datum. Based on first order survey nets of the United States and Canada fit to MSL at 26 tide gauges
NAVD 88	North American Vertical Datum 1988	Official vertical datum for all of North America since June 24, 1993. An upgrade of NGVD 1928 based on more measurements over a larger area and independent of any tide gauge measurements
<i>Three-dimensional datums (based on measurements from space)</i>		
NAD 83	North American Datum 1983	An upgrade using space-based measurements of the previous standard, NAD27
WGS 84	World Geodetic System 1984	US Defense Mapping Agency equivalent of NOAA's NAD 83

^aEach stage of the tide in the time series to be averaged must, by official convention, differ by at least a “0.10 ft” and “2.0 h” from the adjacent measurements in the series. The series usually extend over a specified National Tidal Datum Epoch, but sometimes over a shorter stated duration

^bA high water is a maximum in a tide curve reached by a rising tide. The higher of two unequal high waters during the same day (24 h 50 min the period of the M tidal component) is the higher high water. The other high water is the lower high water

^cDefinitions from US Department of Commerce Tide and Current Glossary (Hicks 1989)

^dA low water is a minimum in the tide curve that is reached by a falling tide. The lower of two unequal low tides during the same day is the lower low water. The other is the higher low water

Elevations are established at a larger set of *secondary gauges* by comparing their shorter tidal records (less than 19 years, but more than 1 year) with simultaneous measurements from nearby primary stations. Tidal datums can be reestablished by leveling short distances from tidal benchmarks. Numerical modeling of the tide can be used to extend datums longer distances between gauge sites including inland across the continent.

The elevation of water levels (and, therefore, tidal datums) in estuaries and coastal lagoons is typically higher than along adjacent open coasts. This super elevation is primarily due to freshwater flowing into the bay and non-linear friction in tidal currents. Adjacent to inlets, the transition is smooth, but relatively steep. Therefore, primary gauges are often located away from their immediate vicinity (see “► [Tide Gauges](#)”).

Tidal Datum Elevation Algorithms

To calculate the elevations of a particular tidal datum at a primary tide gauge site, select the elevation measurements associated with the chosen tide phase (e.g., high tide) from the set of hourly data that covers the current National Tidal Datum Epoch. Official rules specify which hourly readings are associated with the chosen phase of the astronomical tide, but other factors (primarily weather) contributed to the water level measured at that time. Arguments have been made, especially for regions where the wind is strong, to loosen the required association between astronomical tidal phases and selection of which observations to average. Following the official rules (more details in next major section and in Table 1), a single daily high (or low) water elevation was traditionally selected at stations where the tide is classified as being *diurnal* (displaying a single cycle with one high and low per day, Fig. 1, Tidal Day 4); elevations of two high (or low) waters are selected daily at stations classified as *semidiurnal mixed* (displaying two cycles per day, Fig. 1, Tidal Days 1, 2, and 3). Sum the selected observations and divide by the number selected. By thus averaging long series of data, the shorter cyclic variations (astronomical tide) and the random (weather-related) variations are damped. Including the full 19th year avoids biasing related to the strong annual cycle. Significant longer term (19-year) variations arise from astronomical forces, thermal expansion (contraction) of the oceans, crustal movement, glacial melting, etc. These longer-term changes (see “► [Eustasy](#)” and “► [Changing Sea Levels](#)”) are compensated for, as needed at irregular intervals, by adopting evermore recent 19-year periods (Epochs) as the National Tidal Datum Epoch. For North America, the current Epoch is 1960–78; previous Epochs include 1941–59 and 1924–42. Agencies responsible for datum definitions elsewhere agree that tidal epochs should span complete 19-year periods, but they do not all use the same 19-year periods.

Spaced-Based Measurements

After a period of improvement and acceptance, spaced-based measurements derived from the NAVSTAR Global Positioning System (GPS) are now widely used not only as a convenient way to reestablish local datum elevations near benchmarks, but also in defining new coordinate systems (lowest portion of Table 1). Numerical models can introduce tidal datums into these three-dimensional coordinate systems based on idealized reference ellipsoids rather than tidal data. This approach is now quicker and less expensive than collecting tidal measurements or leveling from local tidal datum benchmarks. In the future, direct measurement of sea surface elevations from space may become accurate enough to establish tidal datum elevations essentially continuously along the coast in an iterative blend of observation and tidal dynamic modeling.

Limitations and Precautions in Applications

Temporal and Spatial Variations

The difference between successive high and low waters, *tide range*, is neither uniform nor constant. The phases of the moon and the inclination of its orbital plane with respect to the plane of the earth's orbit around the sun (*declination*) are two of the factors contributing to daily variations in tide range on approximately 28-day and 14-day cycles, respectively. In contrast to the elegant theory explaining temporal variations in the astronomical tide, its spatial variations are complex. Tidal dynamics vary with location on the earth, outline of the coast, bathymetry of the seafloor, and such factors as wind, salinity, and river discharge. Each tidal component is characterized by an amplitude and frequency (see “► [Tides](#)”). The bathymetry and outline of the coast (*basin shape*) modulates sea-level response to the total gravitational force in a manner somewhat analogous to the way the material properties and shape of a poorly tuned musical instrument distorts chords. Some frequencies resonate and are amplified. Others are out of tune with the basin (instrument) and are damped (muffled). Furthermore, frequency-dependent nodes (antinodes) may exist within the basin where responses are maximized (minimized).

Nonuniform Height Differences Between Datums

Complexities of tidal propagation cause the mean differences among tidal datums to vary spatially, for example, the difference between MLW and MLLW varies along shore. The mean tide range varies around the world being about 10 m at Anchorage, Alaska; 1 m at San Diego, California; 0.1 m at Galveston, Texas; and less than a centimeter at Chicago, Illinois, on Lake Michigan.

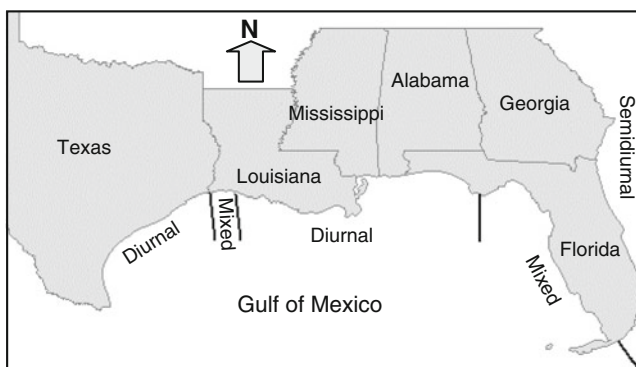
Advent of the www improved the availability and dissemination of the empirically determined elevation differences

among tidal datums in the form of tables issued by federal organizations throughout the world (e.g., <http://co-ops.nos.noaa.gov/bench.html>). Many of these tables also report the highest and lowest observed sea levels. Some sea-level datums (e.g., the British chart datum, Lowest Astronomical Tide) are based on extremes.

Type of Tide

The range and timing of the tide goes through cycles as lunar and solar declinations change. They also vary spatially over the earth's surface. The type of tide varies spatially and temporally as well. The type of tide is predominately *semidiurnal* (two similar, semi-sinusoidal cycles per day with little difference between the two lows or the two highs) over most of the US East Coast and *mixed* (two lows and two highs per day that are unequal in elevation) along most of the US West Coast. Along the Gulf of Mexico, the type of tide is not so uniform, but at most locations it is *diurnal* (only one high and one low per day). Figure 1 shows a mixed type during Tidal Day 1 and Tidal Day 3. Tidal Day 2 is semidiurnal and Tidal Day 4 is diurnal. A real tide curve would transition between these types more gradually (see “► Tides”).

The most common type of tide throughout the world is semidiurnal. Figure 2 depicts regional variations in type of tide along the north shore of the Gulf of Mexico and a portion of the Atlantic. The basin shapes in Tampa Bay and Charlotte Harbor, on the west coast of Florida, are tuned to diurnal tidal forces. Therefore, the tide type in these estuaries is predominately diurnal even though the open Gulf beaches of west Florida are predominately mixed. *Predominately* that the type of tide is as designated for more than half of the time at that particular site. Along the Mediterranean Sea, stretches of coast that are predominately semidiurnal alternate with reaches that are predominately mixed. Tides are predominately semidiurnal around most of England, mixed around most of Australia, and diurnal along much of southeast Asia, Japan, and northeast Siberia.



Tidal Datums, Fig. 2 Regional variations in type of tide. (Modified from Hicks et al. 1988)

Analysis procedures and datum elevations were established in the Gulf of Mexico prior to a full appreciation of these complex changes in type of tide (Hicks et al. 1988). With the benefit of longer term and spatially denser measurements, it became apparent that some initial datum calculation procedures (as well as inconsistencies in terminology) were causing problems. One of the problems was that the MHW and MLW datums had abrupt breaks (*jump discontinuities*) in elevation as one moved alongshore. The global continuity of sea level was a prime feature promoting the widespread adoption of tidal datums. MHW and MLW discontinuities arose because, traditionally, their official definitions included provisions that where the tide was classified as predominately diurnal, only one high (or low) would be included daily in the 19-year series for datum determination. Where the tide was predominately semidiurnal or mixed, both lows and both highs were traditional (and still are) included in these calculations. There are times during the monthly lunar cycle when diurnal forces are small relative to semidiurnal forces. During these times, the tide signal (even at sites where the tide is classified as diurnal) may take on mixed-type characteristics briefly (i.e., exhibit a second daily high and/or low). Likewise at sites that are mixed-type, the second daily high and/or low may become indiscernible for a few days during the month. Tide type is unambiguously classified according to a ratio of amplitude terms for the major tidal constituents. The shape of tide curves vary smoothly along the shore with transition zones between areas with different types of tide. Discontinuities arose from the traditional definitions of MHW and MLW being dependent on tide type. Resulting problems were especially acute in the case of MHW because of its role as a legal state boundary.

Through a series of carefully planned and well-coordinated moves that began as early as the 1970s, the procedures for defining MHW and MLW and the inconsistent naming practices that had crept in at some Gulf sites were changed. Table 1 follows the revised, uniform NOS definitions. Another half dozen modifications were collected into one endeavor that unified analysis, created a single uniform chart datum (MLLW), freed MHW and MLW definitions from dependence on tide type, updated the National Tidal Datum Epoch, moved private-state property boundary seaward along certain reaches, and eliminated all datum discontinuities (Hicks et al. 1988). Though announced much earlier, this comprehensive corrective action officially took effect on November 28, 1980. Complete updating of charts and NOS's Coast Pilot actually proceeded incrementally over the following decade. Tide tables predicting tidal heights for East Coast of North and South America switched chart datum from MLW to MLLW starting with the 1989 volume (making it consistent with NOS tables for other regions). This correction (The National Tidal Datum Convention of 1980) thus simplified standards for tidal datums throughout the Americas.

Nontidal Datums

The *geopotential elevation* of tidal datums vary spatially (i.e., the potential energy of a unit mass at a given elevation with respect to each datum varies geographically). For long periods, a tidal datum can deviate on the order of a meter from an *equipotential surface* (surface of constant specific potential energy) due to prevailing atmospheric pressure, temperature, wind, currents, and salinity differences (see “► [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)”). For example, the geopotential elevation of sea level rises (falls) generally northward along the west (east) coast of the United States. For certain applications (e.g., long-distance pumping, ballistics) it is desirable that the elevation of an equipotential surface be spatially invariant. In such cases, a nontidal datum is clearly required (lower portion of Table 1).

The latest geodetic datum (NAVD 88, middle of Table 1) is not parallel to any tidal datum or to the geodetic datum it replaced (NGVD 29). The NGVD 29 was defined so that it approximated the MSL datum. The new NAVD 88 is on the order of a meter above the MSL datum on the West Coast and 0.1–0.5 m below the MSL datum on the East Coast (Fig. 3) and is independent of sea level.

Shorelines

Conceptually, the intersection of any tidal datum with the beach rigorously defines a shoreline. The acronym for this shoreline is obtained by appending an “L” to the acronym for the selected tidal datum. In the United States, the mean high water line (MHWL) was adopted by most states as the boundary between private and public land. Six northeastern states allow private ownership down to the mean lower low water line (MLLWL). In Texas, private ownership has, at least in certain cases, been restricted to being above the mean higher high water line (MHHWL). Federal submerged lands, exclusive fishing zones, and national economic zones are related back to the MLLWL by their legal definitions.

In practice, charted shorelines are sometimes inferred from rectified aerial photographs taken at specific phases of the tide. Such mapped shorelines only approximate the rigorously defined tidal-datum shore-lines, but are much easier to establish. Unfortunately, imprecise usage often results in them being labeled MHWL or MLWL. Even less-accurate approximations of tidal-datum shorelines are inferred from the wetted bound on the beach, berm line, and toe of the dune as identified in aerial photographs. These charted shorelines may be unlabeled or mislabeled.

Statements like “this chart uses the MLLW datum,” do not imply that the shoreline is the MLLWL or any other tidal datum, only that printed soundings are referenced to MLLW. The shoreline could be the water’s edge at an arbitrary height

of the tide, a high water mark, or follow a geomorphic feature like a berm or toe of dune. Caution is advised not to infer too much from shoreline labels without consulting the description of the procedures used in production of that specific chart. Increased use of digital maps may exacerbate problems because chart developers can no longer anticipate the scale to which the user will take measurements from these zoomable products.

The label “HWL” (for high water line as distinct from MHWL) specifically acknowledges that the shoreline has been located less precisely than tidal-datum shorelines. Nevertheless, the HWL does approximate the MHWL, and is therefore considered “of primary importance in distinguishing the upland from the shore on charts” (Shalowitz 1964). For practical reasons of determination, however, the surveyed HWL is based on observed physical features in the field (such as berm crest). Thus, the HWL stands above the MHWL by an uncertain distance related to wave runup and surveyor judgment.

Concluding Guidance

Tidal datums are listed in Table 1 in the order of decreasing elevation from top to bottom. The vertical offset of orthometric and three-dimensional datums (lower portions of Table 1) from tidal datums varies spatially (e.g., Fig. 3). Chart depths usually refer to a tidal datum (MLLW is the official chart datum to which depths are referred in the United States). Elevations on maps usually refer to an orthometric or a three-dimensional datum (NAVD 88 or NAD 83 for older maps in the United States). By convention, elevations are positive above datum; depths are positive below datum (i.e., an object 10 m below the water has an elevation of 10 m and a depth of 10 m).

In the 1990s, proliferation of personal computer software made it easy to convert between reference systems (see “► [Geographic Information Systems \(GIS\)](#)”). Caution is



Tidal Datums, Fig. 3 Height in centimeters of MSL datum above NAVD88. (Redrawn from USACE 2002)

advised, however, when analyzing measurements made to different references. For example, the effectiveness of coastal erosion mitigation is often assessed by analyzing the difference between variable rates of shoreline change after project completion and a series of measurements made before project initiation. Consider a scrabbled mixture of HWLs and MHWLs. Early shorelines exist only as HWLs. The ratio of HWLs to MHWLs would tend to decrease as aerial surveying and GPS technology became pervasive. The horizontal distance separating a HWL from a MHWL is not recoverable after the HWL survey. Because shoreline positions based on tidal and orthometric datums are more precisely defined than the HWL, MHWL positions may appear more reliable (for analyzing shoreline change rates) than the HWLs that only approximate them. For just the years that have both a HWL and a MHWL, it might seem desirable to delete the less precise of the two measurements, but this would tend to bias the post-construction period and produce the appearance of unwarranted erosion reduction (i.e., deletion the HWLs would tend to displace the shoreline landward more frequently during later years). In most such cases, the more precisely located MHWLs might well be ignored to obtain an unbiased (though imprecise) estimate of changes in the rate of shore retreat.

Adoption of a new National Tidal Datum Epoch is pending. Previous updates have raised datums (reduced the elevations of benchmarks above tidal datums) by amounts that varied with the particular epoch transition, gauge location, and datum selection. For the MHW datum, an increase on the order of 0.25 m would not be unusual from one Epoch to the next. In areas of subsidence (Louisiana, the Netherlands, and portions of Texas, see “► [Coastal Subsidence](#)”) or crustal uplift (e.g., Alaska, see “► [Changing Sea Levels](#)”), datums could be raised or lowered on the order of a meter. Comparisons of shoreline positions documented during different Epochs can, therefore, be problematic. Adjustments can be made to estimate the shoreline positions for a common Epoch based on the vertical shift in tidal datum between two Epochs and the slope of the beach at these elevations. The adjusted results would only be as good as the knowledge of the beach slope used to translate the vertical shift into a shoreline displacement, but the results could be useful. Beaches (including the dune, berm, and long-shore bars) generally migrate upward with the historic long-term rise of sea level. So, comparing unadjusted shorelines, referred to datums that shifted between epochs, can also be valid from the perspective of quantifying how the volume of beach sand changes in concert with sea-level changes. Whether to accept the inter-Epoch shoreline displacement or adjust the shorelines would depend on the time span, magnitude of sea-level change, and knowledge of slope changes. Analysis from both perspectives can be useful. For example, along a 50 km reach of Lake Michigan, the average shore retreat of

17.9 m (over a specific 6 years during which the mean lake level rose persistently) was decomposed into 14.5 m (at a fixed datum elevation) due to erosion and recession of the profile and 3.4 m due to submergence under a 0.39 m higher mean lake level (Fig. 11, Hands 1979).

For accuracy, charts and tables should clearly state the method of determining elevations and shorelines, coordinate systems, units of measurement, and any coordinate transform procedures employed. To maintain accuracies on the order of 0.1 m or better, be certain of and document which particular Epoch was used in the tidal datum definition.

Cross-References

- [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
- [Changing Sea Levels](#)
- [Coastal Changes, Gradual](#)
- [Coastal Subsidence](#)
- [Eustasy](#)
- [Geodesy](#)
- [Geographic Information Systems \(GIS\)](#)
- [Sea-Level Fluctuations over the Last Millennium](#)
- [Tide Gauges](#)
- [Tides](#)

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Tidal Environments

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Definitions

A tidal environment is that part of a marine shore which is regularly submerged and exposed in the course of the rise and fall of the tide. Such environments exhibit particular physical and biological characteristics which, among others, play an important role in coastal dynamics, coastal ecology, coastal protection and engineering works, and integrated coastal zone management.

The coastal area affected by the ocean tides is known as the intertidal or eulittoral zone. Being a long-period wave, the tidal water level oscillates about a mean water level, which usually corresponds to the mean sea level. The vertical distance covered by the tide is known as the tidal range, whereas the part above or below the mean tide level is the tidal amplitude, which can hence have a positive or negative sign. In practice, a number of critical tide levels are distinguished on the basis of longer-term averages. Proceeding from high to low water levels, these are: MEHWS, mean equinoxial high water springs (highest astronomical tide); MHWS, mean high water springs; MHWN, mean high water neaps; MWL, mean water level (commonly mean sea level); MLWN, mean low water neaps; MLWS, mean low water springs; and MELWS, mean equinoxial low water springs (lowest astronomical tide). These mean tide levels are well correlated with various morphological and biological characteristics of tidal environments.

Besides the astronomical modulations, instantaneous tidal elevations can, in the short term, be substantially modified by water-level fluctuations induced by wind and/or wave set-up or set-down. The degree of wave and wind exposure can thus have a substantial influence on the nature of a tidal shore. Where such secondary and irregular fluctuations in coastal water levels are so frequent and strong that they completely mask the tidal signal, the coast is considered to be nontidal (e.g., the Baltic Sea).

In summary, tidal shores are highly variable environments which are not only influenced by the astronomically induced periodic rise and fall of the sea level, but also by numerous secondary processes. In combination, these factors define the local physical nature of a tidal environment (e.g., Davies 1980; Allen and Pye 1992; Allen 1997; French 1997).

Tidal Forcing Factors

The tides are essentially produced by the interactive forces exerted on the oceans by the sun and the moon. Since the

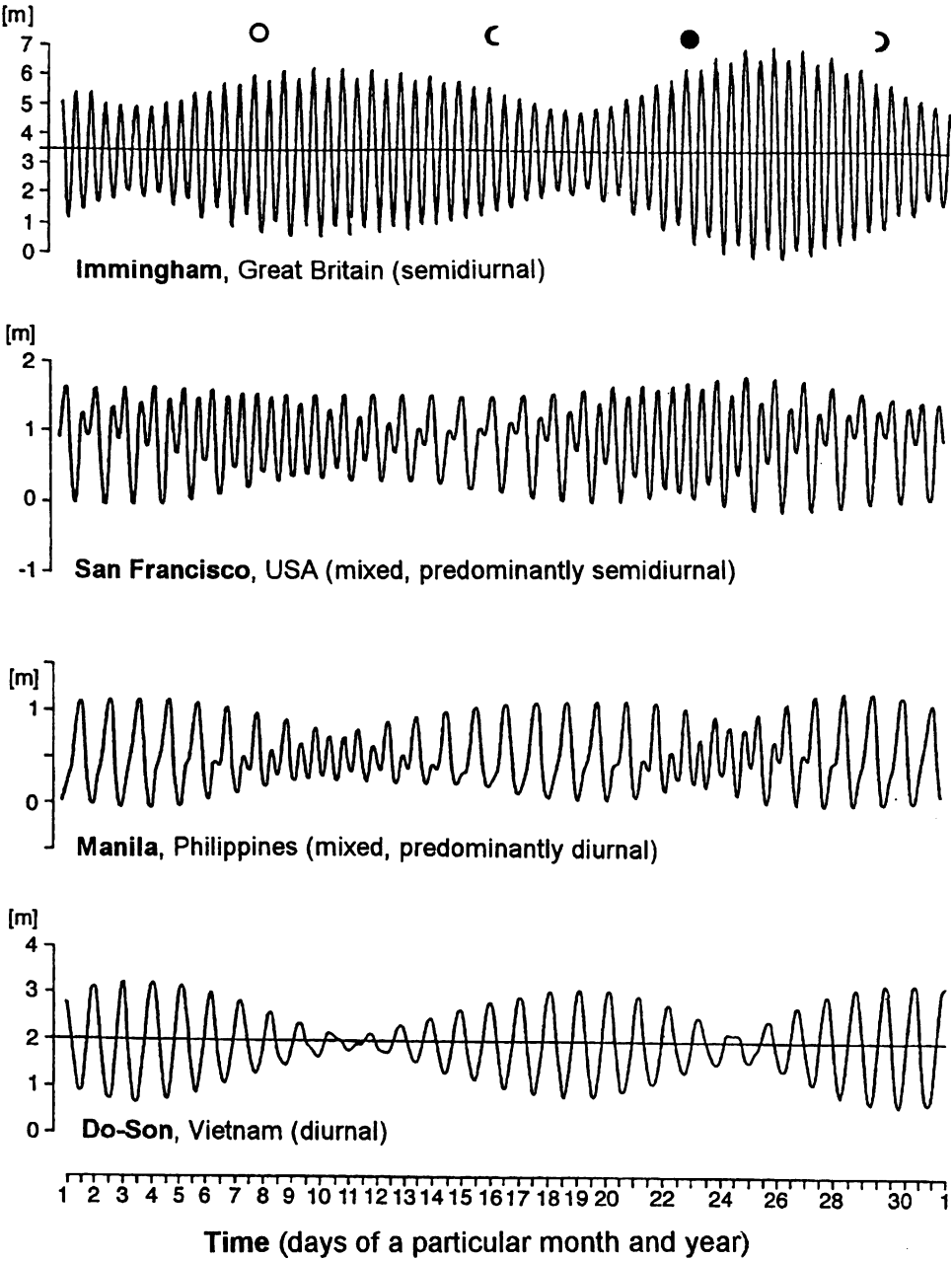
motions of the sun and the earth-moon system are known with great precision, the tide-generating potential can be mathematically resolved into strictly periodic components. These components vary over time as the position of the sun and the moon relative to the position and orientation of the earth changes. The sum of all the tractive forces at any one time defines the total instantaneous potential. Doodson (1922) computed no less than 390 such harmonic components, of which about 100 are long period, 160 diurnal, 115 semidiurnal, and 14 one-third diurnal. Of these, only seven (i.e., four semidiurnal and three diurnal components) are of practical importance (e.g., McLellan 1975). Indeed, only four of these are used to define the character of the tides around the globe (Fig. 1), in which semidiurnal, diurnal, and two mixed tidal types, comprising a predominantly semidiurnal and a predominantly diurnal one, are distinguished (cf. Table 1). Since the tidal type has an important influence on the physical nature of tidal environments, the global distribution of these four types are illustrated in Fig. 2.

As the sun, the earth, and the moon move along their elliptical orbits, they continually change their positions relative to each other. As a result, the total potential defining the height of the astronomical tide is modulated as a function of geographic location in the course of a day, a month, a year, and also on longer timescales. The most prominent tidal period is the fortnightly spring-neap cycle (synodic tide). Thus, spring tides coincide with full moon and new moon, whereas neap tides occur at the corresponding half-moon phases. These tidal ranges are further modulated in the course of a lunar month (anomalistic tides), the highest spring tides occurring when the moon is closest to the earth (perigee), the lowest when the moon is furthest away from the earth (apogee), the difference in distance amounting to roughly 13%. The solar component varies in similar manner between perihelion (currently coinciding with the winter solstice) and aphelion (currently coinciding with the summer solstice), the difference in distance between the two amounting to about 4%.

Another important astronomical feature influencing tidal environments on short timescales is the daily inequality of the tide (declinational tide). This feature results from the inclination of the earth's axis relative to the plane of the ecliptic, and hence the tidal bulge. As the earth rotates around its axis, the position of a geographic locality continually changes relative to the tidal bulge, alternating between a maximum and a minimum tidal elevation every 12.42 h, a feature which affects the elevations of both successive high tides and low tides. On this, the declination of the moon is superimposed which, in turn, causes a progressive change in the daily inequality over time. Thus, with the moon over the equator, successive tides are equal in height (equatorial tides), whereas towards the position of maximum declination successive tides become progressively more unequal in height, dividing into a "large tide" and a "half tide" (tropical tides). The solar tide

Tidal Environments, Fig. 1

Selected examples of tidal curves from different geographic locations illustrating the main tidal types. (Adapted from Dietrich et al. 1975). (a) semi-diurnal tides (e.g., Immingham, Great Britain); (b) mixed, predominantly semidiurnal (e.g., San Francisco, USA); (c) mixed, predominantly diurnal (e.g., Manila, Philippines); (d) diurnal (e.g., Do-Son, Vietnam)



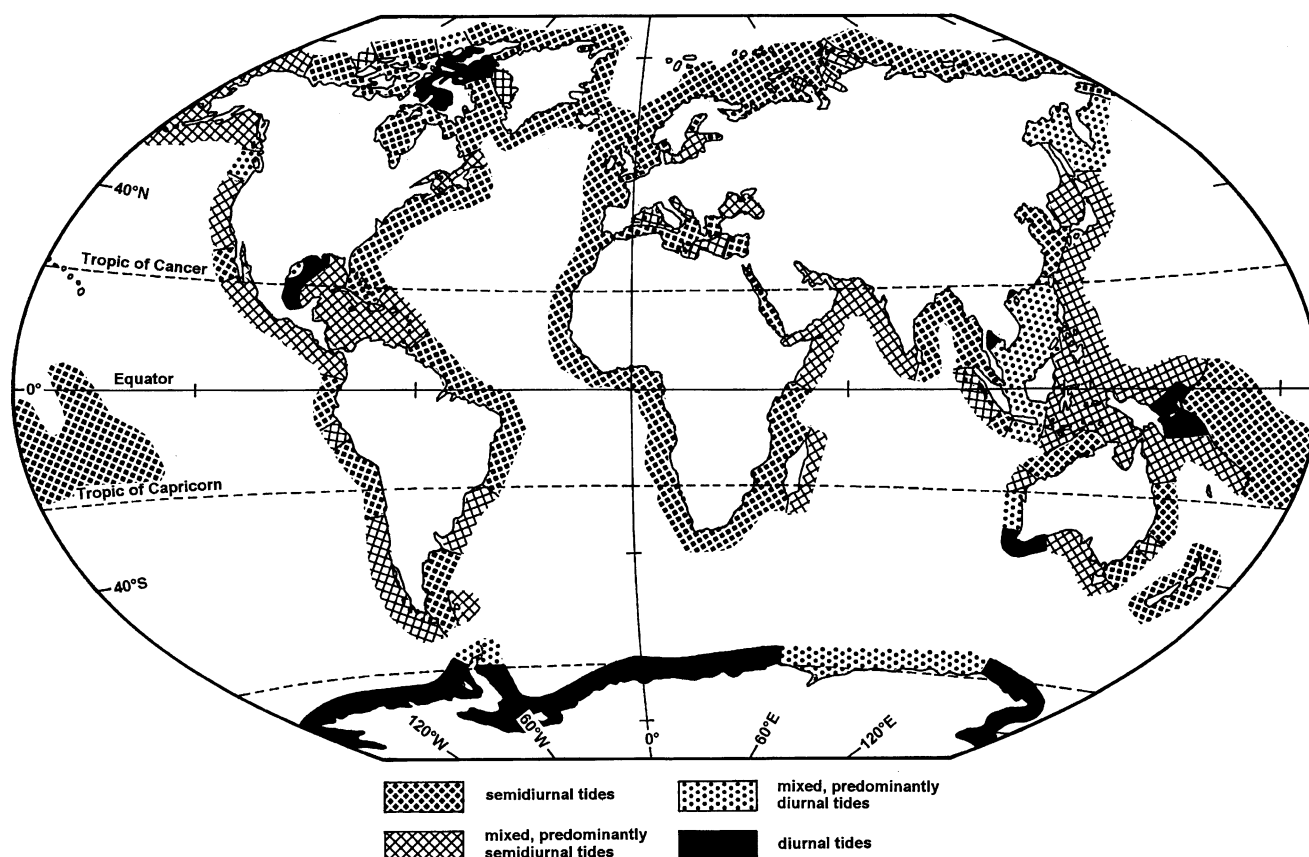
shows a similar inequality, being zero at the equinoxes (spring and autumn) and largest at the solstices (summer and winter). The largest astronomical tides, that is, the greatest spring and lowest neap tides, in the course of a year occur at the vernal and autumnal equinoxes when the orbital path of the earth–moon system crosses the plane of the celestial equator (nodal points), that is, when the total potential of the tide-generating forces is largest because of their closest alignment. In addition to these short-period astronomical cycles, the tide is also modulated by longer period phenomena which need to be considered when analyzing ancient tidal deposits (e.g., Archer et al. 1991; Williams 1991; Oost et al. 1993; de Boer and Smith 1994). Among these are the nutational motion or nodal cycle (≈ 18.6 -year cycle) caused by the

Tidal Environments, Table 1 The tidal character (F) as defined on the basis of the ratio between the sum of the lunisolar diurnal (K_1) and principle lunar diurnal (O_1) and the sum of the principle lunar semidiurnal (M_2) and principle solar semidiurnal (S_2) tidal components

$F = K_1 + O_1 / M_2 + S_2$	Tidal character
0–0.25	Semidiurnal tides
0.25–1.5	Mixed, predominantly semidiurnal tides
1.5–3.0	Mixed, predominantly diurnal tides
>3.0	Diurnal tides

Source: Dietrich et al. (1975)

oscillation of the earth’s axis about its mean position, the precessional motion ($\approx 23,000$ -year cycle) of the earth’s axis, the obliquity ($\approx 41,000$ -year cycle) which defines the angle of inclination of the earth’s axis between 22° and 24.8°



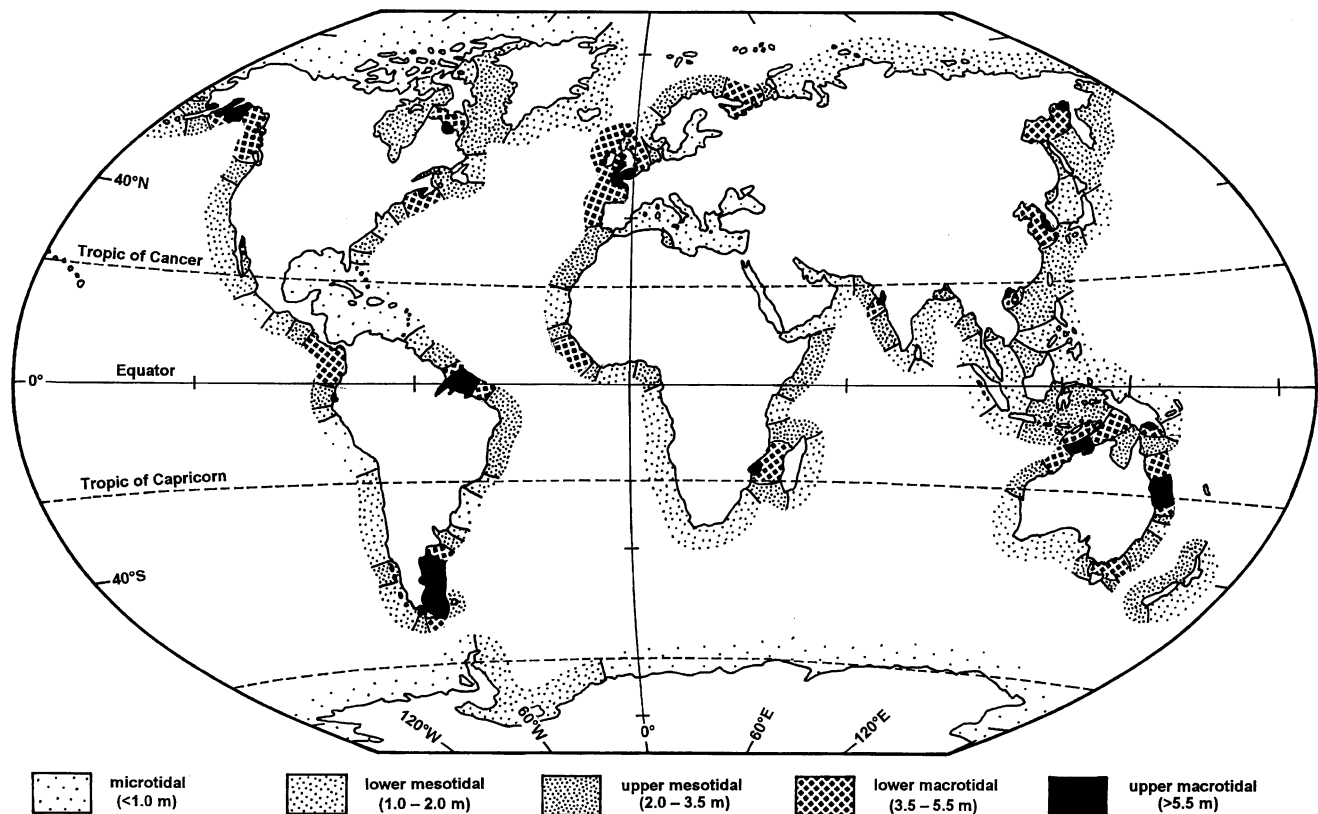
Tidal Environments, Fig. 2 Global distribution of the main tidal types. (Adapted from Davies 1980). Note that the transitions between semidiurnal and diurnal tidal types (here represented by so-called mixed tides) are progressive and not abrupt

(currently 23.5°), and the eccentricity ($\approx 100,000$ -year cycle) controlling the rate of change in the elliptical radius of the earth's orbit around the sun. In addition, it has been shown that the length of the year has decreased from 420 to currently 365 days, while the length of the day has increased from 21 to currently 24 h in the course of the last 500 million years or so (e.g., Williams 1991).

Besides the geographic variation in tidal type and tidal range resulting from astronomical modulations, the physical nature of a tidal environment is also influenced by numerous secondary factors. Among the more important of these are variations in tidal range with distance from the amphidromic point around which the tidal wave rotates, the Coriolis effect as a function of geographic location, the rate of change in water depth, the coastline configuration (plan shape and slope angle), as well as resonant effects resulting from the shape and depth of a tidal basin. Thus, at the center of an amphidrome the tidal range is considered to be zero, but it progressively increases in height with distance along the axis of the two opposing tidal bulges which rotate around the center offset by 180° or 6 h (tidal phases). The Coriolis force, in turn, forces the tidal wave to rotate clockwise in the Southern Hemisphere and anticlockwise in the Northern Hemisphere, but this principle can be upset near the Equator. The direction in which the

tidal wave propagates along a shore thus depends entirely on the geographic location of the coast. Where tidal waves rotating around neighboring amphidromes meet, the tidal water motion can even be perpendicular to the coast.

In addition, the amplification of the tidal wave and the tidal type are strongly influenced by the configuration of the coastline and the rate of shoaling, the effect of this interaction being illustrated in Figs 2 and 3. With decreasing water depth (h) the length of the tidal wave (λ) progressively decreases proportionally in the form $\lambda \propto h^{0.5}$. When propagating into funnel-shaped estuarine water bodies, the initial increase in tidal range due to convergence of the opposite shores is eventually compensated for by friction along the seabed as a result of which the tidal wave gradually decreases in height. This interplay between friction and convergence is proportionally related in the form $a_0 \propto b^{-0.5} * h^{-0.25}$ where a_0 is the amplitude of the tidal wave (m), b is the width (m), and h is the water depth (m) at a particular location along the estuary. This relationship can take on complicated patterns and in nature three basic modes of tidal wave propagation can be distinguished. Thus, in the case where friction dominates over convergence the tidal range progressively decreases in height up-estuary (hypersynchronous mode). In the opposite case, the tidal range increases in height (hyposynchronous mode), and



Tidal Environments, Fig. 3 Global distribution of tidal shores based on tidal range according to the classification scheme of Hayes (1979)

where friction and convergence are in balance the tidal range remains constant (synchronic mode). Even neighboring estuaries will exhibit quite different modes of tidal wave propagation if they differ in shape and water depth (e.g., Borrego et al. 1995). In many estuarine environments a combination of two or even all three modes can be observed.

In some cases, the entrance channel of an estuary or lagoon is so narrow and shallow that the propagation of the tidal wave is “choked” with the effect that the tidal range is dramatically reduced. This filtering mechanism, expressed by the so-called coefficient of repletion (K), has the numerical form of $K = (T/2a_0\pi) * (A_c/A_b) * [2a_0g/(1 + 2gln^2r^{-4/3})]^{0.5}$ where T is the tidal period, $2a_0$ is the tidal range (a_0 being the tidal amplitude), l is the length of the entrance channel, A_c is the cross-sectional channel area, A_b is the surface area of the water body, r is the hydraulic radius of the channel, g is the gravitational acceleration, and n is the Manning’s friction (0.01 – $0.10 \text{ s m}^{-1/3}$). This coefficient of repletion controls reductions in tidal range, phase shifts between ocean and lagoonal tides, non-sinusoidal variations of the lagoonal tides, and flow exchanges between the ocean and the estuarine or lagoonal water bodies (e.g., Kjerfve and Magill 1989).

A final forcing factor which may affect the behavior of tidal waves in shallow water is resonance, a feature associated

with standing waves. Clearly, a standing wave superimposed on a normal tidal wave would dramatically affect the physical nature of a tidal environment. In this context, we distinguish two types. In the case of a half-wave oscillator or seiche, the length of the water body is half the wavelength of the standing wave. The fundamental period (T) of a seiche is defined as $T = 2l/(gh)^{0.5}$ where l is the length of the water body, h is the water depth, and g is the gravitational acceleration. Seiches are particularly common in lakes and marginal seas, where they are forced by wind stress, and hence of less importance in tidal environments. However, quarter-wave oscillators come into operation where the lengths of open-ended elongate gulfs or deep estuaries together with adjacent bays correspond to a quarter of the tidal wavelength ($l = 0.25T * (gh)^{0.5}$). In this case, the period $T = 4l/(gh)^{0.5}$ with notations as above. If this condition is fulfilled, or nearly so, tidal amplification can be quite considerable (e.g., Bay of Fundy, Canada).

Coastal Classification by Tidal Type and Range

Besides classifying the world’s coastline according to tidal type (Fig. 2), coastal tidal environments are also classified on the basis of tidal range, the scheme of Davies (1964, 1980)

Tidal Environments, Table 2 Contrasting two existing classification schemes of tidal shores on the basis of tidal range

Davies (1964)		Hayes (1979)	
Tidal range (m)	Class name	Tidal range (m)	Class name
<2.0	Microtidal	<1.0	Microtidal
2.0–4.0	Mesotidal	1.0–2.0	Lower mesotidal
>4.0	Macrotidal	2.0–3.5	Upper mesotidal
		3.5–5.5	Lower macrotidal
		>5.5	Upper macrotidal

having been the most widely applied to date (Table 2). However, with only three subdivisions, this rather arbitrary approach prevents differentiation where it is most needed, that is, near the lower and upper limits of the potential tidal regime. For example, the Gulf of Mexico coast with an average tidal range of only 0.5 m is very different in character from the west coast of southern Africa where the tidal range averages at 1.6 m, yet both are classified as microtidal. In contrast to this, a more pragmatic classification, comprising five subdivisions (see Table 1), has been proposed by Hayes (1979). This latter scheme takes distinct, process-related geomorphic features into consideration, for example, the upper limit of barrier island occurrence at a tidal range of 3.5 m, which hence marks the transition between upper mesotidal and lower macrotidal regimes in this classification. To contrast the two classification schemes, the global pattern of coastal subdivision using the latter scheme is presented here for the first time (Fig. 3).

Rocky Versus Sandy and Muddy Tidal Environments

The coastlines of the world can be divided into four basic types, namely rocky shores, sandy shores, muddy shores, and bio-shores. In all cases, the character of a particular shore reflects the interaction between the substrate, the local wave climate, the tides, and the biology (e.g., Newell 1979; Davies 1980; Raffaelli and Hawkins 1996). Geographic location, which controls climatic influences and biological species composition (e.g., Chapman 1974), and the Holocene evolution of a coast (e.g., Bird and Schwartz 1985) being important additional factors to consider. Rocky shores occupy the smallest overall area because the intertidal zone is commonly narrow as compared with the other shore types. Shore processes have received considerable attention by engineers for constructional purposes (e.g., Horikawa 1989), whereas biologists have long been intrigued by the distinct faunal and floral zonation patterns along tidal gradients which evidently reflect high degrees of adaptation to the

intertidal environment and the overprinting effects of competition between specific organisms (e.g., Lewis 1972; Raffaelli and Hawkins 1996). However, the interplay between biological and physical factors in defining zonation patterns is still not well enough understood to allow accurate predictions to be made (e.g., Delafontaine and Flemming 1989). Different tidal environments are also characterized by different biogeochemical processes due to different climates, substrates, and biological community structures (Alongi 1998).

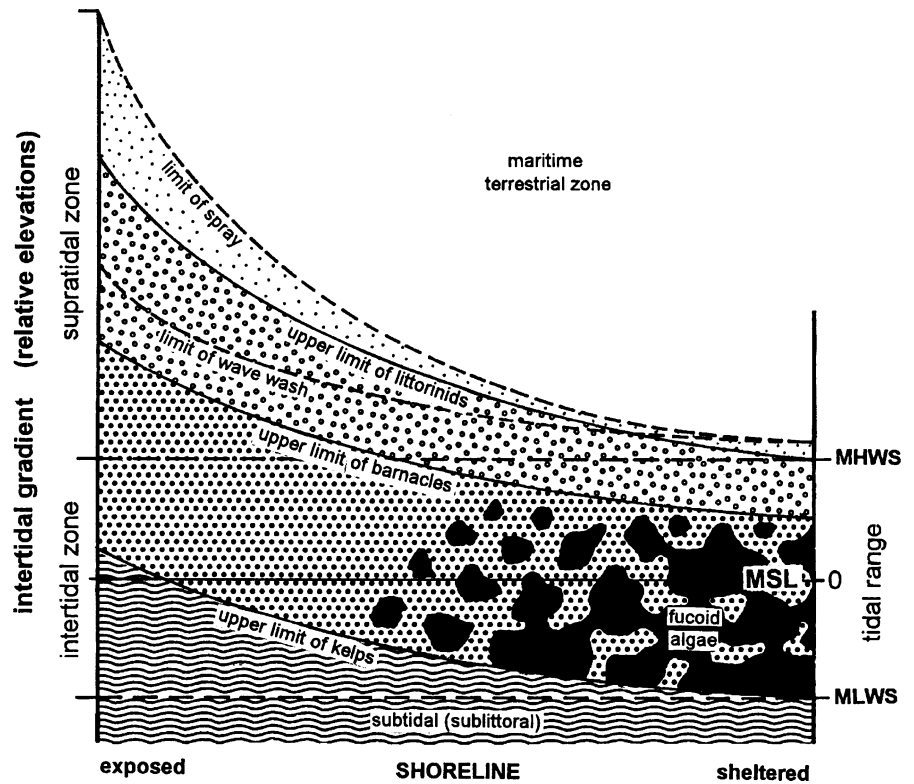
The biological zonation pattern observed along the rocky shores of Great Britain as a function of tidal gradient and the degree of exposure to wave action is illustrated in Fig. 4 (adapted from Lewis 1972; Raffaelli and Hawkins 1996). This basic scheme can be applied to most rocky shores of the world, the only difference being the species composition and distribution, a good example from subtropical Bermuda being shown in Thomas (1985). Important to note here is that the tidal gradient in Fig. 4 is relative, expanding or contracting proportional to the tidal range.

In contrast to rocky shores, into which bio-shores can also be included, sandy and muddy tidal shores can attain shore normal extensions of many kilometers (e.g., Davis 1994; Flemming and Hertweck 1994). French (1997) distinguishes no less than seven intertidal coastal types based mainly on morphology and facies successions. The final transition between land and sea along sheltered coasts is characterized by sharp boundaries separating different floral zones, the typical pattern observed along the barrier island shore of the southern North Sea being illustrated in Fig. 5 (adapted from Streif 1990).

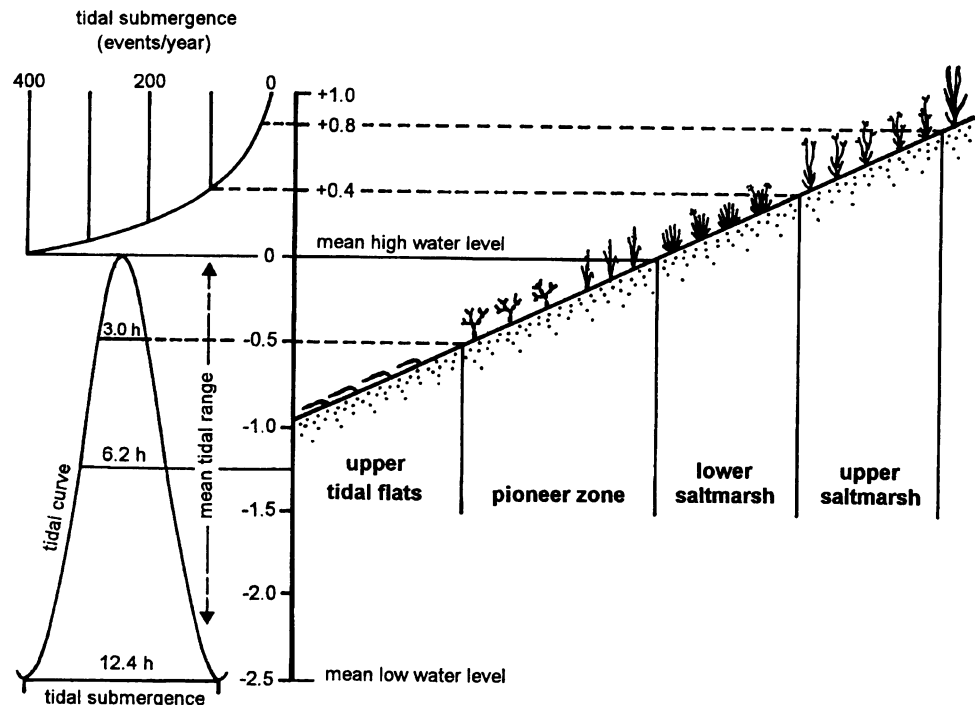
In this example the zones are separated by the frequency of tidal submergence in the course of the year which is controlled by the elevation along the tidal gradient. In principle, this basic pattern should be applicable the world over, individual transition levels being dependent on local floral associations, the tidal range, the seasonal wave climate, and the difference in elevation between mean high tide levels at spring and neap tide (e.g., Chapman 1974; Lugo and Snedaker 1974). A systematic investigation of the factors controlling the upper and lower limits of occurrence of *Spartina anglica* relative to mean sea level along the coast of the United Kingdom has generated quantitative relationships between the spring tidal range, the fetch available for wave generation and propagation, the area of the tidal basin, and in the case of the upper limit also the geographic location (Gray 1992). Thus, the lower limit (L_1) is defined as $L_1 = -0.805 + 0.366R_s + 0.053F + 0.135 * \log_e A_b$, where R_s is the spring tidal range (m), F is the fetch in the direction of the transect (km), and A_b is the area of the tidal basin (km^2). The upper limit, by contrast, is defined as $L_u = 4.74 + 0.4837R_s + 0.068F - 0.199L^\circ$, where L° is the degree North Latitude expressed as a decimal. The correlation

Tidal Environments, Fig. 4

Zonation patterns along rocky tidal shores as a function of exposure to wave action. (Adapted from Lewis 1972; Raffaelli and Hawkins 1996). Note that the scaling is relative to any particular observed tidal gradient

**Tidal Environments, Fig. 5**

Floral zonation pattern observed at the transition between upper intertidal flats and the terrestrial environment of the Wadden Sea (southern North Sea) as a function of tidal elevation and the frequency of tidal submergence in the course of a year. (Adapted from Streif 1990)



coefficients of $r = 0.97$ for L_1 and $r = 0.95$ for L_u demonstrate the predictive potential of this approach.

A comprehensive overview of physical and biological processes active along sandy tidal shores is provided in McLachlan and Erasmus (1983).

Outlook

Many features of tidal environments are still poorly understood. Among these are the quasi-periodic, decadal to sub-decadal fluctuations in the elevation of mean high tide and

mean low tide levels. Being a worldwide phenomenon, one might assume that they result from variations in the astronomical factors defining the tidal potential. A clear correlation, however, is still lacking. As far as sandy tidal environments are concerned, accurate sediment budgets and transport pathways have remained elusive problems whose solution becomes more pressing in view of the predicted acceleration in sea-level rise. The distinction between strictly local features and others of global relevance requires more attention. A number of other unresolved issues have been addressed in the text.

Cross-References

- [Barrier Island Landforms](#)
- [Beach Processes](#)
- [Bioerosion](#)
- [Estuaries](#)
- [Holocene Coastal Geomorphology](#)
- [Littoral](#)
- [Microtidal Coasts](#)
- [Rock Coast Processes](#)
- [Sandy Coasts](#)
- [Tidal Flats](#)
- [Tides](#)
- [Wave-Dominated Coasts](#)

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Tidal Flats

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Definition and Introduction

Tidal flats are low-gradient tidally inundated coastal surfaces. Jackson (1997) defines them as extensive, nearly horizontal,

marshy, or barren tracts of land alternately covered and uncovered by the tide, and consisting of unconsolidated sediment. Tidal flats may be muddy, sandy, gravelly, covered in shell pavements, or locally underlain by rock pavement and, compositionally, be underlain by siliciclastic or carbonate sediments. They are complex coastal systems combining elements of coastal geomorphology, sedimentology, hydrology, hydrochemistry, diagenesis, biology, and ecology.

Geologically, tidal flats have been of great interest to sedimentologists and stratigraphers as coastal systems that are readily accessible to sampling and study, and rich in processes and products resulting from oceanographic, sedimentologic, geohydrologic, hydrologic, hydrochemical, mineralogic, and biotic interactions (Ginsburg 1975; Klein 1976; Alexander et al. 1998; Black et al. 1998). They contrast with steeper-gradient wave-dominated sedimentary coasts such as sandy beaches composed dominantly of sand and with a relatively limited biota, because tidal flats with their generally lower-energy conditions and less scope for physical reworking develop a profusion of natural history coastal features. For instance, there are the sedimentologic products of interactions between waves and tides (e.g., cross-laminated sand, ripple-laminated sand, lenticular bedding, flaser bedding, laminated mud, ripple-laminated silt lenses in clay), the products of interactions between sediments and biota (e.g., various burrow forms zoned tidally across the shore, various types of root-structuring, skeletal remains related to tidal levels), the geomorphic products of tides (e.g., tidal run-off on low-gradient slopes to form meandering tidal creeks), the effect of water temperature and salinity (Kroegel and Flemming 1998), and the products of hydrochemical interactions with sediments resulting in diagenetic products (e.g., dissolution of carbonate by acidic pore water; cemented crusts and their breccia and sand-sized intraclast derivatives; carbonate nodules; gypsum precipitates; and products of redox reactions such as biologically mediated precipitation of iron sulfide). For stratigraphers and students of sedimentary rocks, identifying tidal flats in the geologic record is often an important step in the reconstruction of palaeoenvironments, the location of facies associated with coastlines, and the recognition of such markers in stratigraphic sequences in basin analyses – tidal-flat signatures derived from studies of modern environments provide important analogs in such analyses.

Tidal flats have been of great interest also to biologists and ecologists – firstly, because these systems are habitats for a variety of biota; and secondly, because there are processes and products ranging from the macroscale (such as mangrove forests) to the microscopic (such as microbial communities), with biological responses to and, conversely, biological effects on the sedimentologic, geohydrologic, hydrologic, hydrochemical, and mineralogic features of the environment (Watzin 1983; Stal and de Brouwer 2003); the variation in substrates has an influence on the composition of

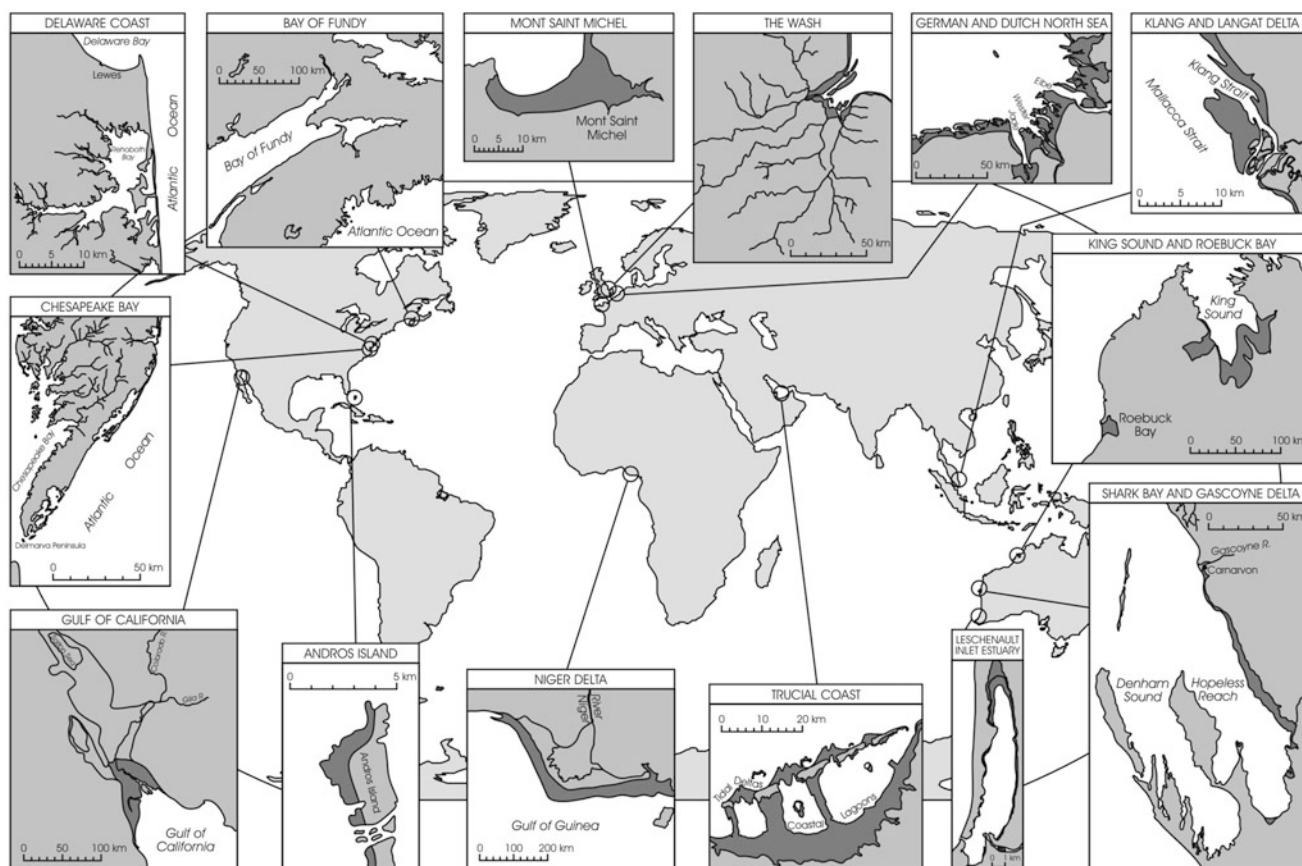
communities (Albrecht 1998; Semeniuk et al. 2000). Further, because of their biodiversity, ecological processing, and productivity, tidal flats are important in their function in the food chain in coastal zones, as food sources for migrating nekton (fish, crabs, snakes), demersal fish, and waterbirds (de Sylva 1975; Dankers et al. 1983; Wolff 1983; Reise 1985, 1991; Hutchins and Saenger 1987; Paterson et al. 2009), and as fish nurseries. In effect, tidal flats are an interactive system of sediments and hydrology/ hydrochemistry influencing biota, and biota effecting and structuring sediments. Again, with their biodiversity and ecological functioning, tidal flats are biologically more complex, and contrast with the ecologically simpler, steeper-gradient wave-dominated sandy shores.

Settings of Tidal Flats

Tidal flats around the globe occur in a variety of regional geomorphic settings (Fig. 1 and Table 1). Since they are surfaces exposed and inundated by tides, they may simply be part of larger coastal systems (Semeniuk 1996, 2008, 2015a; Fan 2012; Flemming 2012), that is, the shores of deltas, estuaries, lagoons, gulfs, bays, straits, rias, sounds, and cusped forelands. Alternatively, they may be the sole coastal form developed along an open coast or broad embayment, or may comprise wholly tidal lagoons leeward of barriers. The best-developed tidal flats occur along estuarine coasts, protected embayments, or barred lagoons where the shore slopes are gentle due to sediment accretion, and tides are large (Fig. 2). Along many coasts, tidal flats are part of prograded shores (Kendall and Skipwith 1968; Thompson 1968; Hagan and Logan 1975; Reineck and Singh 1980); but in some instances, they may comprise modern sediment veneers on wave- or tidal-cut unconformities on rock or Pleistocene sediment, or earlier Holocene sediments (Semeniuk 1981a).

Oceanographically and meteorologically, tidal flats can be tide-dominated, wave-dominated, mixed tide- and wave-influenced, cyclone- and storm-influenced, and/or strongly wind-influenced. Consequently, and depending on the type of sediment delivery, they can be sandy tidal flats, muddy tidal flats, or sedimentologically mixed and zoned tidal flats, with implications for varied sedimentary structures developed across the tidal gradient reflecting the availability of grain sizes and the effects of tides, waves, storms, and wind (Reineck and Singh 1980; Flemming 2012; Zhu et al. 2014).

Sediment types and sedimentation style on tidal flats can also be influenced by climate, viz., tropical humid or tropical arid conditions, at one extreme to boreal and arctic conditions at the other (with ice interacting with coastal deposits, or annual freezing of coastal deposits; cf. Reineck and Singh 1980; Dionne 1998) which would influence rainfall, run-off volumes, wind direction and intensities, storms, water



Tidal Flats, Fig. 1 Location, settings, and sizes of various tidal flats around the globe. Table 1 provides more detailed information and references

temperatures, evaporation, style of diagenesis, and effect of biodiversity, amongst others.

The variety of tidal flats, their substrates, and their various oceanographic and geomorphic settings lend themselves to various types of classification. For instance, Semeniuk (1981b) classified the tidal flats of King Sound into categories based on underlying stratigraphy and local coastal geomorphology, and Fan et al. (2013) classified them and their associated sedimentary features and facies into nine types that presented a continuum of open-coast depositional settings from tidally dominated muddy tidal flats with wave influence through to sandy tidal flats of mixed energy (tide-dominated) to tidal beaches of mixed energy (wave-dominated) to wave-dominated beaches with tide influence.

Tides and Tidal Levels

The tidal ranges that expose tidal flats vary globally from less than 1 m to *ca* 15 m amplitude, and are diurnal (one tide daily), semidiurnal (two tides daily), or mixed (two tides daily, but with inequality between tide maxima and tide minima across the day). Over a lunar cycle, tides vary from a lower amplitude neap range (during quarter and three-

quarter Moon phases) to a higher amplitude spring range (during new and full Moon phases). Higher than normal tides occur annually during equinoctial periods and on an 18.6-year turnaround in response to the Lunar Nodal Periodicity. As a result and depending on shore slope, tidal flat width may vary from being a narrow coastal strip to being broad and expansive coastal forms.

Part of the coast emergent during low tide and submerged during high tide is the *intertidal zone*; that part of the coast permanently submerged below the low-water line is the *subtidal zone*; that occurring above the zone of high-tide inundation is the *supratidal zone* (Fig. 3). Some authors consider the “supratidal zone” as the zone above the mean high-water line but sometimes under water during extremely high tides, or even spring tides, but it is preferable to refer to all gently inclined surfaces and terrain *above* the highest tides as supratidal, and to treat all surfaces flooded by neap, spring, and equinoctial spring tides as intertidal, and to separate these various tidal zones and levels.

Tidal ranges have been classed by Davies (1980) into three groups: *microtidal* <2 m, *mesotidal* 2–4 m, and *macrotidal* >4 m. While this classification generally has been accepted, large tidal ranges >8 m might be further classed as *extreme macrotidal*. Tidal range amplification may occur due to bay

Tidal Flats, Table 1 Some well-known and documented tidal flats ordered in tidal range

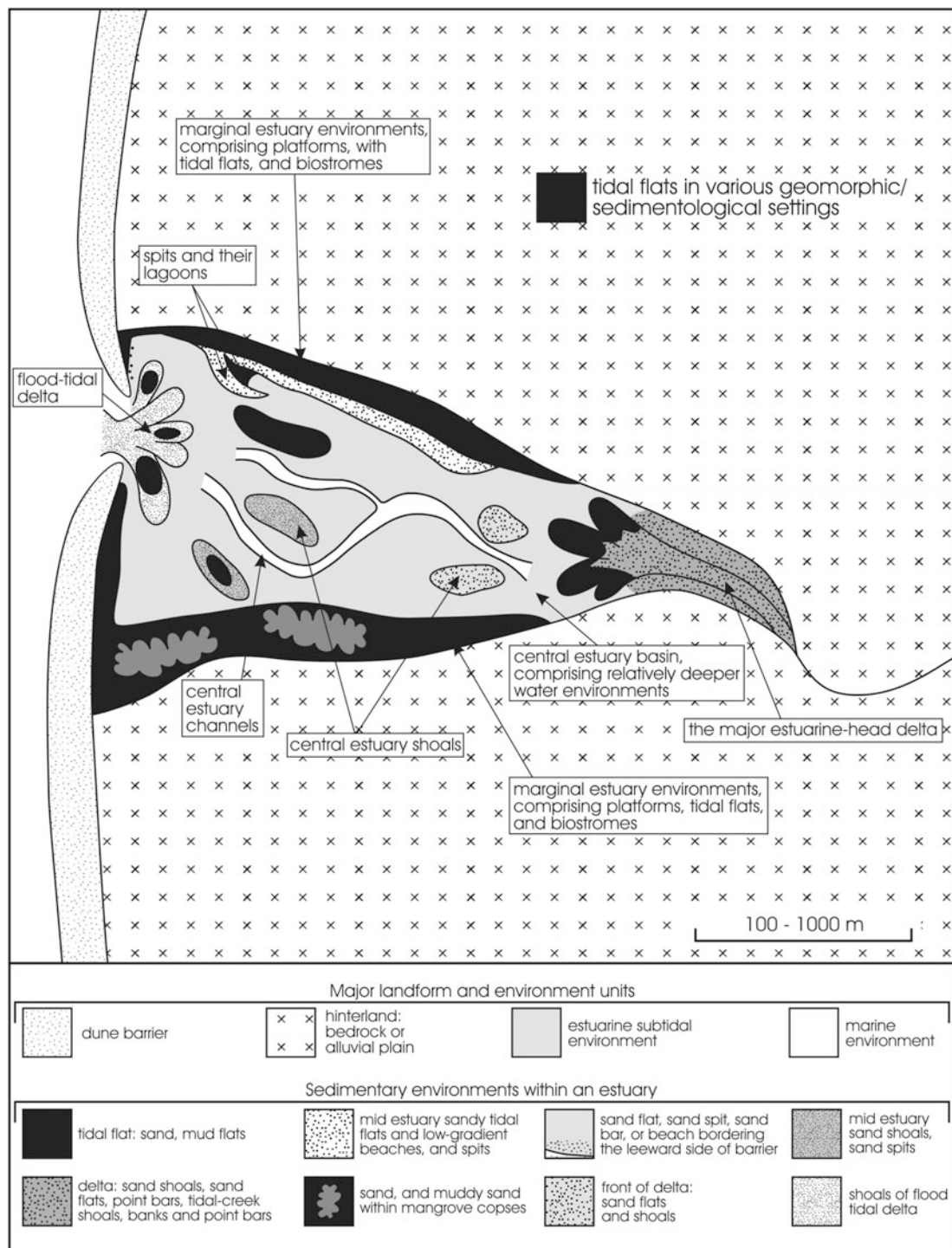
Tidal flat location	Tidal range (m)	Composition	Tidal flat setting
Bay of Fundy (Nova Scotia)	15.0 m Extreme macrotidal	Siliciclastic	Broad tidal flats of gravel, sand, and mud, with local salt marsh peripheral to an estuarine gulf in a humid temperate climate (Knight and Dalrymple 1975)
Bay of Mont St Michel (NW France)	15.0 m Extreme macrotidal	Siliciclastic	Broad tidal flats of sand and mud, with salt marsh within a complex of a funnel-shaped estuary in humid temperate climate (Larsonneur 1975)
King sound (NW Australia)	11.0 m Extreme macrotidal	Siliciclastic	Broad tidal flats of sand and mud, with some cheniers, and erosional tidal channels fringed by mangroves; peripheral to a seasonal estuarine gulf in a semi-arid tropical climate (Semeniuk 1981a)
Roebuck Bay (NW Australia)	10.5 m Extreme macrotidal	Carbonate	Broad tidal flats along semi-sheltered embayment, dominated by mud; erosional tidal channels, fringed by mangroves; in a semi-arid subtropical climate (Semeniuk 1993, 2008)
Gulf of California (USA)	6–8 to 10 m Macrotidal to extreme macrotidal	Siliciclastic	Broad tidal flats dominated by mud, with intermittent beach ridges, with salt marsh; part of the Colorado River Delta within a gulf in a semi-arid subtropical climate (Thompson 1968)
The Wash (England)	7 m Macrotidal	Siliciclastic	Broad tidal flats of sand and mud, with salt marsh; along the shore of a large embayment in a humid temperate climate (Evans 1975).
Delta of the Klang and Langat Rivers (Malaysia)	4.5 m Macrotidal	Siliciclastic	Compound delta with insular/peninsular development of tidal flats, traversed by tidal creeks with mangrove; in a humid tropical climate (Coleman et al. 1970)
The Jade and the Dutch Wadden Sea (North Sea)	2.6–4.1 m Mesotidal	Siliciclastic	Broad tidal flats of sand and mud, with salt marsh, leeward of barriers in a humid temperate climate (Reineck 1975)
Niger Delta (western Africa)	1.0–2.8 m Microtidal	Siliciclastic	Extensive mangrove-vegetated tidal flats of mud and sand developed behind a beach barrier in a delta system in a humid tropical climate (Allen 1970)
Gascoyne Delta (Western Australia)	2 m Microtidal	Siliciclastic	Local narrow tidal flats of sand and mud, with mangrove fringed lagoons in a delta in an arid subtropical climate (Johnson 1982)
Trucial Coast (western Persian Gulf)	2 m Microtidal	Carbonate	Broad tidal flats of carbonate muds and gypsum, with algal mats, salt marsh and mangroves, shoreward of prograding carbonate complex fringing a large gulf in an arid tropical climate (Purser 1973)
Chesapeake Bay (eastern USA)	1.5–2.1 m Microtidal	Siliciclastic	Narrow to broad tidal flats, with salt marsh, within inlets and along the shore of an estuary in a humid temperate climate (Stevenson et al. 1985; Lippson and Lippson 1997)
Delmarva Peninsula to Sapelo Island (eastern USA)	<1 m Microtidal	Siliciclastic	Broad protected tidal flats, with salt marsh, leeward of barriers in a humid subtropical climate (Howard et al. 1972; Harrison 1975)
Andros Island (Bahamas)	0.5 m Microtidal	Carbonate	Broad tidal flats of pelleted mud, with mangrove, salt marsh and algal mats, developed capping the Bahama Bank Carbonate Complex in a humid subtropical climate (Shinn et al. 1969)
Shark Bay (Western Australia)	0.5 m Microtidal	Carbonate	Local tidal flat of sand or pelleted mud, with salt marsh, algal mats, and stromatolites, shoreward of prograding seagrass banks and hypersaline platforms in large, elongate embayments in an arid subtropical climate (Hagan and Logan 1975)
Leschenault Inlet estuary (Western Australia)	0.5 m Microtidal	Siliciclastic	Low-tidal sand, muddy sand / mud bordered by high-tidal salt-marsh-vegetated mud; locally with mangroves; tidal flats border an elongate estuarine lagoon in a humid subtropical climate (Semeniuk et al. 2000)

geometry and coastal constriction, for example, the Bay of Fundy in Nova Scotia, which, because of its basin geometry, amplifies the tide from *ca* 5.4 m at the entrance to the bay to 15 m at its head.

Zoning on Tidal Flats

Tidal flats are typically zoned in terms of geomorphology, sediments, hydrology, hydrochemistry, and biota in

response to gradients of inundation frequency (and conversely, extent of exposure), hydrodynamic energy (wave and tidal-current energy), and salinity. The best zonation is manifest sedimentologically and biologically in response to hydrodynamic variations and physicochemical gradients, respectively (Semeniuk 1983, 2015b). Sedimentologic and biologic zones across the tidal flats are best exhibited in macrotidal settings where there are marked distinctions in slope, sediments, and biology within the interval of the tidal range responding to inundation frequency, wave and

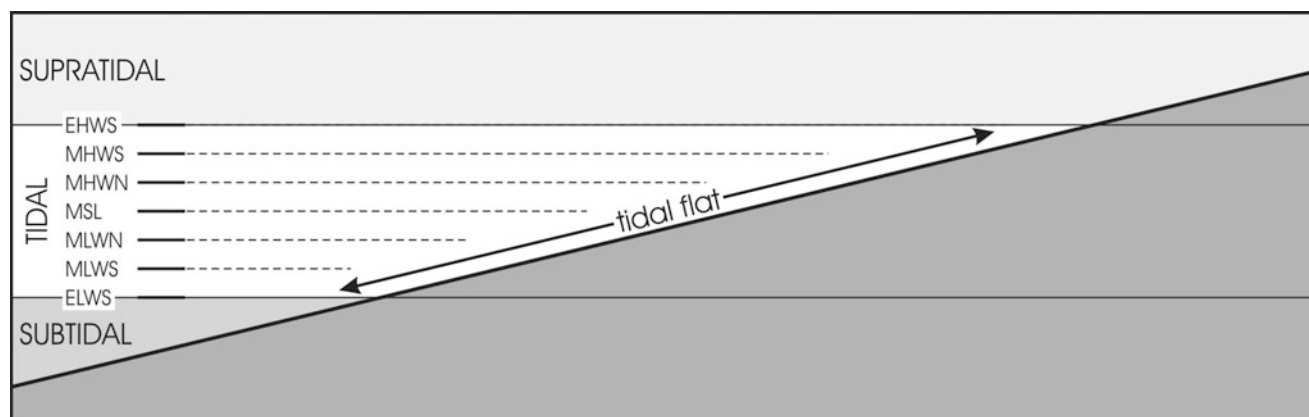


Tidal Flats, Fig. 2 Typical geomorphic and sedimentologic location of tidal flats within an idealized estuary. (From Semeniuk 2015a)

tidal-current energy, and pore-water salinity. On microtidal flats, these various differences related to tidal levels are less pronounced.

Various levels within a tidal flat, often delineated by sediment and/or biological zones, can be distinguished as follows:

- Low tidal flats – exposed by the mean and extreme low spring tides, generally underlain by sand, and vegetation-free
- Middle tidal flats – the flats and low-gradient slopes centered around mean sea level, exposed and inundated by neap tides; the upper parts of these flats may be vegetated



Tidal Flats, Fig. 3 Tidal flat in profile showing the various tidal levels (inclination of tidal flat is exaggerated). Located between the supratidal and subtidal zones, the tidal flat has the following levels: the lowest tidal level at equinoctial low-water spring tide (ELWS), mean low-water

spring tide (MLWS), mean low-water neap tide (MLN), mean sea level (MSL), mean high-water neap tide (MHN), mean high-water spring tide (MHWS), and the highest tide level at equinoctial high-water spring tide (EHWS). (From Semeniuk 2015a)

by samphire in temperate latitude areas, and by mangrove in tropical latitudes

- High tidal flats – inundated by the mean and extreme high spring tides, generally underlain by mud, and vegetated by salt marsh or mangrove, or, in more arid settings, vegetation-free and salt-encrusted (salt flat)

Typical cross sections through some macrotidal to microtidal flats are shown in Fig. 4.

Geomorphic Features of Tidal Flats

The macroscale, mesoscale, and smaller-scale features are the result of wave action or tidal currents, the strength of the waves and tidal currents, and the duration that part of the tidal flat is subject to waves or tidal currents. At the macroscale, the surface of a tidal flat generally is flat to gently inclined, but there may be a range of mesoscale to microscale features therein (Figs. 4 and 5; Table 2) reflecting the effects of position within either the low or high spring tidal zones, or the neap tidal zone. For example, the low tidal zone may be nearly flat or very gently inclined, the middle tidal flat may be more moderately inclined, and the high tidal flat again may be nearly flat or very gently inclined.

At mesoscale, the geomorphic features of tidal flats may include plain flats; local cliffs; spits; cheniers; sand waves; shell mounds; skeletal reefs; and gullies, channels, and creeks (also called tidal creeks). Cliffs, commonly cut into mud, can separate vegetated and vegetation-free plain flat zones, but some cliffs are formed due to either the effect of wave energy concentrated at a specific tidal level or the undercutting of mud through erosion of underlying sand. Tidal creeks may be ramifying or meandering, with point bars and steep banks. At smaller scales, the surface of tidal flats may be planar and

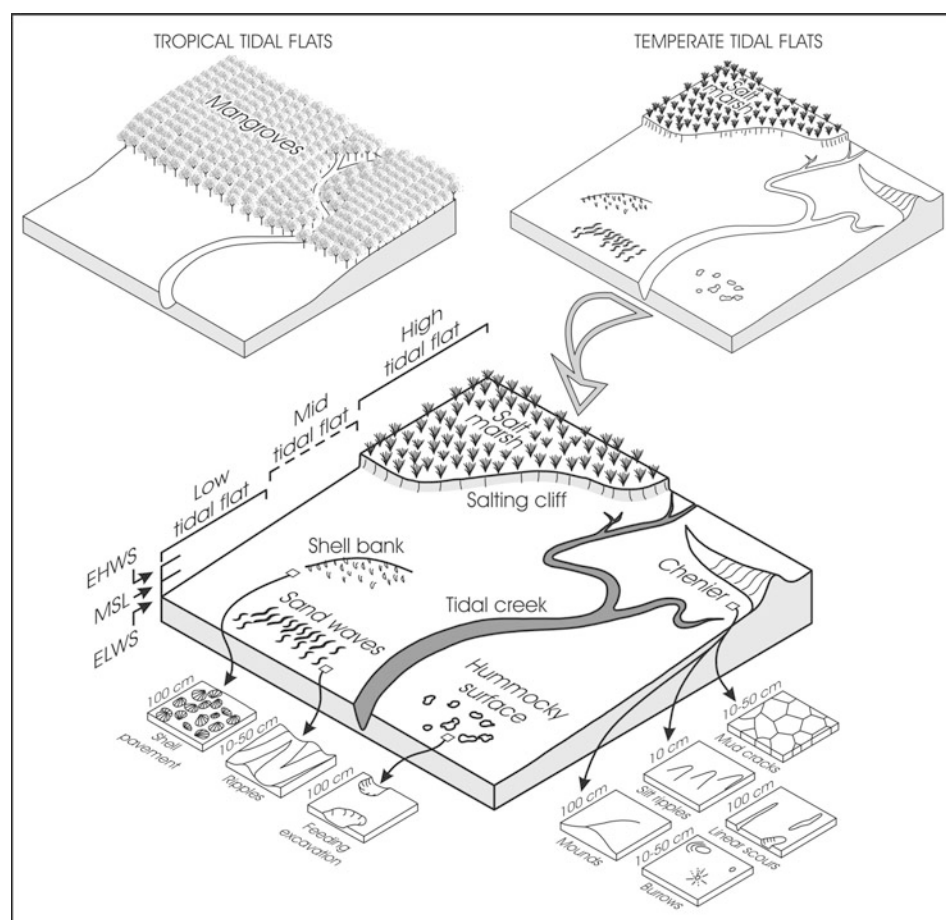
smooth; or hummocky to slightly irregular; or may exhibit linear scours, fish and ray excavation hollows, snail, worm and crab tracks, feeding pellets, exhalent mounds of worm or shrimp burrows, carbonate crust mounding and teepees, desiccation polygons, or mud cracks; or may exhibit silt ripples on a clay floor.

In a geomorphic overview, recently, Zhou et al. (2016) numerically modelled tidal flats in relation to sediments and vegetation with interactions between tides, waves, salt marshes, sediment transport, and sea level rise to determine and predict of tidal flat profile shape and sediment distribution – the tidal flats were depositional or erosional, and varying from convex to concave in profile depending on tide versus wave action, and presence of vegetation.

Tidal Flats and Their Particle Sizes and Sediment Composition

Tidal flats may be underlain by mud, sand, rock gravel, shell pavements and, locally, rock pavements, or mixtures of these. Often, where all particle sizes are present, there is a zonation of sediment types across the flats, and an interlayering at a specific tidal level, but in many instances, one sediment type may dominate across the entire tidal flat. This partitioning of sediments across the tidal flat lends itself to a classification of tidal flats, or zones within tidal flats, according to particle size. For example, those composed wholly of mud may be termed muddy tidal flats, and those composed wholly of sand are sandy tidal flats. Tidal level zones within the tidal flat may be classed according to substrate, for example, sandy low tidal flats, muddy high tidal flats. Many sandy tidal flats, when exposed at low tide, have a glistening film of wet clay on the surface which gives the impression that they are mud flats – scraping away the film of clay will reveal an underlying

Tidal Flats, Fig. 4 Generalised geomorphology of tidal flats. Macroscale geomorphology of tropical-climate tidal flats with mangroves and temperate-climate tidal flats with salt marsh. More detail of mesoscale and microscale features is shown for a temperate climate tidal flat

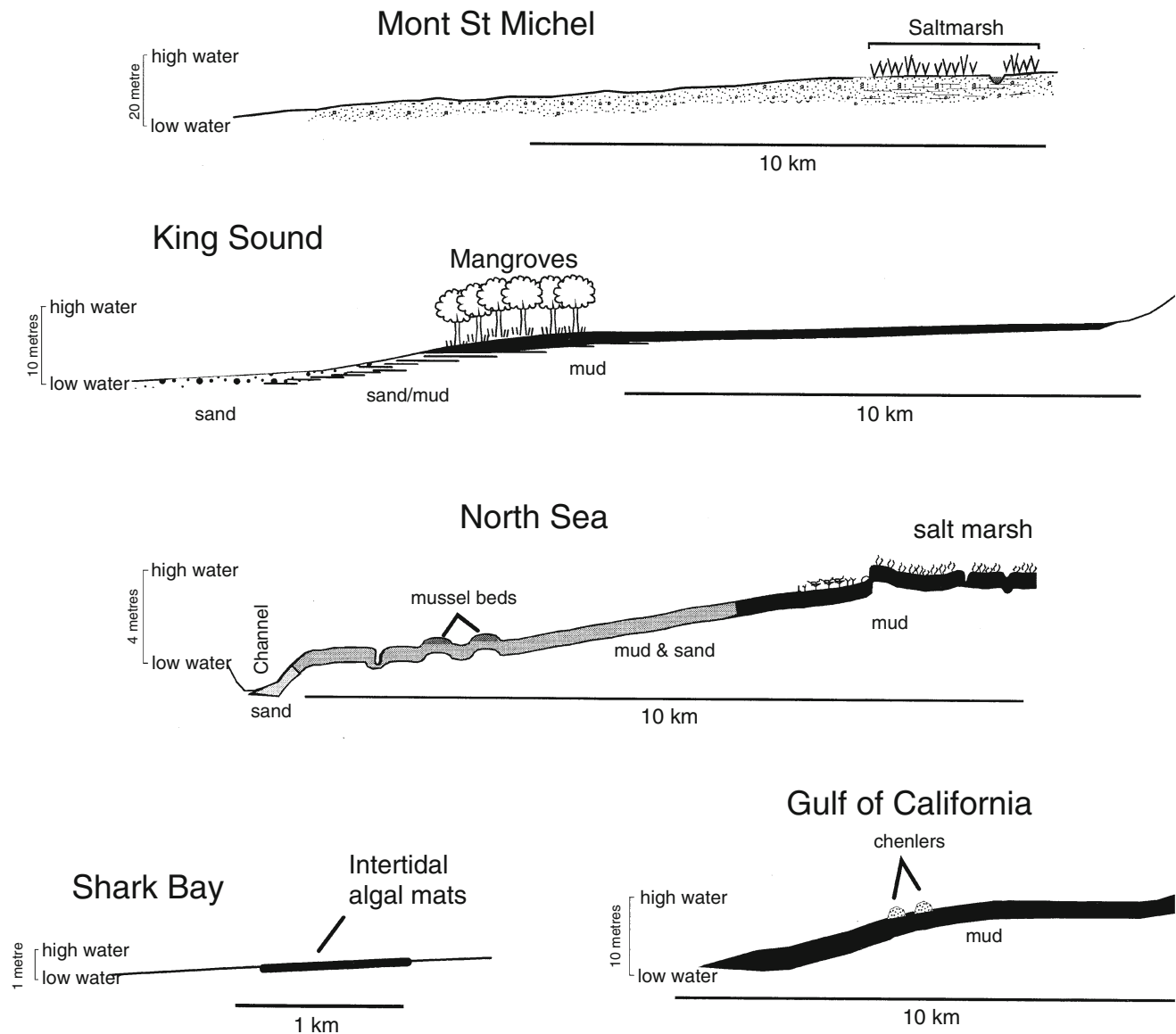


sand substrate and, as such, many a sandy tidal flat with this wet clay surface film has erroneously been called a “mud flat.” A range of possible tidal flat types based on substrate and tidal zone slopes is shown in Fig. 6. A range of possible tidal flat types based on substrate, with field examples, is presented in Table 3.

In regard to sediment composition, two major groups are recognized: *siliciclastic tidal flats* composed of terrigenous sediments such as quartz sand, quartz silt, and phyllosilicate clay, and *carbonate tidal flats* composed of carbonate silt and clay, various sand-sized carbonate grains, and products of cementation (e.g., crusts, breccias, intraclasts). These major groups reflect two extremes in settings: an abundant supply of terrigenous sediment to the tidal coast such as in deltas or estuaries versus a low supply relative to the rate of carbonate sediment production (as along terrigenous sediment-starved coasts). From a historical perspective, the majority of earlier investigations of tidal flats were centered on siliciclastic systems, and much information emerged from studies in the North Sea (Reineck 1972; Evans 1975). Later, as interest in carbonate rocks grew during the 1960s, linked to their petroleum reservoir potential, a range of studies was undertaken in carbonate tidal flats (Shinn et al. 1969; Purser 1973; Hagan and Logan 1975; Shinn 1983).

Generally, regardless of whether the tidal flats are dominantly siliciclastic or carbonate, their sediments commonly contain both siliciclastic *and* carbonate particles. In dominantly siliciclastic settings, there may be minor to moderate carbonate components of shell gravel, shell grit, skeletal sand (e.g., shell fragments, foraminifera), skeletal silt-sized material, and carbonate clay transported to or generated on the flats. Similarly, in dominantly carbonate environments, there may be siliciclastic sand, mud, or gravel from oceanic, aeolian, or local eroding sources. The range and origin of mud, sand, and gravel-sized particles comprising tidal-flat sediments are noted in Table 4.

Some of the best-known siliciclastic tidal flats are the North Sea coast (for example, the Jade and the Dutch Wadden Sea, with detailed information on sediment dynamics, sedimentary structures, sediment types, and tidal-flat stratigraphy), The Wash in south-eastern England, the Gulf of California, the Bay of Fundy, the compound high-tidal delta of the Klang and Langat Rivers, King Sound in north-western Australia, Bay of Mont St Michel in France (Postma 1961; Klein 1963; Thompson 1968; Allen 1970; Coleman et al. 1970; Reineck 1972; Evans 1975; Larssonneur 1975; Semeniuk 1981a; Amos 1995). With most of these examples,



Tidal Flats, Fig. 5 Profiles across various macrotidal to microtidal siliciclastic and carbonate tidal flats (see Table 1) showing nature of slopes, sediments, vegetation, or tidal flat morphology. (From Semeniuk 2005)

there is a grain-size variation across the flats from sand in low-tidal zones, with specific biogenic contributions in particular tidal zones depending on climate setting and biogeography, and sediment types and sedimentary structures are dominantly the result of physical and biologic processes. With increase upslope in pore water salinity, particularly in semiarid and arid climates, the upper parts of siliciclastic tidal flats may develop carbonate nodules or gypsum crystals, or be salt-encrusted.

Carbonate tidal flats generally occur in mid- to low-latitude warm climates. The best known are Andros Island of the Bahama Banks (Shinn et al. 1969), the Trucial Coast along the Persian Gulf Coast (Kendall and Skipwith 1968; Purser 1973), and Shark Bay in

northwestern Australia (Hagan and Logan 1975). An example of macrotidal carbonate tidal flat is Roebuck Bay (Semeniuk 2008). In all these examples, there is little or no terrigenous influx from terrestrial sources to dilute the carbonate accumulation contributed by local biogenic and abiotic sources, and hence the sediments are carbonate-rich. There is a range of diagnostic sediments and structures formed on carbonate tidal flats as a result of tidal deposition, biogenic contributions and alteration, and primary and secondary effects of cementation. Cementation of sediments and formation of their (secondary) structural and sedimentary derivatives is an important and common feature on upper parts of carbonate tidal flats. Under conditions of hypersalinity on the

Tidal Flats, Table 2 Geomorphic features of tidal flats

Geomorphic or surface feature	Origin
Mesoscale surface features (>metre-sized, up to tens of metres long)	
Meandering gullies, channels, creeks, meandering or ramifying	Tidal erosion with local deposition on point bars
Crust-lined and locally brecciated meandering channels	Tidal erosion with local deposition on point bars with mineral precipitation in surface sediments and resultant surface crust expansion
Sand waves	Large bodies of sand developed in low-tidal zones
Spits	Shoestring and sandy gravel body across tidal flat from local headland formed by tidal currents and wave action
Cheniers	Isolated shoestring sand and sandy gravel body on tidal flat, variably formed by tidal currents, wave action, and storms/cyclones
Salting cliff	Small cliff 30–100 cm high, cut into salt marsh, marking junction between high-tidal salt marsh and vegetation-free mid-tidal to low-tidal flat
Mid-tidal cliff	Small cliff up to 100 cm high, marking junction between mangrove front to <i>ca</i> MSL and vegetation-free mid-tidal to low-tidal flat
Microscale surface features (<metre-sized)	
Smooth planar surface	Deposition on, and erosion of the surface
Linear scours (mm to cm deep)	Tidal erosion
Slightly irregular	Tidal erosion of the surface, and/or bioexcavations by small biota and fish
Hummocky surface	Tidal erosion of the surface, and/or excavations by stingrays, fish and large burrowing benthos
Pot-holed surface	Excavations by stingrays, fish, large burrowing benthos
Mud cracks	Desiccation
Surface moundings grading to teepees and brecciation	Mineral precipitation in surface sediments and resultant surface crust expansion
Mounded surface	Mineral precipitation in surface sediments

higher zones of such tidal flats, precipitation of carbonate minerals often is prevalent, and in contrast to siliciclastic tidal flats, since there is an abundance of carbonate grains to act as nuclei for interstitial cements, there is a plethora of diagenetic and sedimentary products such as cemented layers and crust development, progressing to surface mounding, formation of compressional polygons and teepees, and then fragmentation, brecciation, and the formation of intraclasts. Carbonate tidal flats set in the more

arid climates also develop evaporitic mineral suites such as beds of gypsum nodules, gypsum platey crystals, gypsum mud, and halite crusts (Kendall and Skipwith 1968; Hagan and Logan 1975).

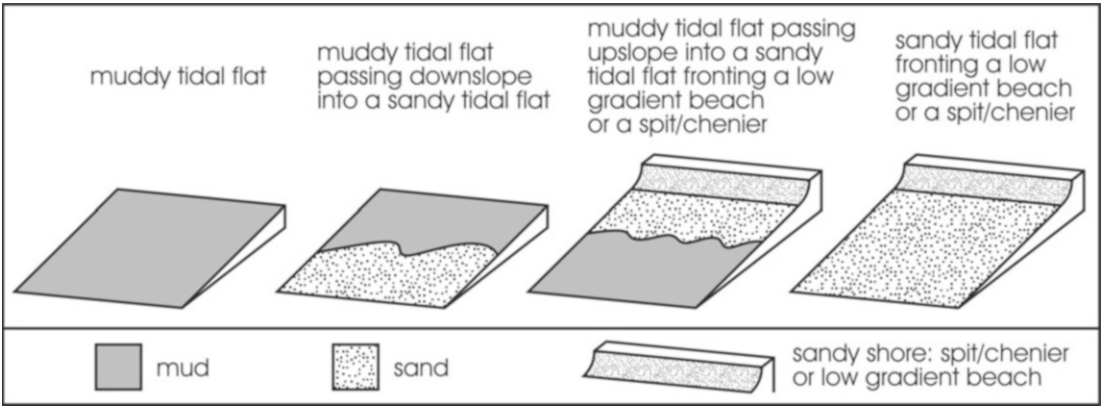
Sedimentology, Sedimentary Structures, and Stratigraphic Sequences

Sediment bedforms, surface features, and near-surface features on tidal flats are produced by oceanographic, and other physical, biotic, and hydrochemical processes. Wave action, tides, and winnowing result in ripples, megaripples, sand waves, sandy plane beds, linear scours, plane mud beds, and gravel pavements. A range of other physical processes result in mud cracks, air escape holes, and bubble structures. Biological activity results in burrow-pocked surfaces, animal tracks, invertebrate and fish burrow structures, fish and ray excavation feeding depressions, crab burrow workings, vesicular structures, crab balls, and accumulation of shell banks and shell gravel. Chemical and physical processes combine to develop sheets of gypsum mush, gypsum crystal boxwork and gypsum nodules, gypsum crystals embedded in sediment, platey gypsum pavements, carbonate crusts, and intraclast breccia pavements.

Sedimentary structures deriving from the burial of sediment bedforms, and the surface and near-surface features include cross-bedding and cross-lamination, herring bone cross-lamination, sand ripple cross-lamination, silt ripple cross-lamination, lenticular bedding, flaser bedding, laminated mud, sand dykes, mud dykes, bubble sand, vesicular mud, root-structuring, vertical burrows, u-burrows, to labyrinthoid burrow networks, shell laminae and beds, shell reefs, silt and sand balls, bioturbation and swirl structures, breccias, nodular gypsum beds, platey gypsum beds, and teepee structures (Fig. 7).

Key sediments, diagnostic of their formative processes, occur in different parts of the tidal flat. For example, mangrove-vegetated muddy tidal flats develop root-structured, burrow-structured, and bioturbated (shelly) mud and crustacean-dominated and polychaete-dominated mixed tidal flats develop burrow-structured, interbedded sand and mud varying to bioturbated muddy sand.

With progradation, siliciclastic tidal flats and carbonate tidal flats develop characteristic stratigraphic sequences (Kendall and Skipwith 1968; Thompson 1968; Coleman et al. 1970; Evans 1975; Hagan and Logan 1975; Harrison 1975; Larssonneur 1975; Semeniuk 1981a, b, 1996, 2008; Shinn 1983; Berkeley and Rankey 2012; Maloof and Grotzinger 2012). A range of stratigraphic sequences is shown in Fig. 8, from various macrotidal to microtidal settings, from



Tidal Flats, Fig. 6 Idealized diagram showing a range of tidal flats underlain by mud, or mud grading down-slope to sand flats, muddy tidal flat grading upslope to sand flats fronting a low-gradient beach or sandy spit/chenier, and sandy tidal flat fronting a low-gradient beach or sandy spit/chenier. (From Semeniuk 2015a)

Tidal Flats, Table 3 Tidal flat types according to substrate

Substrate type underlying tidal flat	Tidal flat type	Examples
Whole tidal flats		
Mix of particle sizes across the whole tidal flat, or differentiation not intended	Tidal flat	North sea; The Wash; Bay of Mont St Michel; King Sound
Tidal flats wholly underlain mainly by:		
Mud	Muddy tidal flat	Gulf of California
Sand	Sandy tidal flat	Southern Tasmania tidal flats
Gravel	Gravelly tidal flat	Parts of the Bay of Fundy
Shelly pavement across tidal flat	Tidal shell pavement	Parts of Shark Bay
Specific zones on tidal flats		
High-tidal flats wholly underlain mainly by mud	Muddy high-tidal flat	King Sound
Mid-tidal flats wholly underlain mainly by mud	Muddy mid-tidal flat	King Sound
Mid-tidal flats underlain mainly by mud and sand	Mixed mid-tidal flat	North Sea
Low tidal flats wholly underlain mainly by sand	Sandy low-tidal flat	North Sea; Bay of Mont St Michel
Low tidal flat underlain by pavement of shell	Low-tidal shell pavement	Parts of Shark Bay
High-tidal flat underlain by crust pavement	High-tidal crust pavement	Parts of Shark Bay
High-tidal flat underlain by breccia	High-tidal breccia pavement	Dampier Archipelago; Shark Bay

flats that are mud-dominated, to sand-to-mud sequences, from temperate to tropical climates. Some examples of sediments and the processes involved in their development from siliciclastic tidal flats and carbonate tidal flats are noted in Table 5.

Hydrology and Hydrochemistry

The groundwater hydrology and hydrochemistry of tidal flats, expressed as the hydrological variation and in the salinity gradient of the watertable under the tidal flat at low tide, or pellicular water in the sediment at low tide (Fig. 9) are important for several reasons. Interstitial pore-water salinity gradients and moisture gradients, for instance, influence macrophytes (such as mangroves and samphires), microbial mats, and invertebrate biota in relation to their occurrence and zonation (Semeniuk

1983, 2015a). Interstitial pore water salinity and moisture gradients also influence precipitation of evaporitic minerals. Microscale shallow groundwater hydrologic recharges and discharges influence development of sedimentary structures (e.g., seepage zones out of sand mounds to initiate sand erosion, or to initiate hydrochemical exchanges and cementation; formation of bubble sand). The hydrologic functioning of tidal flats additionally can drive geochemical processes that diagenetically modify sediments (e.g., formation of iron sulfide precipitation to form grey sediments or the oxidation of buried iron-sulfide-impregnated vegetation to form goethitic pseudomorphs).

Tidal flat groundwater levels fluctuate on a diurnal to semidiurnal basis following the tides, with a dampened effect from mid-tidal levels to upslope. All tidal flat groundwater rises during flood tide and, of course, is inundated on high tide. Recharge and discharge, and lateral groundwater flow

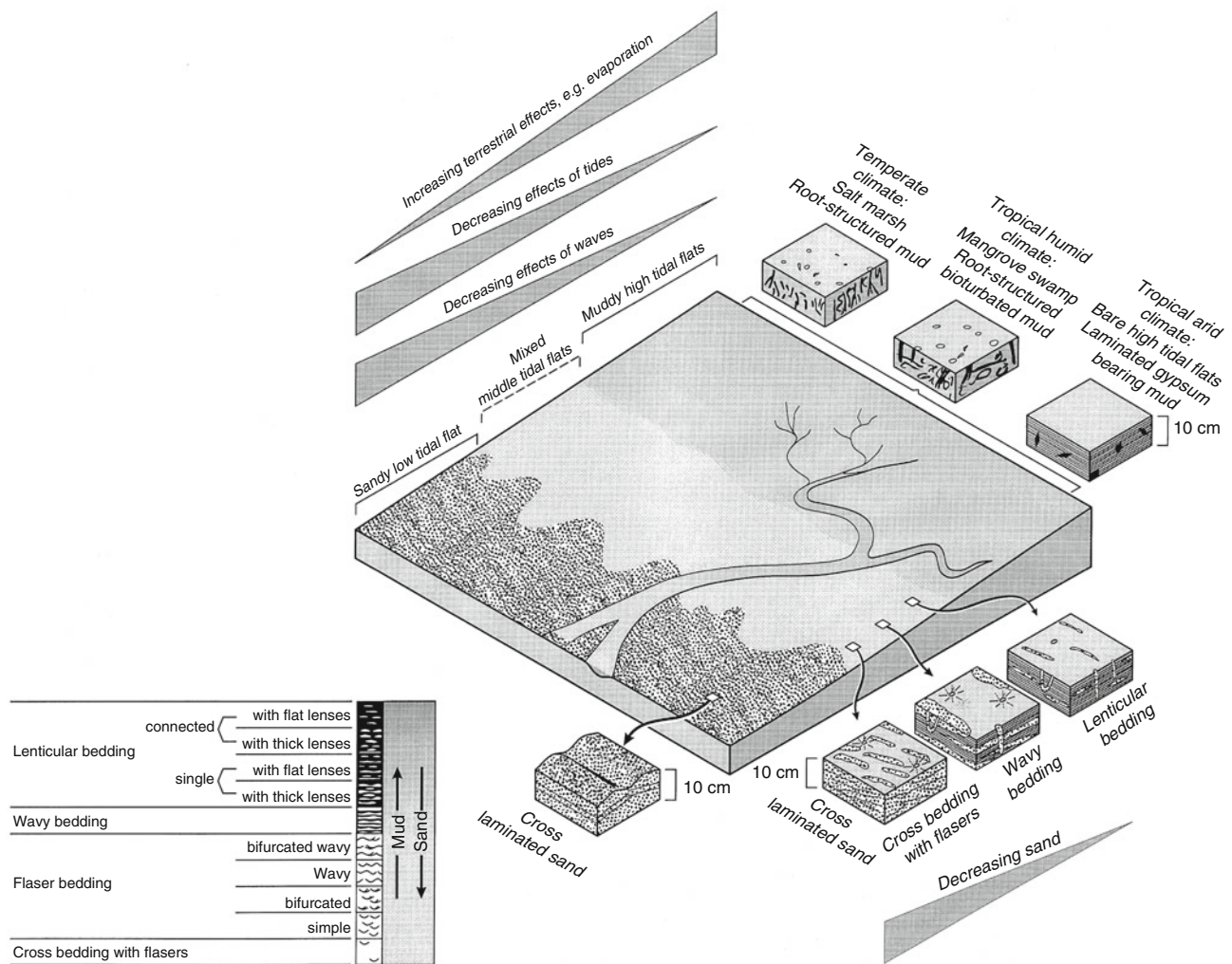
Tidal Flats, Table 4 Types of sedimentary particles on tidal flats and their origin

Sediment particle	Origin (information from various tidal flats)
Clay (<4 µm particle size)	
Phyllosilicate clay (kaolinite, illite, montmorillonite)	1. Fluvially delivered to the coastal system 2. Reworked Pleistocene coastal deposits 3. Reworking of glacial deposits 4. Aeolian
Calcite and aragonite clay	1. Reworked, comminuted skeletons 2. Precipitated from seawater 3. Disintegrated calcareous algae
Goethite	1. Fluvially delivered to the coastal system 2. Aeolian
Quartz clay	Aeolian
Amorphous silica	Diatom (in situ or transported)
Silt (4–63 µm particle size)	
Quartz, feldspar, various silicate minerals	1. Fluvially delivered to the coastal system 2. Reworked Pleistocene coastal deposits 3. Reworking of glacial deposits 4. Aeolian
Skeletal silt	Reworked and in situ comminuted shelly exoskeletons
Amorphous silica	Diatom (in situ or transported)
Sand (63–2000 µm particle size)	
Quartz, feldspar, various silicate minerals, rock fragments	1. Fluvially delivered to the coastal system 2. Reworked Pleistocene coastal deposits 3. Reworking of glacial deposits 4. Aeolian
Skeletal sand	Comminuted to whole reworked and in situ exoskeletons, e.g., shell fragments, foraminifera
Carbonate sand (ooids, pellets)	Generated nearshore and reworked onto tidal flat, and for pellets, carbonate grain destruction by boring algae
Carbonate intraclast sand	Reworking of cemented carbonate crusts
Gravel (<2000 µm particle size)	
Quartz pebbles, rock fragments	1. Fluvially delivered to the coastal system 2. Reworked Pleistocene coastal deposits 3. Reworking of glacial deposits
Mud pebbles and cobbles	Eroded tidal mud
Armoured mud balls	Mud pebbles and cobbles with adhering gravel and shell
Skeletal gravel	Comminuted to whole, reworked and in situ shell
Carbonate intraclast gravel	Reworking of cemented carbonate crusts

through tidal-flat sediments may be facilitated by specific lithologic layers, or stratigraphic intervals, and at the small scale by burrow and root structures.

Groundwater salinity across tidal flats is commonly zoned, generally with near-marine water salinities at about mean sea level, grading to hypersaline and extremely hypersaline upslope (unless the marine waters fronting the tidal flats are hypersaline as at Shark Bay, Australia). In humid wet climates where there is discharge of freshwater into the coastal zone, tidal flat groundwater becomes fresh or brackish where tidal flats adjoin terrestrial freshwater (Semeniuk 1983, 2015a). The main source waters for groundwater of tidal flats are marine water, rain, and (through seepage and land overflow) land-derived freshwater. These water sources recharge, reside in, and interact with the stratigraphic framework underlying the tidal flat (Fig. 10).

Evaporation, macrophyte transpiration, and increasing infrequency of tidal inundation upslope combine to develop a gradient of increasing salinity across tidal flats (Semeniuk 1983). This gradient results in the zonation of biota, exemplified by zonation of mangroves, and zonation of evaporitic minerals and pore-water precipitates. Where marine waters are oceanic (*ca* 35,000 ppm salinity) and evaporation is extreme, high-tidal groundwater may reach 100,000–200,000 ppm salinity, that is, carbonate mineral- and gypsum-precipitating, but where source waters are already hypersaline, tidal-flat groundwater reaches up to *ca* 300,000 ppm salinity, resulting in the precipitation of halite. Hypersaline tidal flats with precipitation of gypsum nodules, gypsum platy crystals, gypsum mud, and halite crusts have been recorded by Kendall and Skipwith (1968), Thompson (1968), Hagan and Logan (1975), and Semeniuk (1981a, b).



Tidal Flats, Fig. 7 Generalised sedimentology of a typical tidal flat, relating sediment types to oceanographic and terrestrial processes, and inset detail of some sediment types in relation to facies setting and

position on tidal flat. (From Semeniuk 2005). Also shown in the systematic variation in bedding types as the sand to mud ratio changes. (Modified after Reineck and Singh 1980)

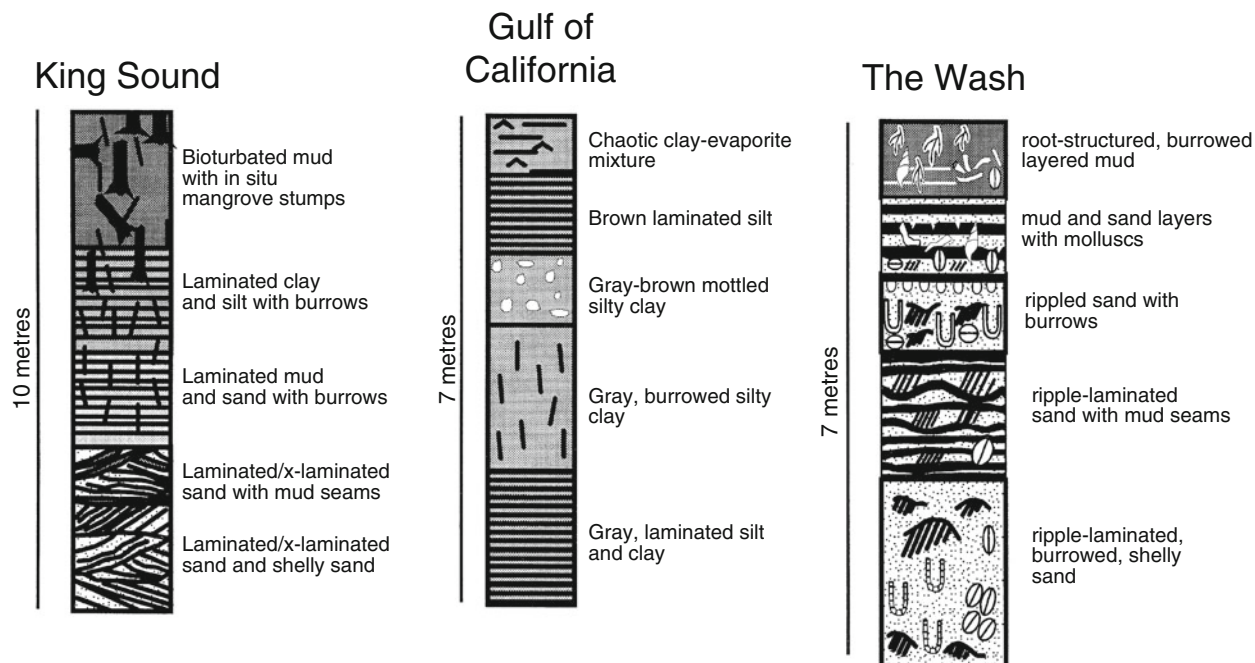
Some Key Biota of Tidal Flats

Depending on climate, tidal level, substrate, hydrology, and salinity, tidal flats may be inhabited in parts by salt marsh, mangroves, seagrass, algal mats, microbial mats, and biofilms (Figs. 11 and 12), as well as mussel beds, oyster beds and reefs, and worm-tube beds and reefs, and by a burrowing benthos of molluscs, polychaetes, and crustacea (Reise 1991; Little 2000; Spalding et al. 2010) and a meiofauna (also called meiobenthos) and microbiota including diatoms, foraminifera, and bacteria (Mithavkar and Anil 2004; Coull 2009). As biologically productive areas, they are feeding grounds for nekton, demersal fish, and shore birds (Fig. 13). Skeletal organisms such as mussels and oysters can form banks, reefs, biostromes, and bioherms on the tidal flat. Mussels and oysters, where they are abundant enough to form continuous sheets over the tidal flat surface, develop distinctive sedimentary deposits

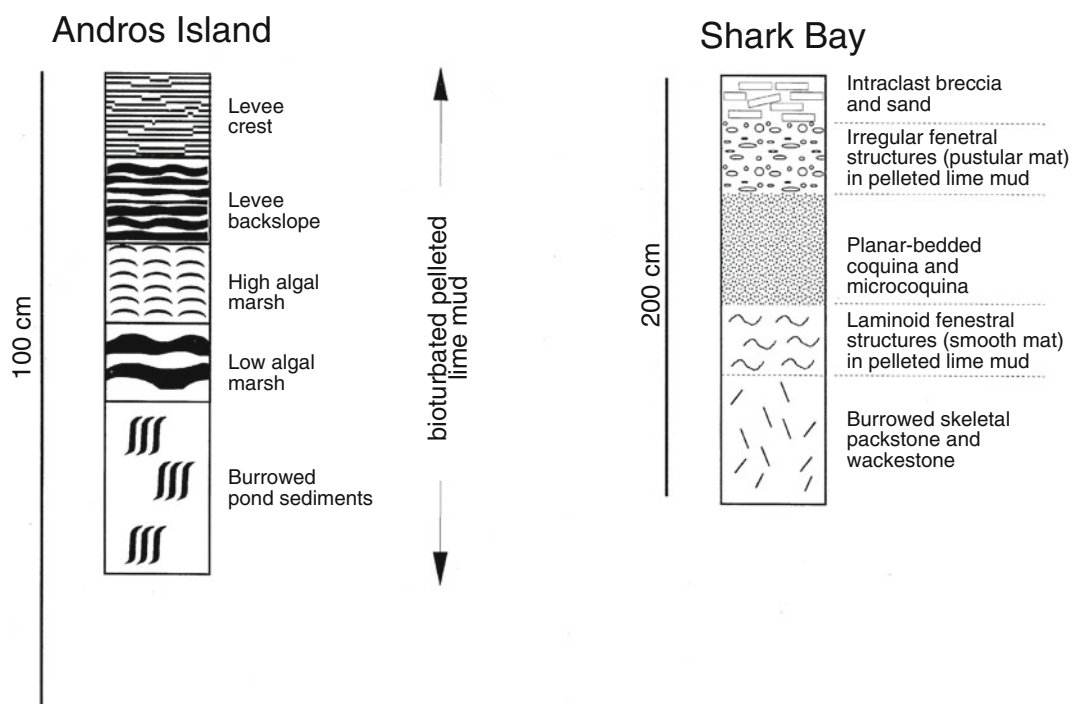
termed “biostromes.” Where organisms such as oysters and worm tube form rigid interlocking skeletal frames as sheets, they are biostromal reefs and, if emergent above the tidal flat surface, form biohermal reefs (or “bioherms”). Terms for biostromes, bioherms, banks, and reefs are defined and discussed in Nelson et al. (1962) and Kershaw (1994).

The well-known biota of tidal flats include mangroves, salt marsh, algal mats and stromatolites, polychaetes, molluscs, crustacea, resident fishes, and invading nektonic and demersal fishes and avifauna (Table 6). Biogeography and climate, substrate, hydrology, and hydrochemistry are major factors determining what biota inhabits tidal flats. Species abundance and zonation at site-specific level are determined by physico-chemical and biological conditions (Semeniuk 1983; Adam 1990; Silvestri et al. 2005). For macrophytes, at a global scale, mangroves dominate mid- to upper-tidal flats in tropical climates and are replaced by salt marsh in temperate climates.

EXTREMELY MACROTIDAL & MACROTIDAL SILICICLASTIC SEQUENCES



MICROTIDAL CARBONATE SEQUENCES

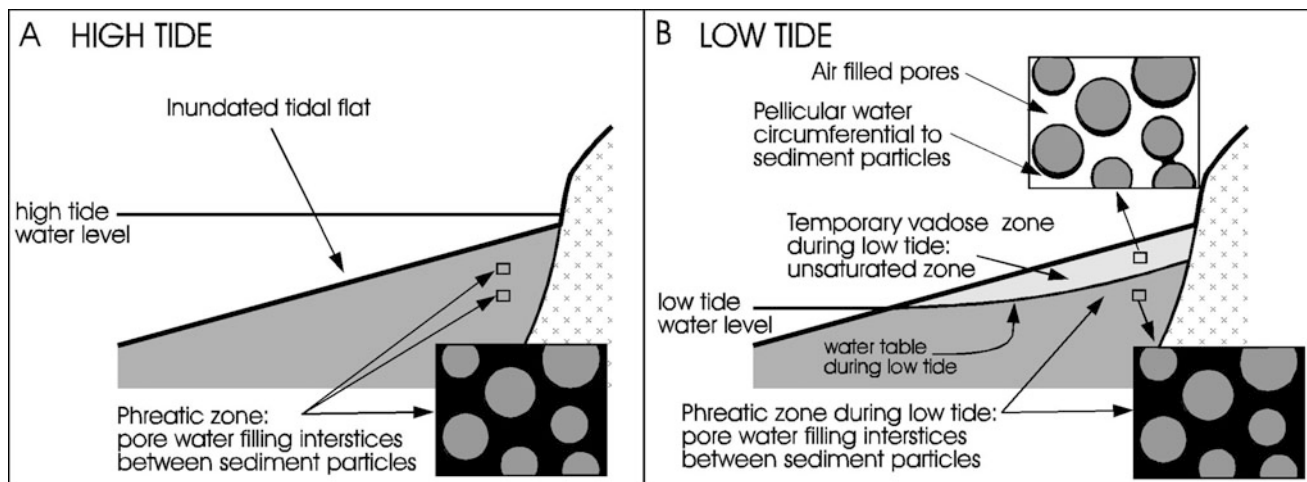


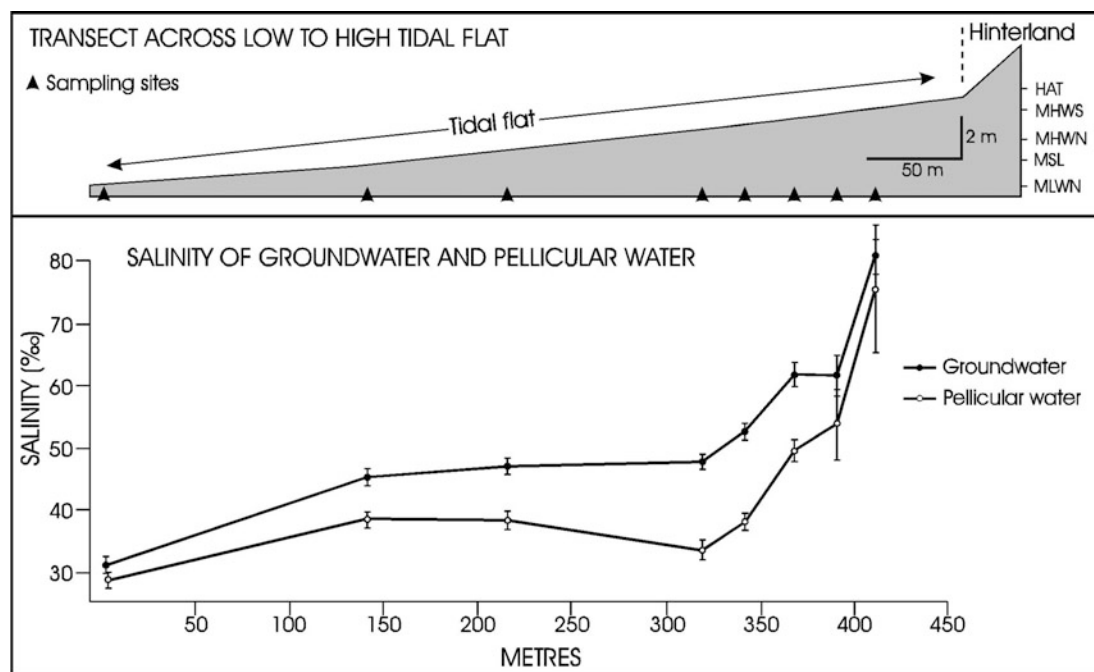
Tidal Flats, Fig. 8 Stratigraphic sequences from various tidal flats (see Table 1). For the macrotidal siliciclastic settings there is a comparison for three sequences: a tropical semiarid mangrove-vegetated tidal flat, shoaling from sand to mud, with burrows, root structures, and in situ mangrove stumps; a subtropical semiarid vegetation-free tidal flat

dominated by mud, with local burrows, and evaporitic mineral structures; and a temperate humid salt marsh vegetated tidal flat, shoaling from sand to mud, with burrows, root structures, and shell. The microtidal carbonate sequences compare the structures of sub-tropical humid tidal flat with that of a subtropical arid tidal flat

Tidal Flats, Table 5 Examples of sediments in their setting, and processes in their development

Environment	Main processes	Resulting sediment(s)
Siliciclastic sediment settings		
Mangrove or salt marsh vegetated high-tidal mudflat	Mud accumulation; root-structuring, bioturbation; shell contribution; groundwater alteration	Grey bioturbated root-structured (shelly) mud
Algal mat-covered high-tidal mudflat	Mud accumulation; binding; trapping; redox reactions; cracking	Laminated mud; desiccated laminated mud
Bare high-tidal mudflat	Mud accumulation; surface shear; cracking of mud; reworking of desiccation polygons; gypsum precipitation	Laminated mud; desiccated mud; mud-chip breccia; gypseous mud
Burrow-structured mid-tidal mudflat	Mud accumulation; surface shear; benthic fauna burrowing	Burrow-structured laminated mud; bioturbated mud
Burrow-pocked mid-tidal mudflat	Mud accumulation; surface shear; benthic fauna burrowing	Burrow-structured laminated mud; bioturbated mud
Mollusc-inhabited mid-tidal mudflat	Mud accumulation; surface shear; accumulation of shell winnowing to concentrate shells	Laminated shelly mud; shell gravel bed
Hummocky to “crated” low-tidal sand flat	Sand accumulation; surface shear; benthos inhabited; nekton and demersal fish feeding on benthos creating excavation craters	Thoroughly bioturbated sand to shelly sand
Mid-tidal mudflat with sand ripples	Mud accumulation; surface shear; traction transport of sand	Flaser bedding
Megarippled low- to mid-tidal sand flat	Traction transport of sand; air trapped by rise and fall of tide	Cross-laminated sand; bubble sand
Mid-tidal burrow-pocked sand flat	Traction transport of sand; benthic fauna burrowing	Burrow-structured cross-laminated sand; bioturbated sand
Carbonate sediment settings		
High-tidal breccia pavement	Mud accumulation; carbonate cementation; root-structuring; groundwater alteration	Limestone breccia sheet
High-tidal alga mat-covered mudflat	Mud accumulation; binding; trapping; redox reactions; cracking	Laminated mud; desiccated laminated lime mud
High-tidal bare mudflat	Mud accumulation; surface shear; cracking of mud; reworking of cracks; gypsum precipitation	Laminated lime mud; desiccated lime mud; mud-chip breccia; gypseous lime mud; laminated gypsum; gypsum nodule bed
Mid-tidal burrow-pocked mudflat	Mud accumulation; surface shear; benthic fauna burrowing	Burrow-structured laminated mud; bioturbated mud

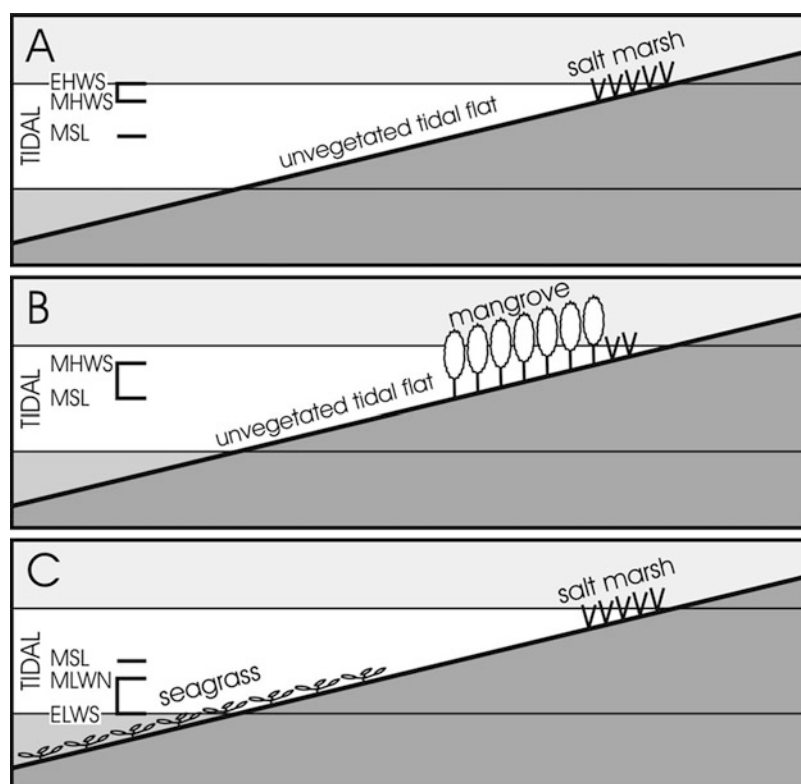
**Tidal Flats, Fig. 9** The characteristics of groundwater residing under tidal flats during high and low tide. (From Semeniuk 2015b)

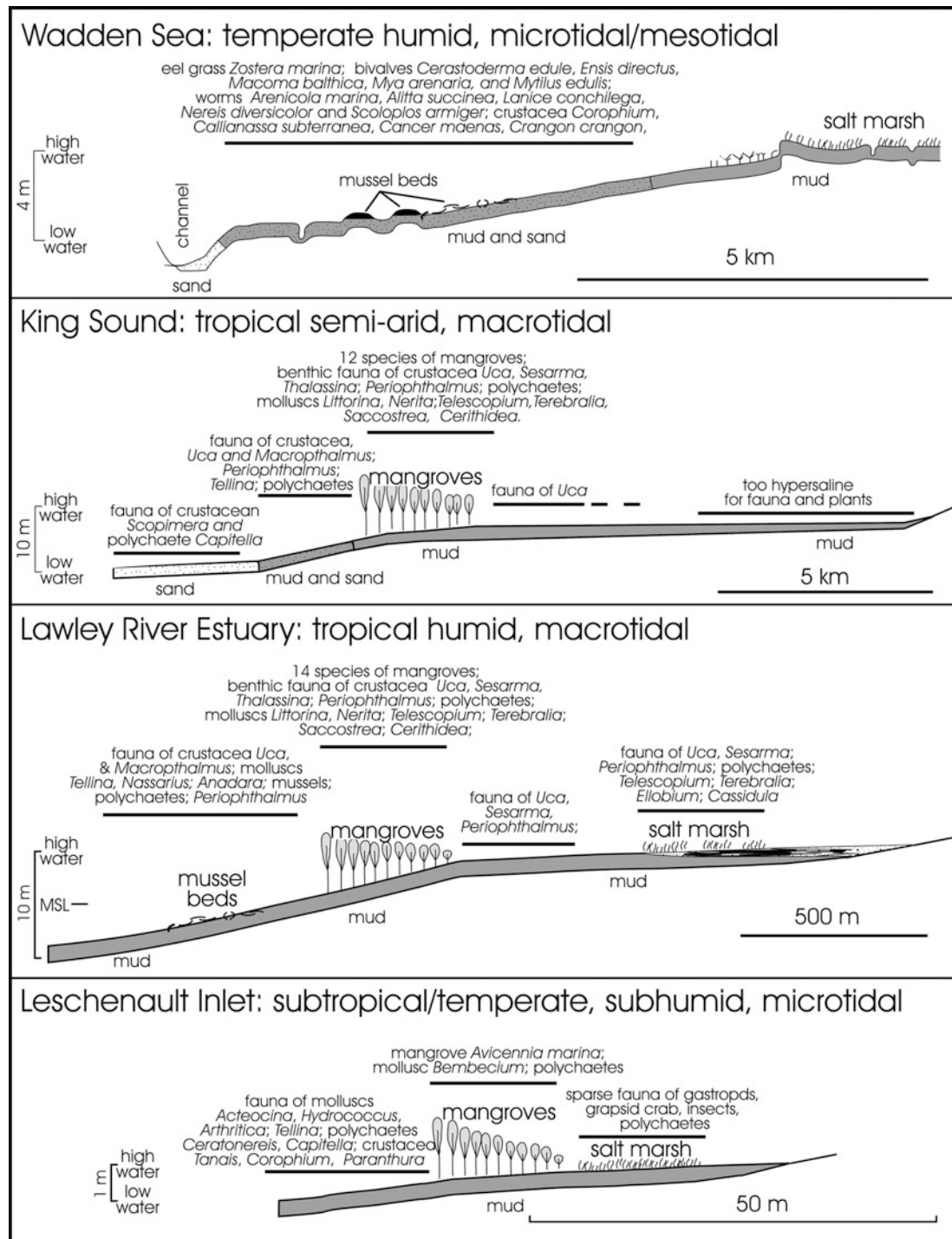


Tidal Flats, Fig. 10 Salinity of groundwater and pellicular water across a tropical tidal flat in north-western Australia. (From Semeniuk 1983, 2015b). The salinity of the groundwater and pellicular increases from the low-tidal zone to the high-tidal zone, with the groundwater being of a

slightly higher salinity than the pellicular water. There is no freshwater seepage in this location, and so the hypersalinity of the high-tidal flats is not diluted

Tidal Flats, Fig. 11 Idealized diagram of tidal flat surfaces showing a range of vegetation and plant life that may occupy specific tidal zones. The positions of the various tide levels for these profiles are shown in Fig. 3: (a) salt marsh on the high tidal flats usually between MHWS and EHWS; (b) mangroves on the high tidal flats usually between MSL and MHWS, bordered in this example by salt marsh on the landward side; (c) seagrass between MLWN and subtidal zone, and salt marsh between MHWS and EHWS. (From Semeniuk 2015a)

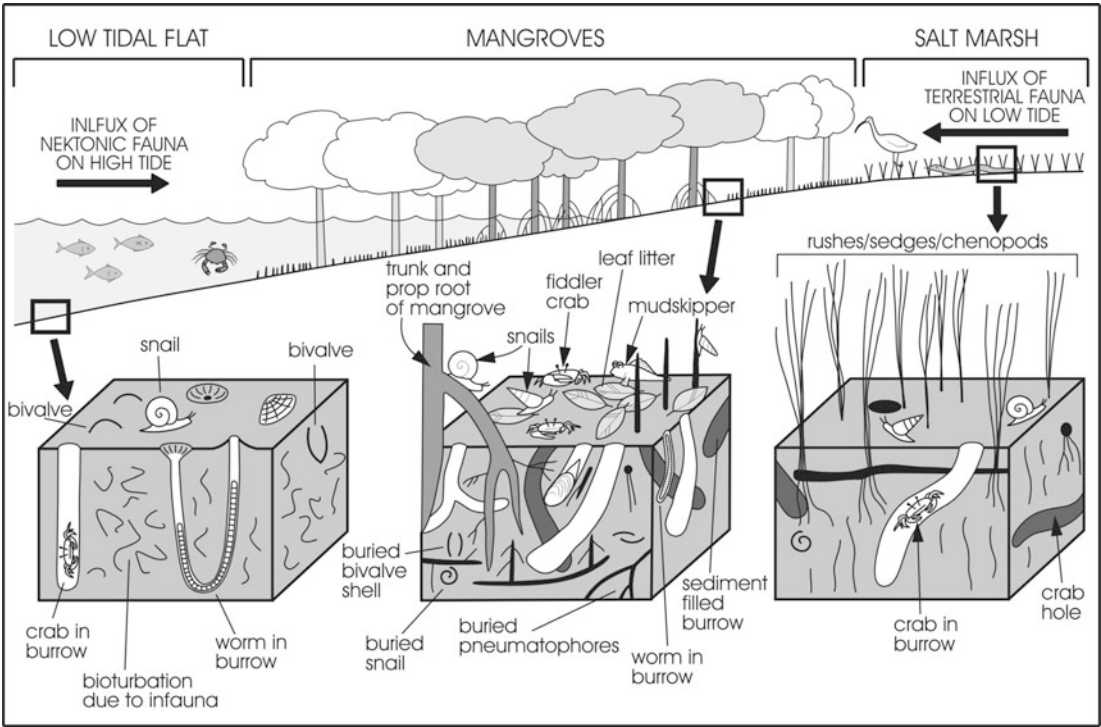




Tidal Flats, Fig. 12 Profiles of some typical tidal flats showing substrate types and generalized and simplified composition of biota (mangrove, salt marsh, invertebrates, mussels). (After Semeniuk 2015a)

With increased salinity, the upper tidal interval may be inhabited by algal mats and stromatolites. Diversity of flora and fauna is linked to climate setting, with high species richness and abundance in tropical areas and relatively lower species richness in temperate areas.

Primary production within specific parts of the tidal flat, for example, from mangroves and salt marsh, often drives the ecosystems of tidal flats. With mangroves and salt marshes, these macrophytes fix nutrients and carbon on the mid- to upper tidal flats, supporting the local resident fauna and the



Tidal Flats, Fig. 13 Diagram showing the large scale ecological relationships of primary production, ecological functioning at a site, and the nekton to terrestrial utilisation/feeding invasions and feeding and also showing blocks of sediment with different types of biota commensurate with substrate type and tidal level setting and the effects of biota on the substrates. (Modified from Semeniuk et al. 1978; Semeniuk 2015a)

Tidal Flats, Table 6 Key biota of the tidal flats

Biota	Occurrence and function
Mangroves	Tropical climate mid- to high-tidal flats; massive primary production in the mid-upper tidal zone; plant detritus sustains biota in the immediate and in the adjoining tidal zones (Tomlinson 1986; Hutchins and Saenger 1987)
Salt marsh – (comprising chenopod vegetation and/or <i>Spartina</i>)	Temperate to tropical climate tidal flats; primary production in the mid-upper tidal zone; plant detritus sustains biota in the immediate and in the adjoining tidal zones (Chapman 1977; Beeftink et al. 1985; Adam 1990; Pennings et al. 2005; Silvestri et al. 2005)
Algal mats and stromatolites	Tropical climate tidal flats; primary production in the mid-upper tidal zone (Kendall and Skipwith 1968; Ginsburg and Hardie 1975); algal mats provide grazing grounds for benthos and fish.
Molluscs, polychaetes, crustacea	These invertebrates occur generally on all tidal flats, although species diversity may decrease towards the temperate climate regions; molluscs, polychaetes, and crustacea are primary and secondary consumers, and sustain higher-level trophic feeders (Dankers et al. 1983; Knox 1986; Semeniuk et al. 2000).
Fish and avifauna	On tidal flats; fish and avifauna generally are primary and secondary consumers, and sustain higher-level trophic feeders; in many instances, they are the highest trophic level in the region (Knox 1986; de Sylva 1975; Owen and Black 1990)

export of detritus sustains benthic biota of polychaetes, molluscs, and crustacea elsewhere on the mid- to low-tidal flats. The biologically rich tidal flat environments also support nekton, demersal fish, and avifauna. Nektonic and demersal fish and other nekton invade the tidal zone for feeding on the high tide, and the avifauna invade the tidal flats at low tide (Fig. 13).

Tidal flats typically are biologically zoned (Figs. 11 and 12). For any benthic group such as polychaetes, molluscs, or

crustacea, there is a species zonation across the flats related to frequency of inundation, substrate type, substrate wetness, pore-water salinity, inter-species competition, and predation pressure, amongst other factors (Semeniuk and Cresswell 2015). Macrophytes (mangrove and salt marsh) also exhibit zonation as related to groundwater salinity, substrates, and elevation of habitat above mean sea level. Many of the benthos are burrowing forms, and the macrophytes have diagnostic root structures, and hence sedimentologically, zonation

Tidal Flats, Table 7 Key processes on tidal flats

Some selected processes	Examples of products on the tidal flat
Oceanographic	
Flood and ebb-tidal currents and slack water	Deposition of mud from suspension, and sand transport by traction; silt ripples, sand ripples, sand waves and megaripples; laminated mud, lenticular bedding, and flaser bedding; scour, small cliffs, and cut-and-fill; tidal creek formation; erosion of mud beds along creek banks and cliff-lines to form mud-ball conglomerate; rolling of mud balls on sandy/gravelly floors to form armoured mud balls
Waves	Winnowed sand sheets and shell gravel; rippling; erosion of cliff-lines cut into mud
Meteorologic/atmospheric	
Wind (erosion, transport, deposition, evaporation)	Sand and silt transport and fall-out deposition onto tidal flats; formation of adhesion ripples; evaporation of wet surfaces at low tide
Solar-induced and/or wind induced evaporation	Mud desiccation and cracking; increasing pore water salinity of groundwater; increasing salinity of groundwater leading to precipitation of minerals
Rain, ice crystallization	Rain imprints; ice crystal imprints; cryogenic disruption
Water temperature variation	Mud deposition from suspension; mortality of benthos
Groundwater hydrologic/ hydrochemical	
Rising and falling of tide	Wetting then drying to form desiccated sediment; erosion of desiccated sediment to form mud flakes; development of bubble sand
Solution/precipitation of carbonate minerals	Shell voids and other vughs; cemented crusts, teepees
Evaporite mineral precipitation	Beds of nodular to platy crystalline gypsum; precipitation of gypsum disrupting primary structures
Iron mineral precipitation	Staining of sediment to dark grey with iron sulfide; staining of sediments to orange-brown with iron oxides
Biological	
Accumulation of shell beds and biostromes, and biostromal to biohermal reefs	Shelly sediments and coquinas, and skeletal biostromes
Plant detritus accumulation	Organic-rich sediments, peat
Sediment trapping and binding by vegetation and algal mats	Root-structured and plant-structured sediment; algal-laminated sediment
Feeding/foraging by nekton and demersal fish	Pocked, pot-holed, excavated, and hummocky surfaces
Burrowing by benthos	Burrow-structured to fully-bioturbated sediment

of the biota results in facies and tidal-level-specific sedimentary signatures across tidal flats; sand-constructed *Arenicola* burrows, for instance, are diagnostic of low-tidal sand flats, vertical to u-shaped to labyrinthoid crustacean burrows in a root-structure-free mud are diagnostic of mid- to low-tidal flats, coarse root-structured substrates and associated faunal burrows are diagnostic of mangrove-vegetated high-tidal flats, while fine root-structured substrates are diagnostic of salt marsh-vegetated high-tidal flats. Some diagnostic biogenic structures and biofacies related to tidal assemblages are often signatures of specific tidal levels and lithofacies within a given region.

Summary

The coastal zone is one of the most complex environments on Earth, located at the triple junction between land, sea, and atmosphere (Brocx and Semeniuk 2009). In this context, as low-gradient shores, tidal flats exhibit a myriad of products resulting from interactive, interrelated and overlapping exogenous and endogenous agents and processes, which

include oceanographic, meteorologic, atmospheric, fluvial, hydrologic and hydrochemical, and biological processes (Table 7). As outlined above, these processes commonly are distributed along physicochemical gradients (e.g., gradients of tidal effects, hydrology, hydrochemistry, geochemistry) and operate on a range of basic sediment types such as mud, sand, and shell gravel to develop a complex of geomorphic, sedimentologic, and diagenetic products which are commonly zoned across the tidal flats and are often specific to a coastal setting, sediment setting, climate, and biogeography.

Cross-References

- [Bioherms and Biostromes](#)
- [Coastal Sedimentary Facies](#)
- [Coastal Wind Effects](#)
- [Mangroves, Ecology](#)
- [Microtidal Coasts](#)
- [Muddy Coasts](#)
- [Salt Marsh](#)
- [Spits](#)

- Tidal Creeks
- Tidal Datums
- Tidal Environments
- Tide-Dominated Coasts
- Vegetated Coasts

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Tidal Inlets

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Introduction

Tidal inlets are an integral part of coastal barrier systems throughout the world. They serve as natural conduits for the exchange of water, nutrients, and organisms between coastal ocean and the backbarrier. In addition, along many coastal plain settings, including much of the East and Gulf Coasts of the United States, the southern and western coasts of the North Sea, and along many deltaic coasts, the only safe harborages, including major ports, are found behind barrier islands. The large number of improvements that are performed at the entrance to inlets such as the construction of jetties and breakwaters, dredging of channels, and the operation of sand bypassing facilities demonstrate the importance of inlets in providing navigation routes to these harborages.

Diversity in the morphology, hydraulic signature, and sediment transport patterns of tidal inlets attest to the dynamics and complexity of their processes. The variability in oceanographic, meteorologic, and geologic parameters, such as tidal range, wave energy, sediment supply, storm magnitude and frequency, freshwater influx, geologic controls, and the

interactions of these factors, are responsible for a wide range in tidal inlet morphologies and channel geometries and make them highly dynamic geomorphic systems.

Tidal Inlet Definition

A tidal inlet is defined as an opening in the barrier shoreline through which water penetrates the land thereby proving a connection between the coastal ocean and bays, lagoons, and marsh and tidal creek systems. Tidal currents maintain the main channel of a tidal inlet and keep it flushed of sediment.

The second half of this definition distinguishes tidal inlets from large, open embayments or passageways along rocky coasts. Tidal currents at inlets are responsible for the continual removal of sediment dumped into the main channel by wave action. Thus, according to this definition tidal inlets occur along sandy or mixed sand/gravel barrier coastlines, although one side may abut a bedrock headland. Some tidal inlets coincide with the mouths of rivers (estuaries) but in these cases, inlet dimensions and sediment transport trends are still governed, to a large extent, by the volume of water exchanged at the inlet mouth and the reversing tidal currents, respectively.

At most inlets over the long term, the volume of water entering the inlet during the flooding tide equals that leaving during the ebbing phase. This volume is referred to as the tidal prism. It is a function of the open water area and tidal range in the backbarrier, as well frictional factors that govern the enhancement or attenuation of tidal wave propagation into the backbarrier.

Inlet Morphology

A tidal inlet is specifically the area between two barriers or between the barrier and the adjacent bedrock or glacial headland. Commonly, recurved ridges of spits, consisting of sand that was transported toward the backbarrier by refracted waves and flood-tidal currents, form the sides of the inlet. The deepest part of an inlet, the inlet throat, is usually located where accretion by one or both of the bordering barrier spits constricts the inlet channel to a minimum width and minimum equilibrium cross-sectional area. Here, tidal currents normally reach their maximum velocity. Commonly, the strength of the currents at the throat causes sand to be removed from the channel floor leaving behind a lag deposit composed of gravel or shells and, in some locations, channel scour may terminate in shallow bedrock or indurated sediment. On the continental shelf, the throat sections are often the only vestiges of former inlet channels and thus aid in reconstructing ancient barrier positions in areas where their superstructure is otherwise eroded during sea-level rise.

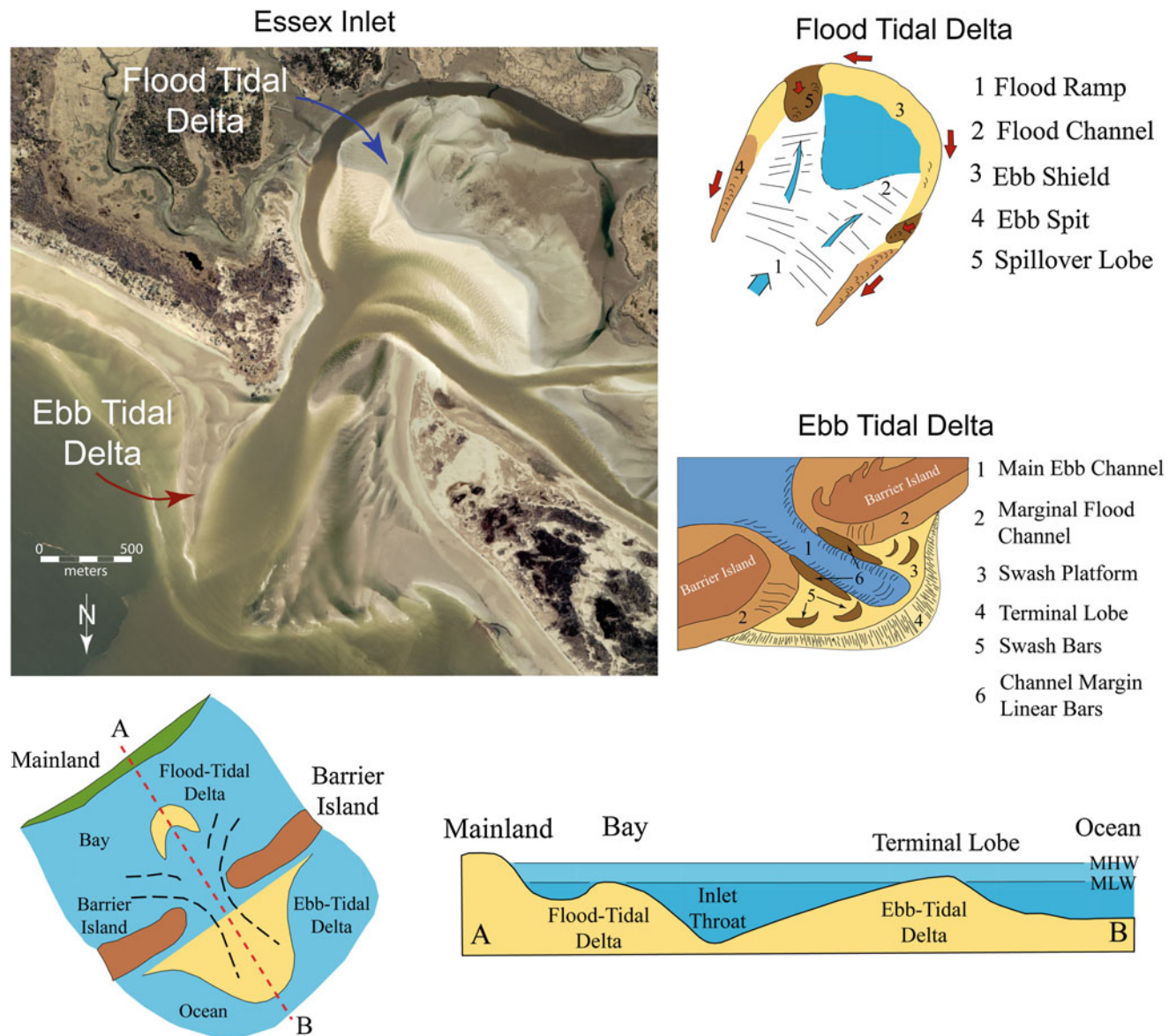
Tidal Deltas

Closely associated with tidal inlets are sand shoals and tidal channels located on the landward and seaward sides of the inlets. Flood tidal currents carry sand landward and deposit it in the backbarrier forming flood-tidal deltas; ebb-tidal currents deposit sand on the seaward side of the inlet forming an ebb-tidal delta.

Flood-Tidal Delta. Their presence or absence, size, and development are related to a regional tidal range, wave energy, sediment supply, and backbarrier setting. Tidal inlets that are backed by a system of tidal channels and salt marsh (mixed-energy coast) usually contain a single horseshoe-shaped flood-tidal delta (e.g., Armona Inlet, Portugal; Rangaunu Harbor Inlet, New Zealand; Essex River Inlet, Massachusetts; Fig. 1). Contrastingly, inlets that are backed by large shallow bays may contain multiple flood-tidal deltas. Along some microtidal coasts, such as Rhode Island, flood deltas form at the end of narrow inlet channels cut through the barrier. Changes in the locus of deposition at these deltas produce a multi-lobate morphology resembling a lobate river delta (Boothroyd et al. 1985). Flood delta size commonly increases as the amount of open water area in the backbarrier expands. In some regions, especially following inlet closure, flood deltas become colonized and altered by marsh growth that masks their original morphology. At other sites, portions of flood-tidal deltas are dredged to provide navigable waterways and thus are highly modified.

Flood-tidal deltas are best revealed in areas with moderate to large tidal ranges (1.5–3.0 m) because in these regions they are well exposed at low tide (Fig. 2). As tidal range decreases, flood deltas get converted into subtidal shoals. Most flood-tidal deltas have similar morphologies consisting of the following components (Hayes 1975, 1979).

1. **Flood ramp:** This is a landward shallowing channel that slopes upward toward the intertidal portion of the delta. Strong flood-tidal currents and storm waves cause landward sand transport in the form of landward-oriented sandwaves.
2. **Flood channels:** The flood ramp splits into two shallow flood channels. Like the flood ramp, these channels are dominated by flood-tidal currents and flood-oriented sand waves. Sand is delivered through these channels onto the flood delta.
3. **Ebb shield:** It defines the highest and landwardmost part of the flood delta and may be partly covered by marsh vegetation. It shields the rest of the delta from the effects of the ebb-tidal currents, deflecting the outflow laterally toward the ebb spits.
4. **Ebb spits:** These spits extend from the ebb shield toward the inlet. They form from sand that is eroded from the ebb shield and transported back toward the inlet by ebb-tidal currents.



Tidal Inlets, Fig. 1 Tidal inlet morphology and components as illustrated in a vertical aerial photograph of Essex River Inlet, Massachusetts containing well-developed flood- and ebb-tidal deltas. Tidal delta models are from Hayes (1975)

5. **Spillover lobes:** These are lobes of sand that form where the ebb currents have breached through the ebb spits or ebb shield depositing sand in the interior of the delta.

Through time, some flood-tidal deltas accrete vertically and/or grow in size. This is evidenced by an increase in areal extent of marsh grasses, which require a certain elevation above mean low water to exist. At migrating inlets, new flood-tidal deltas are formed as the inlet moves along the coast and encounters new open water areas in the backbarrier. At most stable inlets, however, sand comprising the flood delta is largely recirculated. The transport of sand on flood deltas is controlled by the elevation of the tide and the strength and direction of the tidal currents. During the rising tide, flood

currents reach their strongest velocities near high tide when the entire flood-tidal delta is covered by water. Hence, there is a net transport of sand up the flood ramp, through the flood channels and onto the ebb shield. Some of the sand is moved across the ebb shield and into the surrounding tidal channel. During the falling tide, the strongest ebb currents occur near mid to low water. At this time, the ebb shield is emerged and diverts the currents around the delta. The ebb currents erode sand from the landward face of the ebb shield and transport it along the ebb spits and eventually into the inlet channel where once again it will be moved onto the flood ramp, thus completing the sand gyre.

Ebb-Tidal Delta. This is an accumulation of sand that has been deposited by the ebb-tidal currents and that has been



Tidal Inlets, Fig. 2 Well-developed flood-tidal delta at Barnstable Harbor Inlet, Cape Cod, Massachusetts. This region has a tidal range of 3 m

subsequently modified by waves and tidal currents. Ebb deltas exhibit a variety of forms dependent on the relative magnitude of wave and tidal energy of the region, as well as geological controls. Best developed along mixed energy coasts, most ebb-tidal deltas contain the same general features including (Fig. 1):

1. **Main ebb channel:** This is a seaward shallowing channel that is scoured into the ebb-tidal delta sands. It is dominated by ebb-tidal currents, although a mutually evasive landward flow may exist within the channel during part of the tidal cycle.
2. **Terminal lobe:** Sediment transported out the main ebb channel is deposited in an arcuate shoal of sand forming the terminal lobe. The deposit slopes relatively steeply on its seaward side and its outline is well defined by breaking waves during storms or periods of large swell at low tide.
3. **Swash platform:** This is a broad shallow sand platform located on both sides of the main ebb channel, defining the general extent of the ebb delta.
4. **Channel margin linear bars:** These are bars that border the main ebb channel and sit atop the swash platform. They tend to confine the ebb flow and are exposed at low tide.
5. **Swash bars:** Waves breaking over the terminal lobe and across the swash platform form arcuate swash bars that migrate onshore. The bars are usually 50–150 m long, 50 m wide, and 1–2 m in height.
6. **Marginal-flood channels:** These are shallow channels (0–2 m deep at mean low water) located between the channel margin linear bars and the ocean beaches. The channels are dominated by flood-tidal currents that may encounter ebb flow in the main ebb channel during parts of the tidal cycle.

Ebb-Tidal Delta Morphology

The general shape of an ebb-tidal delta and the distribution of its sand bodies tell us about the relative magnitude of different sand transport processes operating at a tidal inlet. Ebb-tidal deltas that are elongated in a shore-normal direction, with a main ebb channel and channel margin linear bars that extending far offshore, occur at tide-dominated inlets (e.g., Georgia and southern South Carolina). Wave-generated sand transport plays a secondary role in modifying delta shape at these inlets. Because most sand movement is in onshore-offshore direction, the ebb-tidal overlaps a relatively small length of inlet shoreline. This has important implications concerning the extent to which the inlet shoreline undergoes erosional and depositional changes.

Wave-dominated inlets tend to be small relative to tide-dominated inlets. Their ebb-tidal deltas are driven onshore, close to the inlet mouth by the dominant wave processes (e.g., North Carolina, Maryland, Texas). Commonly, the terminal lobe and/or swash bars form a small arc outlining the periphery of the delta. In many cases, the depositional elements of the delta are entirely subtidal. In other instances, sand bodies clog the entrance to the inlet leading to the formation of several major and minor tidal channels.

At mixed energy tidal inlets, the shape of the delta is the result of a relative balance between tidal and wave processes (e.g., northern South Carolina). These deltas have a well-formed main ebb channel, which is a product of ebb-tidal currents. Their swash platform and sand bodies substantially overlap the inlet shoreline many times the width of the inlet throat due to wave processes and flood-tidal currents.

Ebb-tidal deltas may also be highly asymmetric such that the main ebb channel and its associated sand bodies are positioned predominantly along one of the inlet shorelines. This configuration normally occurs when the major back-barrier channel approaches the inlet at an oblique angle or when a preferential accumulation of sand on the updrift side of the ebb delta causes a deflection of the main ebb channel along the downdrift barrier shoreline.

Tidal Inlet Formation

The formation of a tidal inlet requires the presence of an embayment and the development of barriers. In coastal plain settings, the embayment or backbarrier was often created through the construction of the barriers themselves, like much of the East Coast of the United States or the Friesian Island coast along the North Sea. In other instances, the embayment was formed due to rising sea level inundating an irregular shoreline. The embayed or indented shoreline may have been a rocky coast such as that of northern New England and California or it may have been an irregular unconsolidated sediment coast such as that of Cape Cod in

Massachusetts or parts of the Oregon coast. The flooding of former river valleys has also produced embayments associated with tidal inlet development.

Breaching of a Barrier

Rising sea level, exhausted sediment supplies, and human influences have led to thin barriers that are vulnerable to breaching. The breaching process normally occurs during storms after waves have destroyed the foredune ridge and storm waves have overwashed the barrier depositing sand aprons (washovers) along the backside of the barrier. Even though this process may produce a shallow overwash channel, seldom are barriers cut from their seaward side. In most instances, the breaching of a barrier is the result of the storm surge heightening waters in the backbarrier bay. When the level of the ocean tide falls, the elevated bay waters flow toward the ocean as an ebb surge, gradually incising the barrier and cutting a channel. If subsequent tidal exchange between the ocean and bay is able to maintain the channel, a tidal inlet is established. The breaching process is enhanced when offshore winds accompany the falling tide and if an overwash channel is present to facilitate drainage across the barrier (Fisher 1962). Many tidal inlets that are formed by this process are ephemeral and may exist for less than a year, especially if stable inlets are located nearby. Barriers most susceptible to breaching are long, thin, low, and wave-dominated.

Spit Extension Across a Bay

The development of a tidal inlet by spit construction across an embayment usually occurs early in the evolution of a coast. The sediment to form these spits may have come from erosion of the nearby headlands, discharge from rivers, or from the landward movement of sand from inner shelf deposits. Most barriers along the coast of the United States and elsewhere in the world are 3,000–5,000 years old, coinciding with a deceleration of rising sea level. It was then that spits began growing and enclosing portions of the irregular rocky coast of New England, the West Coast, parts of Australia, and many other regions of the world. As a spit builds across a bay, the opening to the bay gradually decreases in width and cross-sectional area. It may also deepen. Coincident with the decrease in size of the opening is a corresponding increase in tidal flow. The tidal prism of the bay remains approximately constant, so as the opening gets smaller, the current velocities must accelerate. A tidal inlet is formed as the spit reaches a stable configuration.

Drowned River Valleys

Tidal inlets have formed at the entrance to drowned river valleys due to the growth of spits and the development of barrier islands which have served to narrow the mouths of the estuaries. It has been shown through stratigraphic studies,

particularly along the East Coast of the United States, that in addition to drowned river valleys, many tidal inlets are positioned in paleo-river valleys that are not occupied by fluvial systems at present (Halsey 1979). These are old river courses that were active during the Pleistocene when sea level was lower and they were migrating across the exposed continental shelf. Tidal inlets become established in these valleys because tidal currents easily remove the sediment filling the valleys.

Tidal Inlet Migration

Some tidal inlets have been stable since their formation whereas others have migrated long distances along the shore. In New England and along other glaciated coasts, stable inlets are commonly anchored next to bedrock outcrops or resistant glacial deposits. Along the California coast, most tidal inlets have formed by spit construction across an embayment with the inlet becoming stabilized adjacent to a bedrock headland. In coastal plain settings, stable inlets are commonly positioned in former river valleys. One factor that appears to differentiate migrating inlets from stable is the depth to which the inlet throat has eroded. Deeper inlets are often situated in consolidated sediments that resistant erosion. The channels of shallow migrating inlets are eroded into unconsolidated sand.

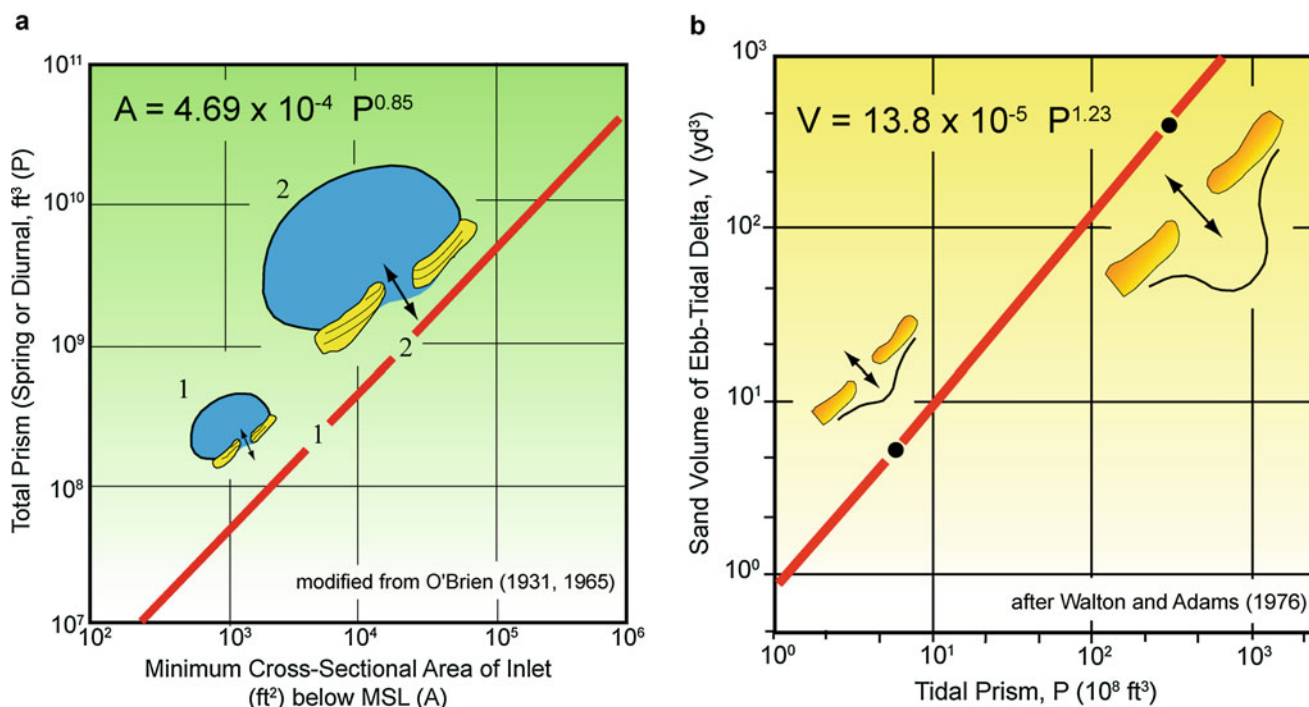
Although the vast majority of tidal inlets migrate in the direction of dominant longshore transport, some migrate updrift (Aubrey and Speer 1984). In these cases, the drainage of backbarrier tidal creeks controls flow through the inlet. When a major backbarrier tidal channel approaches the inlet at an oblique angle, its ebb-tidal flow is directed toward the margin of the inlet throat. If this is the updrift side of the main channel, then the inlet will migrate in that direction. This is similar to a river where strong currents are focused along the outside of a meander bend causing erosion and channel migration. Inlets that migrate updrift are usually small to moderately sized and occur along coasts with small to moderate net longshore sediment transport rates.

Tidal Inlet Relationships

Tidal inlets throughout the world exhibit several consistent relationships that have allowed coastal engineers and marine geologists to formulate predictive models: (1) inlet throat cross-sectional area is closely related to tidal prism, and (2) ebb-tidal delta volume is a function of tidal prism.

Inlet Throat Area: Tidal Prism Relationship

The size of a tidal inlet is tied closely to the volume of water going through it (Fig. 3a; O'Brien 1931, 1969). Although inlet size is primarily a function of tidal prism, to a lesser



Tidal Inlets, Fig. 3 (a) Graph depicting relationship between tidal prism and inlet throat cross-sectional area. (From O'Brien 1969). (b) Graph showing correspondence between tidal prism and volume of the ebb-tidal delta. (From Walton and Adams 1976)

degree inlet cross-sectional area is also affected by the delivery of sand to the inlet channel. For example, at jettied inlets, tidal currents can more effectively scour sand from the inlet channel and therefore they maintain a larger throat cross-section than would be predicted by the O'Brien Relationship. Similarly, for a given tidal prism, Gulf Coast inlets have larger throat cross-sections than Pacific Coast inlets. This is explained by the fact that wave energy is greater along the West Coast and therefore the delivery of sand to these inlets is higher than at Gulf Coast inlets. Jarrett (1976) has improved the tidal prism – inlet cross-sectional area regression equation for U.S. inlets by separating into three classes the low energy Gulf Coast inlets, moderate energy East Coast inlets, and higher energy West Coast inlets. Even better correlations are achieved when structured channels are distinguished from natural inlets.

Variability. It is important to understand that the dimensions of the inlet channel are not static but rather the inlet channel enlarges and contracts slightly over relatively short time periods (<1 year) in response to changes in tidal prism, variations in wave energy, effects of storms, and other factors. For instance, the inlet tidal prism can vary by more than 30% from neap to spring tides due to increasing tidal ranges. Consequently, the size of the inlet varies as a function of tidal phases. Along the southern Atlantic Coast of the United States, water temperatures may fluctuate seasonally by 35–40 °F. This causes the surface coastal waters to expand, raising mean sea level by 30 cm or more. In the summer and

early fall when mean sea level reaches its highest seasonal elevation, spring tides may flood backbarrier surfaces that normally are above tidal inundation. This produces larger tidal prisms, stronger tidal currents, increased channel scour, and enlarged inlet cross-sectional areas. At some Virginia inlets, this condition increases the inlet throat by 5–15% (Byrne et al. 1975). Longer-term (>1 year) changes in the cross-section of inlets are related to inlet migration, sedimentation in the backbarrier, morphological changes of the ebb-tidal delta, and human influences.

Ebb-Tidal Delta Volume: Tidal Prism Relationship

In the mid-1970s, Walton and Adams (1976) showed that the volume of sand contained in the ebb-tidal delta is closely related to the tidal prism (Fig. 3b). They also showed that the relationship was improved slightly when wave energy was taken into account in a manner similar to Jarrett's divisions. Waves are responsible for transporting sand back onshore thereby reducing the volume of the ebb-tidal delta. Therefore, for a given tidal prism, ebb-tidal deltas along the West Coast contain less sand than do equally sized inlets along the Gulf or East Coast.

Variability. The Walton and Adams Relationship works reasonably well for inlets all over the world. However, field studies have shown that the volume of sand comprising ebb-tidal deltas changes through time due to the effects of storms, changes in tidal prism, or processes of inlet sediment bypassing (FitzGerald 1982). The process of moving sand

past a tidal inlet is commonly achieved by large bar complexes migrating from the ebb delta and attaching to the landward inlet shoreline. These large bars may contain more than 300,000 m³ of sand and represent more than 10% of sediment volume of the ebb-tidal delta (FitzGerald 1988; Gaudiano and Kana 2001).

Sand Transport Patterns

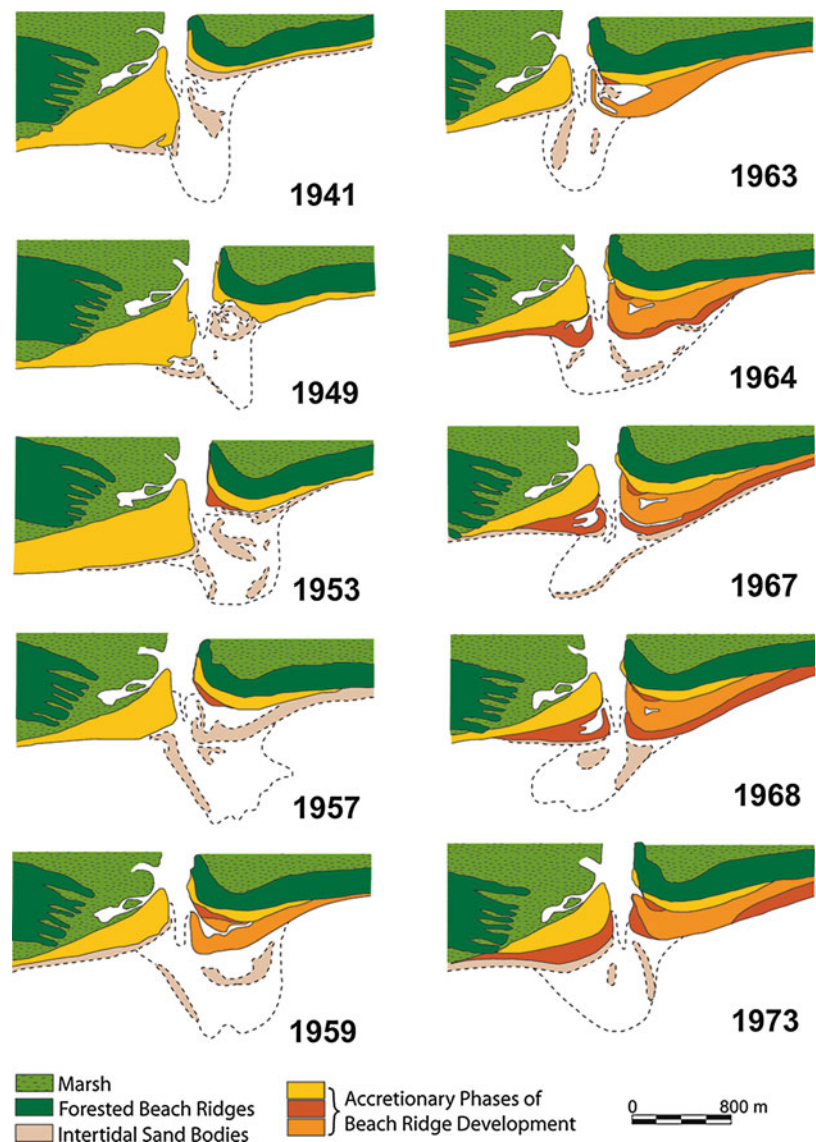
The movement of sand at a tidal inlet is complex due to reversing tidal currents, effects of storms, and interaction with the longshore transport system. The inlet contains short-term and long-term reservoirs of sand varying from the relatively small sandwaves flooring the inlet channel that migrate meters each tidal cycle to the large flood-tidal delta shoals where some sand is recirculated, although the entire

deposit may remain stable for hundreds or even thousands of years. Sand dispersal at tidal inlets is complicated because, in addition to the onshore-offshore sediment transport by tidal and wave-generated currents, there is constant delivery to and transport away from the inlet produced by the longshore transport system. In the discussion below, the patterns of sand movement at inlets are described including how sand is moved past a tidal inlet. The result of these inlet-related sand transport patterns can cause substantial changes in inlet shoreline erosional-depositional trends (Fig. 4).

General Sand Dispersal Trends

The ebb-tidal delta contains segregated areas of landward versus seaward sediment transport that are controlled primarily by the way water enters and discharges from the inlet, as well as the effects of wave-generated currents. During the ebbing cycle the tidal flow leaving the backbarrier is

Tidal Inlets, Fig. 4 Tidal inlet shoreline changes at Price Inlet, South Carolina. (From FitzGerald 1988)



constricted at the inlet throat causing the ebb currents to accelerate. Once out of the confines of the inlet, the ebb flow expands laterally and the velocity slows. Sediment in the main ebb channel is transported in a net seaward direction and eventually deposited on the terminal lobe due to this reduction in current velocity. One response to this seaward movement of sand is the formation of ebb-oriented sandwaves having heights of 1–2 m.

In the beginning of the flood cycle, the ocean tide rises, while water in the main ebb channel continues to flow seaward as a result of momentum. Due to this phenomenon, water initially enters the inlet through the marginal flood channels, which act as the pathways of least resistance. The flood channels are dominated by landward sediment transport and are floored by flood-oriented bedforms. On both sides of the main ebb channel, the swash platform is most affected by landward flow produced by the flood-tidal currents and breaking waves. As waves shoal and break, they generate landward flow, which augments the flood-tidal currents but retards the ebb-tidal currents. The interaction of these forces acts to transport sediment in a net landward direction across the swash platform. In summary, at many inlets there is a general trend of seaward sand transport in the main ebb channel, which is countered by landward sand transport in the marginal flood channels and across the swash platform.

Inlet Sediment Bypassing

Along most open coasts, particularly in coastal plain settings, angular wave approach causes a net movement of sediment, which along the East Coast of the United States varies from 100,000–200,000 m³/year. The manner whereby sand moves past tidal inlets and is transferred to the downdrift shoreline is called inlet sediment bypassing. The primary mechanisms of sand bypassing natural inlets include: (1) stable inlet processes; (2) ebb-tidal delta breaching, and (3) inlet migration and spit breaching. One of the end products of all the different mechanisms is the landward migration and attachment of large bar complexes to the inlet shoreline. Discussion of this topic can be found in FitzGerald (1982) and FitzGerald et al. (2001).

Stable Inlet Processes. This mechanism of sediment bypassing occurs at inlets that do not migrate and whose main ebb channels remain approximately in the same position. Sand enters the inlet by: (1) wave action along the beach, (2) flood-tidal and wave-generated currents through the marginal flood channel, and (3) waves breaking across the channel margin linear bars. Most of the sand that is dumped into the main channel is transported seaward by the dominant ebb-tidal currents and deposited on the terminal lobe.

At lower tidal elevations, waves breaking on the terminal lobe transport sand along the periphery of the delta toward the landward beaches in much the same way as the sand moved in the surf and breaker zones along beaches. At higher

tidal elevations, waves breaking over the terminal lobe create swash bars on both sides of the main ebb channel. The swash bars (50–150 m long, up to 50 m wide) migrate onshore due to the dominance of landward flow across the swash platform. Eventually, they attach to channel margin linear bars forming large bar complexes. The latter tend to parallel the beach and may be more than a kilometer in length. They are fronted by a steep (25–33°) slipface that may reach up to 3 m in height.

The stacking and coalescence of swash bars to form a bar complex is the result of their slowing onshore migration up the nearshore ramp. As the bars gain a progressively greater intertidal exposure, the wave bores driving their onshore movement act over an increasingly shorter period of the tidal cycle. Eventually, the entire bar complex migrates onshore and welds to the upper beach, followed by one or more similar landforms. When a bar complex attaches to the downdrift inlet shoreline, some of this newly accreted sand is then gradually transported by wave action to the downdrift beaches, thus completing the inlet sediment bypassing process. It should be noted that some sand bypasses the inlet independent of the bar complex. In addition, some of the sand comprising the bar re-enters the inlet via the marginal flood channel and along the inlet shoreline.

Ebb-Tidal Delta Breaching. This process of sediment bypassing occurs at inlets with a stable throat position, where the main ebb channels migrates downdrift like the wag of a dog's tail. Sand enters the inlet in the same manner as described above for *Stable inlet processes*. However, at these inlets the delivery of sediment by longshore transport produces a preferential accumulation of sand on the updrift side of the ebb-tidal delta. The deposition of this sand causes a deflection of the main ebb channel until it nearly parallels the downdrift inlet shoreline. This circuitous configuration of the main channel results in an inefficient tidal flow through the inlet, ultimately leading to breaching and establishment of a new channel through the ebb-tidal delta. The process normally occurs during spring tides or periods of storm surge when the tidal prism is greatly enlarged. In this state, the ebb discharge piles up water at the entrance to the inlet where the channel bends toward the downdrift inlet shoreline. This causes some of the tidal waters to exit through the marginal flood channel or flow across low regions on the channel margin linear bar. Gradually, over several weeks or convulsively during a single large storm, this process cuts a new channel through the ebb delta thereby providing a more direct and efficient pathway for tidal exchange through the inlet. As more of the tidal prism is diverted through this new main ebb channel, tidal discharge through the former channel decreases, often causing it to infill and become abandoned.

The sand that was once on the updrift side of the ebb-tidal delta, and is now on the downdrift side of the new main channel, is moved onshore by wave-generated and flood-tidal currents. Initially, some of this sand aids in filling the

former channel while the rest forms a large bar complex that eventually migrates onshore and attaches to the downdrift inlet shoreline. The ebb-tidal breaching process results in a large batch of sand bypassing the inlet. Similar to the stable inlets discussed above, some sediment bypasses these inlets in a less dramatic fashion, grain-by-grain on a continual basis.

It is noteworthy that at some tidal inlets, the entire main ebb channel is involved in the ebb-tidal delta breaching process, whereas at others just the outer portion of main ebb channel is deflected. In both cases, the end product of the breaching process is channel realignment that more efficiently conveys water through the inlet, as well as sand being bypassed in the form of a bar.

Inlet Migration and Spit Breaching. Yet another method of inlet sediment bypassing occurs at migrating inlets. In this situation, an abundant sand supply and a dominant longshore transport direction cause spit extension from the barrier terminus. To accommodate this process, the inlet migrates by eroding the downdrift barrier shoreline. Along many coasts, as the inlet is displaced farther along the downdrift shoreline, the inlet channel to the backbarrier lengthens and retards the exchange of water between the coastal ocean and backbarrier. This condition leads to large water level differences between the ocean and bay, making the barrier highly susceptible to breaching, particularly during storms. Ultimately, when the barrier spit is breached and a new inlet is formed in a hydraulically more favorable position, the tidal prism is diverted to the new inlet, often followed by closure of the old inlet. When this happens, the sand comprising the ebb-tidal delta of the former inlet is transported onshore by wave action and commonly takes the form of a landward migrating bar complex. It should be noted that when the inlet shifts to a new position along the updrift shoreline, a large quantity of sand has effectively bypassed the inlet. The frequency of this inlet sediment bypassing process is dependent on inlet size, rate of migration, storm history, and backbarrier dynamics.

Bar complexes. Depending on the size of the inlet, the rate of sand delivery to the inlet, the effects of storms, and other factors, the entire process of bar formation, its landward migration, and its attachment to the downdrift shoreline may take from 6 to 10 years. The volume of sand bypassed can range from 100,000 to $>1,000,000 \text{ m}^3$. The bulge in the shoreline that is formed by the attachment of a bar complex is gradually eroded and smoothed as sand is dispersed to the downdrift shoreline and transported back toward the inlet.

In some instances, a landward migrating bar complex forms a saltwater pond as the tips of the arcuate bar weld to the beach and stabilize its onshore movement. Although the general shape of the bar and pond may be modified by overwash and dune building activity, the overall shoreline morphology is frequently preserved. Lenticular-shaped coastal ponds or marshy swales become diagnostic of bar migration processes and are common features at many inlets.

Tidal Inlet Effects on Adjacent Shorelines

In addition to the direct consequences of spit accretion and inlet migration are the effects of volume changes in the size of ebb-tidal deltas, sand losses to the backbarrier, processes of inlet sediment bypassing, and wave sheltering of the ebb-tidal delta shoals (FitzGerald 1988).

Number and Size of Tidal Inlets

The degree to which barrier shorelines are influenced by tidal inlet processes is dependent on their size and number. As the O'Brien Relationship demonstrates, tidal prism governs the size or cross-sectional area of an inlet. This concept can be expanded to include an entire barrier chain in which the size and number of inlets along a chain are primarily dependent on the amount of open water area behind the barrier and the tidal range of the region. In turn, these parameters are a function of other geological and physical oceanographic factors. Wave-dominated, microtidal coasts tend to have long barrier islands and few tidal inlets and mixed-energy coasts have short stubby barriers and numerous tidal inlets (Hayes 1975, 1979). Presumably, the mesotidal conditions produce larger tidal prisms than along microtidal coasts, which necessitate more openings in the barrier chain to maintain the tidal exchange. Many coastlines follow this general trend but there are many exceptions due to the influence of sediment supply, bay area, and other geological controls (Davis and Hayes 1984).

Tidal Inlets as Sediment Sinks

Tidal inlets not only trap sand temporarily on their ebb-tidal deltas but are also responsible for the longer-term loss of sediment moved into the backbarrier. At inlets dominated by flood-tidal currents, sand is continuously transported landward enlarging flood-tidal deltas and building bars in tidal creeks. Sand can also be transported into the backbarrier of ebb-dominated tidal inlets during intense storms. During these periods, increased wave energy produces greater sand transport to the inlet channel. At the same time the accompanying storm surge heightens the water surface slope at the inlet resulting in stronger than normal flood tidal currents. The strength of the flood currents, coupled with the high rate of sand delivery to the inlet, results in landward sediment transport into the backbarrier. Along the Malpeque barrier system in the Gulf of St. Lawrence (New Brunswick, Canada), it has been determined that over a 33-year period, 90% of the sand transferred to the backbarrier took place at tidal inlets and at former inlet locations along the barrier (Armon 1979).

Sediment may also be lost at migrating inlets when sand is deposited as channel fill. If the channel scours below the base of the barrier lithosome, then the beach sand that fills this channel will not be replaced entirely by the deposits excavated on the eroding portion of the inlet throat. Because up to

40% of the length of barriers is underlain by tidal inlet fill deposits ranging in thickness from 2 to 10 m (Moslow and Heron 1978; Moslow and Tye 1985), this volume represents a large long-term loss of sand from the coastal sediment budget. Another major sediment sink at migrating inlets is associated with the construction of recurved spits that extend into the backbarrier. For example, along the East Friesian Islands of Germany, recurved spit development has caused the lengthening of barriers along this chain by 3–11 km since 1650. During this stage of barrier evolution, the large size of the tidal inlets permitted ocean waves to transport large quantities of sand around the end of the barrier forming recurves that extend far into the backbarrier. Due to the size of the recurves and the length of barrier extension, this process has been one of the chief natural mechanisms of bay infilling (FitzGerald and Penland 1987).

Changes in Ebb-Tidal Delta Volume

Ebb-tidal deltas represent huge reservoirs of sand that may be comparable in volume to that of the adjacent barrier islands along mixed-energy coasts (i.e., northern East and West Friesian Islands, Massachusetts, southern New Jersey, Virginia, South Carolina, and Georgia). For instance, the combined ebb-tidal delta volume of Stono and North Edisto Inlets in South Carolina has $197 \times 10^6 \text{ m}^3$ of sand and the intervening Seabrook-Kiawah Island barrier complex contains $252 \times 10^6 \text{ m}^3$ (Hayes et al. 1976). In this case, the deltas comprise 44% of the sand in the combined inlet-barrier system. The magnitude of sand contained in ebb-tidal deltas suggests that small changes in their volume dramatically affect the sand supply to the adjacent ocean shorelines.

A transfer of sand from the barrier to the ebb-tidal delta takes place when a new tidal inlet is opened, such as the formation of Ocean City Inlet when Assateague Island, Maryland, was breached during the 1933 Hurricane. Initially, the inlet was only 3 m deep and 60 m across but quickly widened to 335 m when it was stabilized with jetties in 1935. Since the inlet formed, more than $14 \times 10^6 \text{ m}^3$ of sand have been deposited on the ebb-tidal delta. Trapping the southerly longshore movement of sand by the updrift north jetty and growth of the ebb-tidal delta have led to serious erosion along the downdrift beaches. The northern end of Assateague Island has been retreating at an average rate of 11 m/year. The rate of erosion lessened when the ebb tidal delta reached an equilibrium volume and the inlet began to bypass sand (Stauble and Cialone 1996).

In contrast to the cases discussed above, the historical decrease in the inlet tidal prisms along the East Friesian Islands has had a beneficial effect on this barrier coast. From 1650 to 1960, the reclamation of tidal flats and marshlands bordering the German mainland, as well as natural processes, such as the building and landward extension of recurved spits, decreased the size of the backbarrier by 80%. In turn, the

reduction in bay area decreased the inlet tidal prisms, which led to smaller sized inlets and smaller ebb-tidal deltas. Wave action transported ebb-tidal delta sands onshore as tidal discharge decreased. This process increased the supply of sand to the beaches and aided in lengthening of the barriers (FitzGerald et al. 1984).

Wave Sheltering

The shallow character of ebb-tidal deltas provides a natural breakwater for the landward shorelines. This is especially true during lower tidal elevations when most of the wave energy is dissipated along the terminal lobe. During higher tidal stages, intertidal and subtidal bars cause waves to break offshore expending much of their energy before reaching the beaches. The sheltering effect is most pronounced along mixed-energy coasts where tidal inlets have well-developed ebb-tidal deltas.

The influence of ebb shoals is particularly well illustrated by the history of Morris Island, South Carolina, that forms the southern border of Charleston Harbor. Before human modification, the entrance channel to the harbor paralleled Morris Island and was fronted by an extensive shoal system. In the late 1800s, jetties were constructed at the harbor entrance to straighten, deepen, and stabilize the main channel. During the period prior to jetty construction (1849–1880), Morris Island had been eroding at an average rate of 3.5 m/year. After the jetties were in place, the shoals eroded and gradually diminished in size and so did the protection they afforded Morris Island, especially during storms. From 1900 to 1973, Morris Island receded 500 m at its northeast end increasing to 1100 m at its southeast end, a rate three times what it had been prior to jetty construction (FitzGerald 1988).

Effects of Inlet Sediment Bypassing

Tidal inlets interrupt the wave-induced longshore transport of sediment along the coast, affecting both the supply of sand to the downdrift beaches and the position and mechanisms whereby sand is transferred to the downdrift shorelines. The effects of these processes are exhibited well along the Copper River Delta barriers in the Gulf of Alaska. From east to west along the barrier chain, there is an increase in width of the tidal inlets and size of the ebb-tidal deltas (Hayes 1979). In this case, inlet width can be used as a proxy for its cross-sectional area. These trends reflect an increase in tidal prism along the chain, which is caused by an increase in bay area from east to west while tidal range remains constant. Also quite noticeable along this coast is the greater downdrift offset of the inlet shoreline in a westerly direction. This morphology is coincident with an increase in the degree of overlap of the ebb-tidal delta along the downdrift inlet shoreline. The offset of the inlet shoreline and bulbous shape of the barriers are produced by sand being trapped at the updrift eastern end of the barrier. The magnitude of shoreline progradation (seaward accretion) is a function of inlet size and extent of its ebb-tidal

delta. What we learn from the sedimentation processes along the Copper River Delta barriers is that tidal inlets can impart a very important signature on the size and form of the barriers (FitzGerald 1996).

Drumstick Barrier Model. In an investigation of barrier islands shorelines in mixed energy settings throughout the world, Hayes (1979) noted that many barriers exhibit a drumstick barrier island shape. In this model the “meaty” portion of the drumstick barrier is attributed to waves bending around the ebb-tidal delta producing a localized reversal in the longshore transport direction. This process reduces the rate at which sediment bypasses the inlet, resulting in a broad zone of sand accumulation along the updrift end of the barrier. The thinner downdrift “bone” of the drumstick is formed through spit accretion. Later studies demonstrated that landward-migrating bar complexes from the ebb-tidal delta guide barrier island morphology and overall shoreline erosional-depositional trends, particularly in mixed-energy settings.

Studies of the Friesian Islands demonstrate that inlet processes exert a strong influence on the dispersal of sand and in doing so dictate barrier morphology (FitzGerald et al. 1984). In addition to drumsticks, the East Friesian Islands exhibit many other shapes. Inlet sediment bypassing along this barrier chain occurs, in part, through the landward migration of large swash bars (>1 km in length) that deliver up to 300,000 m³ of sand when they weld to the beach. In fact, it is the position where the bar complexes attach to the shoreline that determines the form of the barrier along this coast. If the ebb-tidal delta greatly overlaps the downdrift barrier, then the bar complexes may build up the barrier shoreline some distance from the tidal inlet, forming *humpbacked barriers*. If the downdrift barrier is short and the ebb-tidal delta fronts a large portion of the downdrift barrier, then bar complexes weld to the downdrift end of the barrier forming *downdrift bulbous barriers*.

Human Influences

Dramatic changes to inlet beaches can also result from human influences, including the obvious consequences of jetty construction that reconfigures an inlet shoreline. By preventing or greatly reducing the ability of the inlet to bypass sand, the updrift beach progrades while the downdrift shoreline erodes due to depletion or exhaustion of sediment supply. There can also be more subtle human impacts that can equally affect inlet shorelines, especially those associated with changes in inlet tidal prism, sediment supply, and the longshore transport system. Nowhere are these types of impacts better demonstrated than along the central Gulf Coast of Florida where development has resulted in the construction of causeways, extensive backbarrier filling and dredging projects, and the building of numerous engineering structures along the coast. A detailed study of this region by Barnard and Davis (1999)

has revealed that since the late 1980s, 17 inlets have closed along this coast and at least five closures can be traced to human influences caused primarily by changes in tidal prism. For example, access to several barriers has been achieved through the construction of causeways that extend from the mainland across the shallow bays. Along most of their lengths, the causeways are dike-like structures that partition the bays, thereby also altering bay areas and tidal flow patterns. In some instances, tidal prisms were reduced to a critical value causing inlet closure. At these sites, the tidal currents were unable to remove the sand dumped into the inlet channel by wave action. Similarly, when the Intracoastal Waterway was constructed along the central Gulf Coast of Florida in the early 1960s, the dredged waterway served to connect adjacent backbarrier bays thereby redistributing the volume of water that was exchanged through the connecting inlets. This major waterway lessened the flow going through some inlets (causing some to close) while at the same time increasing the tidal discharge and causing the expansion of other inlet systems (Barnard and Davis 1999).

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Tidal Power

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Tidal Energy

Tidal energy is derived from the earth's inherent force, the earth's rotation within the disturbing field of moon and sun. In relatively shallow seas, friction dissipates almost all the energy. Though tides can be predicted very accurately, they

are not in phase with moon and sun movements, the tide waves being distorted as landmasses, narrow passages and shallow depth areas impede and/or influence their progression (Charlier 1982). The tidal current is the rotary current accompanying the turning tide crest in an open ocean. It becomes a reversing current, nearshore, moving in and out as flood and ebb currents. Both of these can be harnessed to produce mechanical and/or electrical power (Charlier and Justus 1993).

Major disadvantage of tidal power electricity generation is that tides are linked to the lunar rather than to the solar cycle and vary in range throughout the year due to their periodic components. These negative aspects can be partially overcome by ingenious engineering and retiming of use of potential energy accumulated at low-demand periods. Such retiming had even been suggested several decades ago using retaining basins, compressed-air, or hydrogen – even electrolytically produced – usable as fuel when tide and peak-power demand are not synchronic (Gilbert 1982). Storing has been proposed in exhausted deposits, abandoned mines, and artificial cavities.

Of the 3000 million kilowatts equivalent dissipated by tidal energy, one billion (10^9) develop in shallow seas. But only a fraction of this “power” can be captured. The geographical site must be suitable from engineering and economics viewpoints and the “usable head” has to be high (5 m or more). The latter requirement has drastically changed with the development of low, and ultra-low, head turbines, greatly increasing the number of potential sites; thus, the 200 million kW theoretically harnessable 20 years ago has sizably increased. The economic geographical factor has also been altered as a result of improvement in transmission possibilities; indeed, national grid systems and high voltage transmission lines have minimized the problem of distance between generating plant and consumer (e.g., in Canada), and the problem of protection (insulation) against extreme temperature amplitudes was resolved more than a decade ago by the Soviets. Finally, where the emphasis has been placed on huge, even gigantic (e.g., Chausey Islands, France) projects, small plants are often favored currently, an advantage for developing countries or isolated areas (Suriname, Half Moon Cove [Maine, USA], China).

The Plant

The tide mill (gm) and the tidal power plant are similar. The latter has an electric generator. The plant's major components are a dam which houses the powerhouse, a retaining basin, and a link to the electricity grid. Construction of a barrage or dam is necessary; earthen dams have been used in the construction of Chinese plants. The dam consists additionally of dikes connecting with the natural embayment and a sluiceway. A passage way for fishes is now provided. A reasonable tidal range is required, although quite small ones will suffice today. Estuaries and gulfs in shallow areas are privileged sites. A closed basin is thus created which fills and empties daily,

sometimes more than once, depending on the local tidal regime. Equipped with sluices and turbines placed in the barrage, the system retains the water entering at flood tide and releases it at ebb tide.

Generally, turbines produce electricity as the water flows out of the retention basin (or “pool”) but reversible blade (the so-called *bulb*) turbines, originally invented by Harza and later developed by French engineers (Hollenstein and Soland 1982), can produce electricity both as the water enters and when it exits the basin. Practice – based upon 30 years of operation of the Rance River, France, plant – has, however, shown that entering water generation is perhaps not always economically profitable. By judiciously selecting tide gates opening and closing times, power generation can be synchronized with peak-demand periods, even if these do not coincide with tide peaks. The hydraulic head can be increased by reversing power units, temporarily turning the turbine-generator into a pump-motor; the bulb turbine precisely regulates flow in both directions and acts as turbine and pump. Pumping, however, consumes electricity.

A so-called “site value coefficient” for plants (k) was calculated by Robert Gibrat based on dam length and natural energy ($k = L/NE$); however, these are not the only “factors” involved in site selection; considered should be tidal range, potential pool size, ratio of basin aperture section to basin surface, length of dam, basin characteristics (such as geometric shape and surface, opening, widening shape), geological structure, lithology and petrology, foundation soil quality, gradient of resistance layer, probability of silting, rate of sedimentation, climate, market distance, competitiveness of conventional and/or alternative (e.g., nuclear, aeolean) power sources. Some economists maintain that today tidal power is competitive and may even be less expensive than fossil fuel and nuclear generation. Calculations made by Voyer and Penel (1957) have been revised (Charlier 1998). All the factors are seldom, if ever, simultaneously favorable, which, with the high capital investment for large plants, probably accounts for the small number of plants constructed (Charlier and Justus 1993).

Types of Plants

Plants are single- or multiple-basins systems, a tide-powered air storage scheme, tide-powered hydrostatic pump scheme or Gorlov setup. Single-basin plants function either as a one-way operation (ebb generation only), two-way operation (generation at both ebb and flood flow), two-way operation with pump-turbines generation (excess used to pump water in storage reservoir[s]), or high-tide pumped storage (nontide-connected power production). Multiple-basin schemes are either double-pool (station placed in-between basin, filling one basin at flood, emptying the other at ebb), both basins

pumping (pumping high pool up and low pool down, pumping at off-peak), pool-to-pool dam pumping, pool-to-pool dam pumping combined with pump-turbines, tide-boosted pumped storage (basins may be of different sizes and in different places), or variations of the latter.

The tide-powered air storage scheme differs in that it uses the tidal energy to drive air turbo-compressors and stores the compressed air. In the tide-powered hydrostatic pump, a propeller turbine is linked to a pump and the tidal energy is converted in a flow of high-pressure oil which turns a Pelton wheel coupled to an alternator. The Gorlov proposal involves a thin plastic barrier hermetically anchored to bottom and bay sides supported by a bay-spanning cable. The tidal energy would be converted into power by an air motor piston at ebb with direct generation or compressed air storage (Gorlov 1982). Finally, one technology would anchor in line a series of floating turbine and generator units along the flow of the tidal current (Charlier and Justus 1993). This approach, less onerous probably than a “traditional” tidal power plant, would perhaps be attractive for small local or regional schemes.

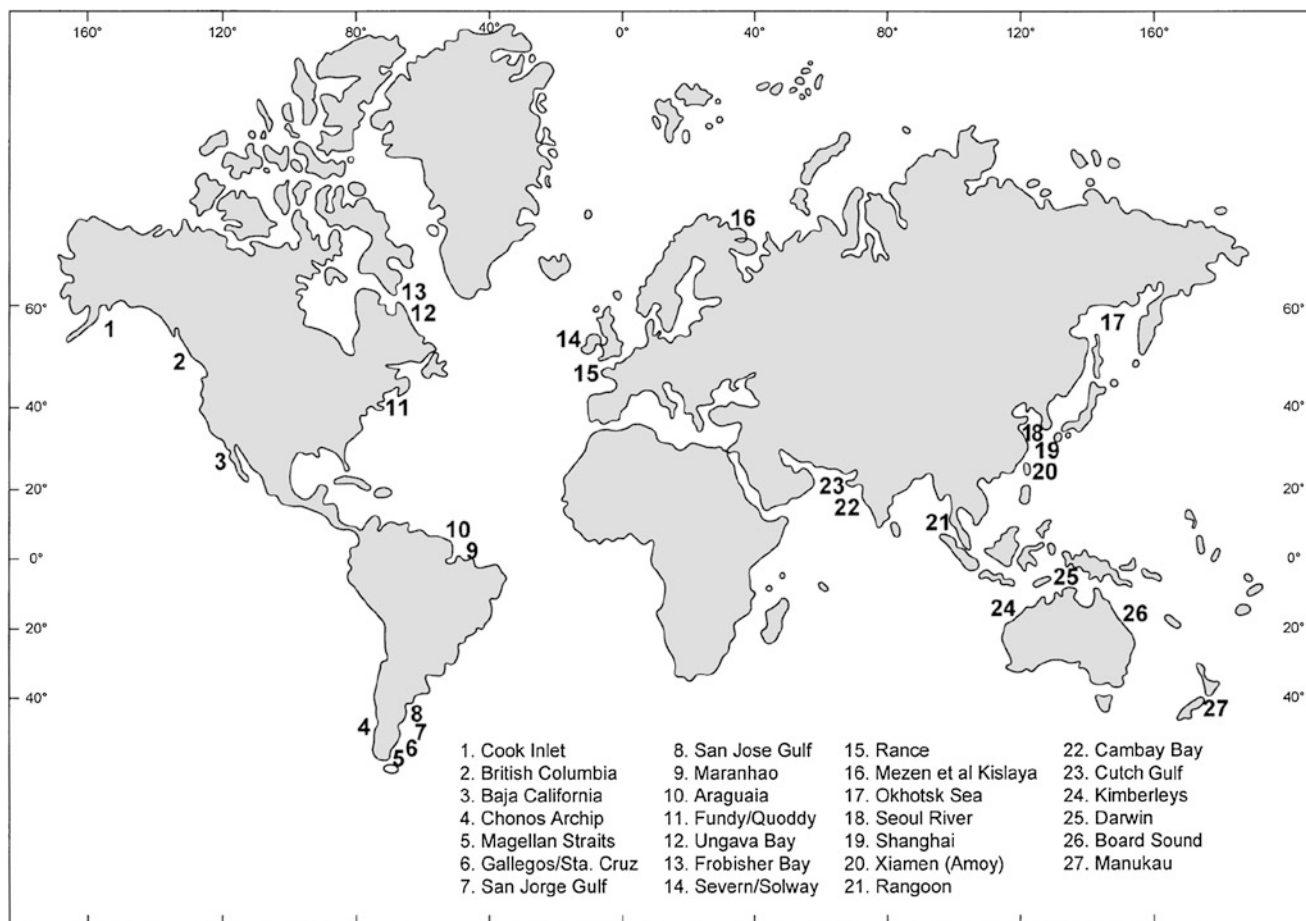
Reversible blade (bulb) turbines have been placed in the Rance River (France) and the Kislogubskaja (Russian Federation) single-basin plants. The Russian plant was built in modules, which considerably reduced construction costs and avoided construction of expensive cofferdams. The turbines can be used as low-head pumps.

Operating Plants

If the literature on tidal power plants is extensive (cf. Charlier 1982, 1998, 2003; Charlier and Justus 1993), it is not easy to always determine where plants were built (Fig. 1). Little is known of a plant built in the mid-1920s in Suriname (then Dutch Guyana) or in Boston Bay (end of nineteenth century and dismantled because of harbor extension). Information on China’s post-World War II “more than one hundred” plants is scant and even in the country itself is poor, though a few papers have been recently published (Ch’iu 1958). Occasionally, construction of plants is announced, as in Korea or India (cf. Sharma 1982; Song 1987), but proves premature. Three major plants have been built and are operating: the Rance River (France), the Kislogubskaja (Russia), and the Annapolis-Royal (Nova Scotia, Canada) plants. The first one is large and dates from 1956, the second, a pilot plant, is much smaller and was heralded as a forerunner of more ambitious ones to come, and the third one is an experimental and pilot installation.

The Rance River Plant

Twenty-four bulb groups of 10,000 kW were placed in horizontal hydraulic ducts entirely surrounded by water at the Rance plant. The cost of the cofferdams represented one-



Tidal Power, Fig. 1 Location of major tidal power plants (after Charlier and Justus 1993, with permission from Elsevier Science)

third of the building cost, an expense which can now be dispensed with as was in the Russian plant and to some extent in the Canadian plant. The cost of the project ran about US\$100 million (in 1966-\$).

Started in 1993 and completed in 1966, the 53 m wide and about 390 m long plant protrudes 15 m above water, with foundations 10 m below sea level, accommodates a four-lane roadway (eliminating a ferry) and has substantially contributed to the economic development of the former lethargic region besides furnishing 544 MW h of usable power. The mean net annual energy production exceeds slightly 500 GW h and the mean capacity throughout the year is 65 MW. It has six modes of operation. The operation policy has aimed at optimizing the value of the energy generated, reaching maximum profit, instead of eyeing to generate the maximum amount of energy. Generation by overemptying the basin (by pumping) has been done only rarely during the last decade.

Environmental aspects at the Rance Plant. Environmental impact, at the French Rance River Plant, has been extremely mild causing only the disappearance of one species of fish. No major biological modifications have occurred. New species

have appeared. Sandbanks have disappeared; high speed currents have appeared near sluices and also near the powerhouse where sudden surges have been observed. Tidal ranges have been reduced from 13.5 m to 12.8 m and minima increased. The fishing industry has not been affected contrarily to fears, and tourism has increased.

The Russian Plant

The Kislaya Bay plant is located near Mezen on the Arctic. The small Russian Arctic coastal bays hold a 3.2 billion kW potential. Completed in 1968, the 18 m long, 36 m wide, 15 m high, with a 15.2 m head, plant was built by towing preconstructed concrete caissons to the site and then sinking them into place. A narrow passage connects bay and sea; no long dikes were necessary as the powerhouse itself closes the basin. Transmission lines existed close to the site. The power set capacity is 400 kW. The reversible hydraulic turbine is coupled with a synchronous generator, with the help of a multiplier.

Russians are still considering a variety of schemes and stress that by harnessing tides in the Sea of Okhotsk bays, some 174 billion kW h could be produced.

The Canadian Plant

An experimental plant was constructed in the early 1980s at Annapolis- Royal (Nova Scotia) in the Bay of Fundy whose “electricity potential” is huge; studies on Fundy tidal power go back to at least 1944. It is here that the first tide mill was built in North America (1607) using tidal and river water. A low-head STRAFLO turbine has been installed in the new plant whose generator and turbine form a coupled unit without driving shaft, particularly compact, and thus cutting powerhouse costs (Douma and Stewart 1982). The costs amounted to CA\$46 million (approximately US\$44 at the time). Situated on Hog’s Island in the middle of an intake canal on the lower reaches of the Annapolis River where tidal ranges reach 7 m, it took advantage of an existing dam protecting farm land. The Chinese, similarly, have commonly used existing (often earthen) dams to install tidal plants.

The turbine is a single effect one and generates electricity only during discharge of water from the retaining pool. In the event that the pool level would be below sea level, the turbine’s water passage would sluice seawater into the basin during each tidal cycle. A cofferdam of local material was used to construct, in the dry, the powerhouse. Turbine and generator are located in the center of the turbine pit. The dam is 225 m long, 60 m wide at low water level, and 18 m wide at the crest. A fish pathway is provided. Fifty million kilowatts are generated.

Environmental impact. The Canadian project was designed to assess both the operational characteristics and the reliability of the Straflo turbine, and, of course, the possibility of implementing major tidal power utilization in eastern Canada. The impact study showed that interactions can be mitigated and that physical, biological, and human impact would be minimal and residual impacts small, while economic outfall would be considerable both for the short- and long-terms.

The Chinese Plants

Information is difficult to secure, even in China. Claims of a large number of plants – as many as 105 – and their successful operations have been repeatedly made and so have plans to undertake new construction (Ch’iu 1958). There are also records of the “temporary” abandon of some plants due to siltation.

Chinese engineers acknowledged a problem with siltation whose extent proves difficult to assess, but it has been said that some tidal power plants were put out of commission because of sediments accumulation. A comparative study on devising siltproof systems was carried out using data from the Baishakou tidal power plant and envisioned environmental protection to control sedimentation in the tallwater channel and reservoir (Zhu 1992); at about the same time other studies explored the optimum patterns for double-effect single-basin plants (Wang et al. 1991) and calculated (modelization) of the

optimum tidal energy at any location (Soo et al. 1994). Zhikui Zhu presented in 1992 an analysis of data pertaining to the Baishakou tidal power station. It describes measures apparently taken to control sedimentation in the tallway channel and the reservoir area. Comprehensive management is outlined, including mechanical sandproofing methods and environmental protection.

The interest of Chinese researchers in tidal power plants has not waned, nor has it been somnolent as in Europe and the United States, due undoubtedly to the abundance and generally low cost of fossil fuel. An exploratory study and modelization were recently conducted (Ji 1991).

A review article on Chinese activity in tidal power utilization covering the 1950–1990 period was published in the somewhat unfamiliar [to westerners] *Collection of Oceanographic Works*. The information rather contradicts other releases which talk about a hundred small tidal power plants. Guixiang Li (1991) mentions eight small power stations and discusses their tangible economic and social benefits. He further discloses that “many more” small and medium size, even one or two large, plants “will be built by the year 2000” along the coasts of China’s mainland and the coastal islands. The publication is rather difficult to obtain, except perhaps through the good offices of the University of Karlsruhe (Germany).

Other Geographical Locations

No accurate information is available about the two-basin plant that functioned in Boston harbor at the turn of the preceding century until it was dismantled to make room for the port’s extension, nor of the Van Bemmelen plant in Suriname. The tidal current plant in Iceland seems to have left no traces either. Plans were afoot in the 1990s to build a tidal power plant in [South] Korea (e.g., Garolim Bay) but came to a halt for political reasons (recognition of North Korea by France whose Sogreah was to build the facility) (Song 1987). Other sites which have been considered and for which periodically plans resurface are the Kimberleys in Australia, San Jorge Gulf in Argentina (Aisiks and Zyngierman 1984), and the Severn River in Great Britain (Severn Barrage Committee 1986). There are of course numerous other suitable sites worldwide (Charlier and Justus 1993), and studies were conducted for some of them, for instance for India (Sharma 1982).

Tidal stream’s rapid currents have been recently examined as sources of power, for example, in the Orkney and Shetland islands (Bryden et al. 1993–1995), perhaps using Darrieus turbines (Khio et al. 1996), and conversion of kinetic to electrical energy with a barrage in New York’s East River (Birman 1994). In Maine and South Carolina, experiments proved that using tidal stream power with speeds up to 1.5 knots cuts the cost of seeding rafts with a wooden scoop on the bow that

directs the following water to an enclosed compartment containing upwelling units and is less expensive than using other sources of power (Hadley et al. 1994; Rhode et al. 1994).

Cross-References

- [Engineering Applications of Coastal Geomorphology](#)
- [Microtidal Coasts](#)
- [Tidal Environments](#)
- [Tidal Prism](#)
- [Tide Mill](#)
- [Tide-Dominated Coasts](#)
- [Tides](#)
- [Wave Power](#)

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Tidal Prism

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The tidal prism is the amount of water that flows into and out of an estuary or bay with the flood and ebb of the tide, excluding any contribution from freshwater inflows. For this reason, it is often reported as the volume of the incoming tide, and the contribution of river inflow calculated from the difference of the ebb and flood volumes. The tidal prism can be determined from hydrographic charts, where the volume in the estuary between low water and high water is calculated from sounding data. The tidal prism can also be determined by measuring the amount of water flowing into an estuary using a technique known as a tidal gauging. This is normally undertaken at a narrow section in an estuary of regular shape, and at a time when river inputs are low. Measurements of current velocity and water depth are made continuously at various points throughout the section using current meters suspended from a boat or bridge, along with measurements of water (tide) level in the section over the tide. The tidal prism is computed as the product change in cross-section area and mean velocity in section, integrated over half the tidal cycle. Today a vessel mounted ADCP (acoustic doppler current profiler) which returns depth, tidal change, and current at many points in the water column as the vessel travels back-and-forth across the section greatly simplifies and improves the accuracy of a tidal gaugings.

The tidal prism is an important metric for an estuary. It is an indicator of the hydrodynamic processes operating in an estuary. In shallow estuaries where the tidal prism forms a large proportion of the water in the estuary at high tide, tidal

processes dominate and flushing is good. In deep estuaries with relatively small tidal prisms density-driven flows and river inputs play a greater role in the hydrodynamics. In estuaries where there are semidiurnal tides and a large difference in spring and neap tidal ranges, the amount of water flowing into the estuary (the tidal prism) on a spring tide may be double that on a neap tide and tidal currents and sand transport increase along with this. Empirical relationships developed between tidal prism and various estuary parameters help with the conceptual understanding of processes and provide a simple tool for easy calculations of stable channel design (e.g., Bruun 1990).

The tidal prism is related to inlet dimensions, the amount of sand stored in tidal deltas, and has been used to describe inlet stability. The empirical relationship between the gorge cross-sectional area and tidal prism of a tidal inlet has been described in numerous studies (O'Brien 1931; Jarrett 1976; Hume and Herdendorf 1992) by:

$$A = c\Omega^n,$$

where A is the gorge cross-sectional area (m^2), Ω is the tidal prism (m^3) and c and n are constants. Inlet gorges that are stable conform to the relationship and there is considered to be a balance between the inlet geometry and tidal flow through the gorge. Those lying off the line are out of equilibrium and characterized by either scour or deposition. A similar relationship has been found to exist between the ebb tidal delta sand volume, which increases with increasing tidal prism. In this situation, the volume of the sand body also increases with decreasing wave energy and as the angle that the ebb jet makes with the beach on the adjacent barrier shore increases (Walton and Adams 1976; Hicks and Hume 1996). On coasts where there is little wave energy and estuaries have large tidal prisms the delta will be elongated offshore under the influence of the ebb tidal jet. Where estuaries with small tidal prisms occur on coasts with high wave energy the ebb tidal sand body will be more flattened against the shore. A – Ω relationships have been used to quantify the morphological stability of tidal inlets. Bruun and Gerritsen (1960) showed that the size of the tidal gorge is one of the main factors determining the ability of flow to transport sediment through the entrance. Inlet gorges that are morphologically stable (i.e., have the ability to return to their original configuration after a disturbance) conform to the relationship because there is a balance between tidal flow and littoral drift to the gorge, so the inlet stays open. They demonstrated that when the ratio of the tidal prism to total (gross) annual littoral drift delivered to the inlet from the ocean is in excess of 300 the inlet has a high degree of stability. In cases where the ratio is less than 100 there will be a low degree of stability and entrance bars will shallow and navigation difficult. A – Ω relationships like those for tidal inlets on sandy shores also hold for many

different estuary types ranging from lagoon to river mouth and even to large coastal embayments, and have been used to characterize and classify inlets (Hume and Herdendorf 1993).

Cross-References

- Beach and Nearshore Instrumentation
- Estuaries
- Tidal Environments
- Tidal Flats
- Tidal Power
- Tide Gauges
- Tide-Dominated Coasts
- Tides

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Tide Gauges

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Historical Origin

Tide gauges have a relative long history. At some places, systematic sea level observations have been performed and recorded since the late seventeenth or early eighteenth

Paolo A. Pirazzoli: deceased.

century. This is the case in Amsterdam, for instance, since 1682, in Liverpool since 1768, or in Stockholm since 1774.

The first devices were simply graduated rods, usually called tide poles, placed at locations where the instantaneous water level height of the sea could be read off at any time by an observer. Most of such measurements were undertaken in or at the entrance of harbors. They were restricted to observations of high and low water levels, as well as the time of their occurrences (Cartwright 1999).

Automatic recording devices appeared only in the 1830s, although basic instructions were already detailed and published in an Italian journal in 1675. An extract was transcribed in the “*Journal des Sçavans*” of April 22, 1675 (Observatoire de Paris 1675). These were mechanical gauges, equipped with float, wires, counterweights, clock, pen, paper chart recorder, and stilling well. They provided the first complete tidal curves which could be examined in detail and digitized for further analysis.

Tide or Sea-Level Gauges?

Whatever technique is employed, the basic quantity provided by tide gauges is an instantaneous height difference between the level of the sea surface and the level of a fixed point on the adjacent land. The vertical reference point is a nearby benchmark which can be observed by traditional surveying techniques. Thence, tide gauges not only record ocean tides but also a large variety of sea-level signals that can be caused by variations in atmospheric pressure, density, currents, continental ice melt. . . as well as vertical motions of the land upon which the measurement instrument is located, due to tectonic changes, isostatic adjustments, volcanism inflation, sediment consolidation, pier subsidence, etc. Records of such devices are indicative of what are called relative sealevel changes (Pugh 1987). The recorded processes have characteristic timescales from several minutes to centuries.

Therefore, tide gauge data is valuable information to a wide range of activities over a variety of timescales, for scientific research as well as for many practical applications. For instance, data have uses: for navigation and ship traffic (guidance, tidal timetables. . .), for coastal engineering design (dike building, dredging works. . .), for statistics of extreme levels over long periods, for studies of upwelling and fisheries, for hydrographic surveys (sound charts, chart datum. . .), for storm-surge predictions and alert, for analysis of the risk of flooding and coastal protection, for input or validation of ocean circulation models, for long-term trend of sea-level variations, for global change research and monitoring and, of course, for tidal analysis, prediction and validation of tidal models. In spite of the numerous applications, the historical and conventional term of tide gauge still prevails for such devices, although sea-level gauge would be more adequate.

Technical Evolution

The most common type of gauge in use around the world still consists of the float and stilling well system which was devised more than a century ago. In this system, the height measurement of a floating gauge is taken by measuring, on a reduced scale, the length of the wire holding a counter-balanced float. The float sits on the surface of the water inside a well. The vertical movement of the float is transmitted and reduced in scale through a more or less sophisticated system of wires, pulleys, and counterweights to a pen. A continuous record of water height against time is obtained in this way as a drawn curve on paper, the paper being in motion at a fixed speed in a normal direction to the pen displacement. The stilling well is a vertical tube long enough to cover any possible range of tide. It prevents the float from drifting in the presence of currents or winds. The well is designed to provide a mechanical filtering of short-period oscillations due to waves by restricting the flow of water into and out of the well.

Traditional mechanical float devices are progressively replaced by new technologies. Modern type gauges are mainly based either on the measurement of the subsurface pressure or on the measurement of the time of flight of a pulse, acoustic or radar.

The principle of a pressure system is the measurement of the hydrostatic pressure of the water column above a fixed point below the lowest expected tide level. The conversion of that hydrostatic pressure into a sea-level equivalent height is performed according to the law: $h = (p - p_a)/(\rho g)$ by measuring or assuming values of water density ρ , air pressure p_a , and local acceleration due to gravity g . Pressure sensors usually exploit strain gauge or ceramic technology. Water pressure translates then into changes in resistance or capacitance in the pressure element. The resulting signal is normally a specific frequency which is converted into physical units of pressure.

The acoustic or radar systems determine the vertical distance from a transducer, located above the sea surface, to the water by measuring the elapsed time of a pulse that is emitted, reflected, and returned back to the sensor. The distance to the water is then derived from the velocity of the type of wave considered (sound or radar).

Whereas technical description may be provided by manufacturers, IOC manuals (1985, 1993, 1997) are a helpful source of information. Valuable detailed information can be found there on each type of gauge, their respective advantages, drawbacks, performances, and limitations, as well as advice on operational methods and environmental conditions of use.

A critical part of the tide gauge system is probably the benchmark on land as it provides the fundamental zero point or datum to which the values of sea level are referred. This benchmark is extremely important as it serves to build useful

long-term sea-level time series, even if parts of the time series were obtained from different gauges and different benchmarks (as long as they were geodetically connected). Tide gauge benchmarks are ultimately the source of the long-term coherence and stability of the measurements. It is therefore common sense to preserve the datum by installing and connecting a set of 5–10 benchmarks within a few hundred meters of the tide gauge. Usually, one of them is arbitrarily called the tide gauge benchmark, the “most” stable, the “most” secure or the closest, although all of them are representative of the datum. Even though the station is equipped with the most modern equipment, long-term instrumental drift and local stability surveying should be performed at least annually.

In the era of modern communication technologies, data in electronic form is essential as it can be retrieved immediately from a gauge to a data center by automated modem dial-up or satellite transmissions. Data can then easily be made accessible worldwide via the Internet. Paper chart recorders are no longer acceptable. They require slow laborintensive digitization and are cost-effective. Moreover, they contain many sources of inaccuracy and are inadequate for certain applications.

Scientific Applications

Many scientific applications other than the natural tidal research and modeling benefit from tide gauge records. For instance, tide gauge data are used to establish vertical reference systems on land and on sea in order to define the height and depth datums. The belief, about a century ago, that the average level of the sea was constant over long periods of time led to define the concept of geoid and, subsequently, to establish the origin of the leveling networks on “mean sea level.” Typically, countries chose one tide gauge station to compute this quantity over an arbitrary time period: in France, for example, the datum was determined at Marseilles from continuous tide gauge records performed during the period 1885–97; in Britain, the Ordnance Datum was determined at Newlyn from records extending from May 1915 to April 1921. However, mean sea level varies from place to place and at one specific place over time. Today, mean sea level at Marseilles is about 11 cm above the local 1885–97 datum, whereas it is about 0.2 m above the Ordnance Datum at Newlyn. Thus, the datums no longer represent the “real” average of the sea level at these sites.

Tide gauge data are also used to establish the datums to which the depths are referred on nautical charts and above which tide predictions are provided for practical purposes. Since 1996, the International Hydrographic Organization has recommended that Lowest Astronomical Tide be adopted as the International Chart Datum. This datum is defined as the

lowest tide level which can be predicted in average meteorological conditions and in any combination of astronomical conditions.

Owing to the rhythmic nature of the tide, the components of its oscillations can be represented by a series of sinusoidal curves, mainly depending on the relative positions of the moon and the sun. In each place, the major tide components are determined empirically from hourly records during at least 29 days (a moon cycle), by the use of harmonic analysis. Because astronomical orbits are known, tides can be predicted at any time. Random deviations from the predicted tide can be due to changes in air pressure or wind (sea surges), currents, changes in water density, discharge (especially at a river mouth) and can be analyzed after filtering the astronomical tide. The occurrence of short-lived oscillations in random deviations, following a sea surge, may correspond to seiches. Finally, if periodic, short-lived and random components are removed by filtering processes, and if the series investigated are long enough, long-term trends will appear.

Tide gauges have been carefully studied for indications of recent global sea-level rise. They include land movement and sea-level movement and depend, therefore, on local or regional tectonics as well as global climate change effects (eustasy). By analyzing the difference in trend between the records in two stations, local and regional components can be easily revealed, though their interpretation may be not univocal. In order to keep the statistical accuracy of a trend estimation below ± 0.5 mm/year, almost continuous records of at least 40–60 years long are usually necessary. At the present time, less than 300 stations in the world can provide such long records. Their geographical distribution is unfortunately very uneven, most of the stations being located in the Northern Hemisphere, with a great majority on both sides of the North Atlantic and few or no data in very wide coastal areas. Since the late 1980s, international projects like GLOSS have made deserving efforts to improve the existing tide-gauge network, but the duration of the new records is still too short to assess long-term trends. In spite of such globally unrepresentative distribution, of the difficulty in separating the various components of relative sea-level change, and of the fact that long-term trends show a great spatial variability, several authors have attempted an estimation of the recent global (average) sea-level rise from tide-gauge records, with various approaches (Emery and Aubrey 1991; Pirazzoli 1996). The results obtained are quite variable: a “global” sea-level rise between 0.5 and 3.0 mm/year is inferred by various authors, with several estimates around 1 mm/year; higher rates are obtained when tide-gauge trends are “corrected” using isostatic models; however, other authors infer that an accurate global sea-level trend is indeterminable of the basis of tide-gauge data alone. This precludes for the moment a more precise quantification of the recent global rise in sea level only with tide gauges.

Perspectives: Synergy With Space Techniques

Recent advances in space geodesy and gravity measurements allow consideration of monitoring of land movements in a global geocentric reference frame rather than in a local one. Repeated or continuous precise positioning of tide gauge benchmarks by geodetic techniques like GPS or DORIS, over periods of a decade or so, will enable vertical crustal movement to be determined and, subsequently, provide a possible discrimination between eustatic sea-level rise and land subsidence within tide gauge records (IOC 1997; Neilan et al. 1998). Moreover, tide gauge data will then be expressed in the same global geodetic reference frame as satellite altimeter observations and can therefore be directly compared and combined with the altimetric sea levels, providing at last a more reliable and accurate estimate of sea-level variations.

Probably the most significant improvement in sea-level research comes from satellite radar altimetry. However, altimetry cannot provide detailed local high-frequency sea-level information due to the specific sampling pattern. Tide gauges continue to possess important attributes for certain applications, like continuity with historic measurements, high accuracy, continuous sampling, and ability to record at the coast at a relatively low cost.

Cross-References

- ▶ [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
- ▶ [Changing Sea Levels](#)
- ▶ [Coastal Climate](#)
- ▶ [Coastal Currents](#)
- ▶ [Geodesy](#)
- ▶ [Global Positioning Systems \(GPS\)](#)
- ▶ [Storm Surge](#)
- ▶ [Submerging Coasts](#)
- ▶ [Tectonics and Neotectonics](#)
- ▶ [Tidal Datums](#)
- ▶ [Tides](#)
- ▶ [Uplift Coasts](#)

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Tide Mill

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Tide Mills (*moulins à marée*, *molinos de mar*, *Gezeitenmoehlen*, *getijdenmolens*) dotted the coasts and estuaries for several centuries until more efficient, but not less costly, alternatives all but wiped them off the map. A few are making a comeback. Some were still functioning during World War II in England and Wales, but most were derelict or their buildings transformed and used for other purposes (Wailles 1941). History records show that tide mills once functioned on the rivers Thames (London Bridge), Tiber (in besieged Rome), and Danube (Charlier and Menanteau 1998). A mill stood at the entrance of Dover Harbor in (1000) according to the *Domesday Book*.

Tide Mills consisted of a dam with sluices, a retaining basin, and a float or a water wheel and transformed the energy of running water into mechanical power to run flour-mills, saw-mills, even breweries, and as late as 1880 to pump sewage. They apparently were also put to work in the *polder-works*.

Tide Mills require sites with tidal amplitude-and thus a tidal current-though even a tidal creek may do. They are the forerunners of today's tidal power plant in the same manner as the windmill is the precursor of the contemporary air turbines. The tide mill is in fact a conventional water mill using the tidal current as its source of power, occasionally both ebb and flood tide, but most used only the ebb current as the retaining basin filled at flood time, emptied at ebb time. A few mills used a proportion of freshwater from the stream in addition to seawater.

Roger H. Charlier: deceased.

The most common idea was the “float method” by which the incoming water raised a floating mass which, as it fell down to original position, provided “work”; another approach included a shaft-mounted rotating paddle wheel activated by ebb and flood, with power transmitted by the shaft; finally somewhat more sophisticated types would let air contained in a metal or concrete conduit be compressed by the incoming tide, thus furnishing compressed air power, an idea which has resurfaced today in proposed schemes (Gorlov 1982). The fourth system, even more elaborate, dams part of the sea (bay, gulf) and this pool fills up at incoming tide; the water when released at low tide passes through turbines to flow back to sea or to another basin. The latter scheme appeared in the nineteenth century and with its wheel rotating at as much as 150 rpm, had a much higher yield.

Tidal mills were quite numerous in England, Wales, The Netherlands, Brittany, and Spain. The Europeans brought them to the United States, seemingly on both coasts (Creek 1952) and Canada. There are some claims that they were not uncommon along the South China Sea coast. Besides the touristic value of restored mills, as near Southampton (Ewing Mill), Plougastel (Brittany), and in Massachusetts (Chatham Spice Mill), tide mills are making a timid come back, in an improved version, to provide power in remote areas.

Cross-References

- Polders
- Tidal Creeks
- Tidal Power
- Tidal Prism
- Tides

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Tide-Dominated Coasts

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As first enunciated by Price (1955), and later elaborated upon by others (e.g., Hayes 1975), tides play an important role in

defining the geomorphology of depositional shores, namely coastal and deltaic plains. On such shores, it is the ratio of tidal energy, usually dictated by tidal range, to wave energy, a function of average wave height, that determine the morphology of the coast, with sediment supply being an important modifier near major river mouths. As a generalization, coasts with small tidal ranges (microtidal ≤ 2 m (Davies 1964)) are dominated by wave energy. Such coasts were termed *wave-dominated coasts* by Price (1955). On the other hand, coasts with large tidal ranges (macrotidal ≥ 4 m) are typically dominated by tidal energy, and were hence termed *tide-dominated coasts* by Price. The effectiveness of wave energy diminishes (i.e., waves cannot break in a concentrated area for a long period of time), and tidal current energy increases as the vertical tidal range increases. Exceptions to this generalization occur where waves are so small that even small tides generate adequate energy to shape the coast. An example of a *tide-dominated coast* with a relatively small tidal range occurs at the head of the embayed coastline of northwest Florida (Hayes 1979). Coasts with intermediate wave energy and tidal ranges (typically mesotidal = 2–4 m) were termed *mixed-energy coasts* by Hayes (1979).

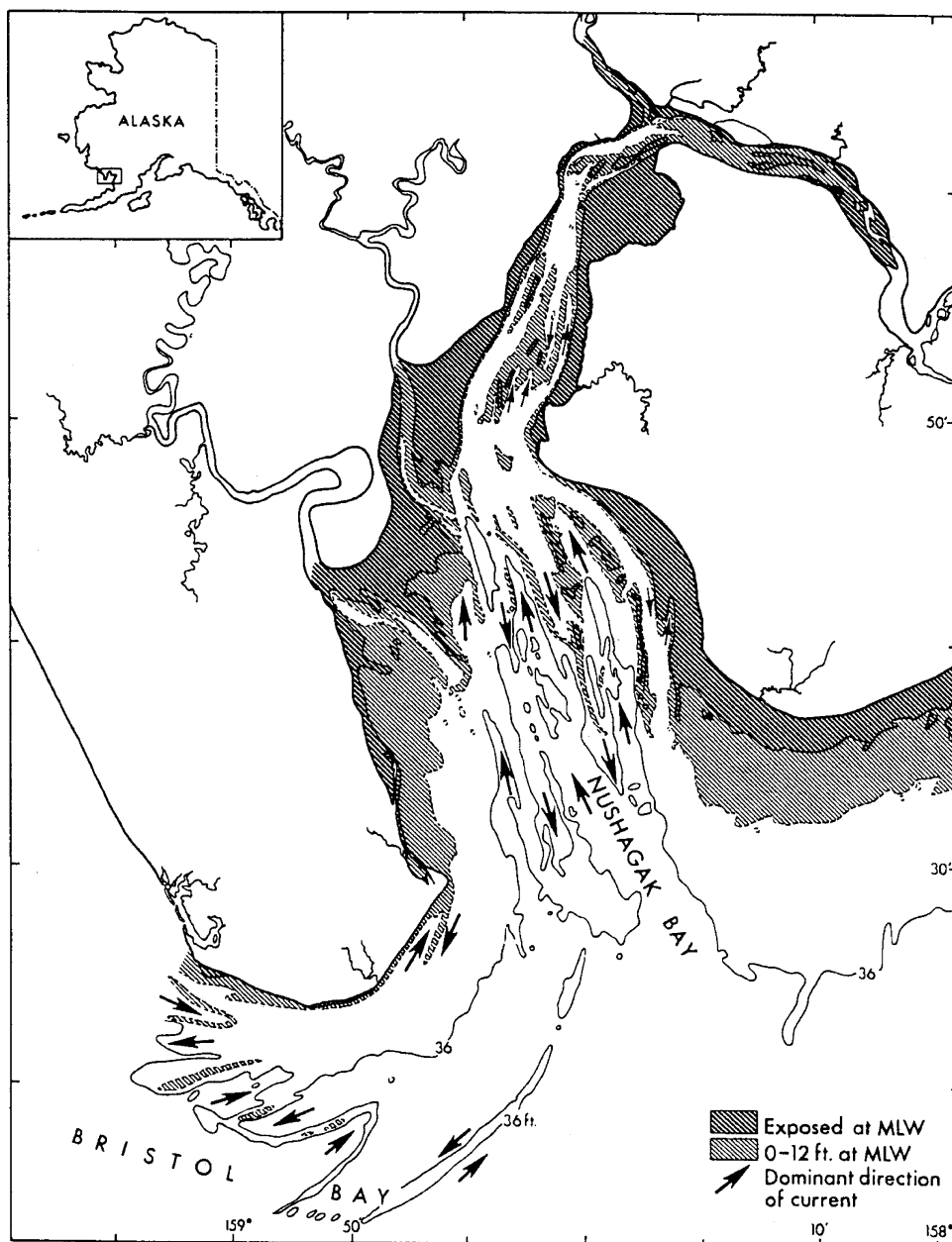
Tide-Dominated, Non-Deltaic Coasts

The bathymetry of a typical tide-dominated, non-deltaic shore, Nushagak Bay, Alaska, is illustrated in Fig. 1. The coastal morphology of major river mouths on tide-dominated coasts are most commonly open-mouthed estuaries, as shown in Fig. 1. Between major rivers on these types of coasts, the shore is occupied by extensive salt marshes and tidal flats. Barrier islands are completely missing, because wave action is not focused enough at a single topographic level to build a barrier island and tidal currents are strong enough to disperse the sand to offshore regions. Examples of this type of coast occur in northwest Australia (Coleman and Wright 1979), western Korea (Kim et al. 1999), the Bay of Bengal, northern end of the Gulf of California (Thompson 1968), the Wash, England (Evans 1975), and many other depositional coasts with tidal ranges greater than 4 m. Generally speaking, the sediment distribution patterns on coasts of this type are exactly opposite to those on wave-dominated coasts, inasmuch as finest sediments occur on mudflats and in wetlands of the upper intertidal zone and coarsest sediments occur lower in the intertidal zone and offshore where tidal currents are strongest.

Studies by Evans (1975) documented the characteristics of the sediments of the tidal flats of the Wash, an indentation in the coastline of the east coast of England, which has a tidal range of 7.0 m. The seaward portions of the flats are made up of complex sand bodies covered by sand waves, whereas the upper part is fringed by a salt marsh composed of fine-grained sediments. Studies of the tidal flats of the Bay of Fundy, by Knight and

Tide-Dominated Coasts,

Fig. 1 Nushagak Bay, Alaska, a tide-dominated embayment (tidal range = 5.5 m). Note lineation of shoals parallel with tidal currents



Dalrymple (1975), Dalrymple et al. (1990), Yeo and Risk (1981), and numerous others, describe the sedimentology, sediment transport dynamics, and potential stratigraphy of this intertidal zone, which has the largest tidal range in the world.

Figure 2 gives a hypothetical prograding stratigraphic sequence for the intertidal zone of a tide-dominated coast based on the references cited above and other sources.

Tide-Dominated Deltas

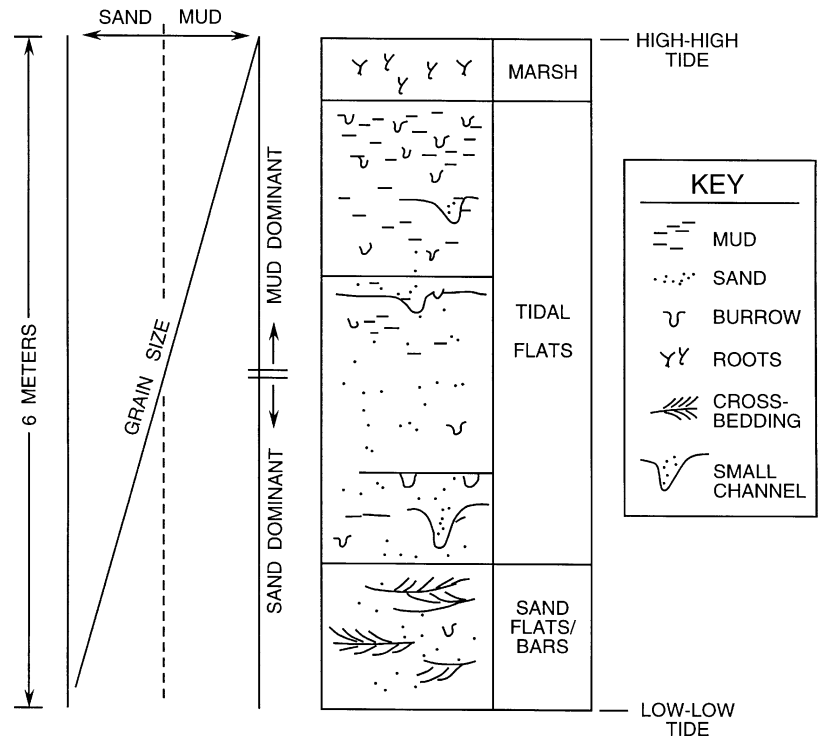
On tide-dominated coasts where the rivers have enough sediment load to have filled the antecedent lowstand valley and

build a bulge in the shore (within the time frame of the present highstand), the resulting river deltas are typically referred to as *tide-dominated deltas* (Fisher et al. 1969; Galloway 1975). However, the distinction between such deltas and estuaries is somewhat obscure. Tide-dominated deltas usually are composed of a series of funnel-shaped water bodies (estuaries?) at multiple river mouths with a series of shore perpendicular sand ridges that extend offshore of the river mouths. Broad tidal flats and marsh or mangrove wetlands occur along the shore of the embayments. A generalized model of a tide-dominated delta is given in Fig. 3.

The tide-dominated Ord River Delta, described by Coleman and Wright (1979), is located on the coast of

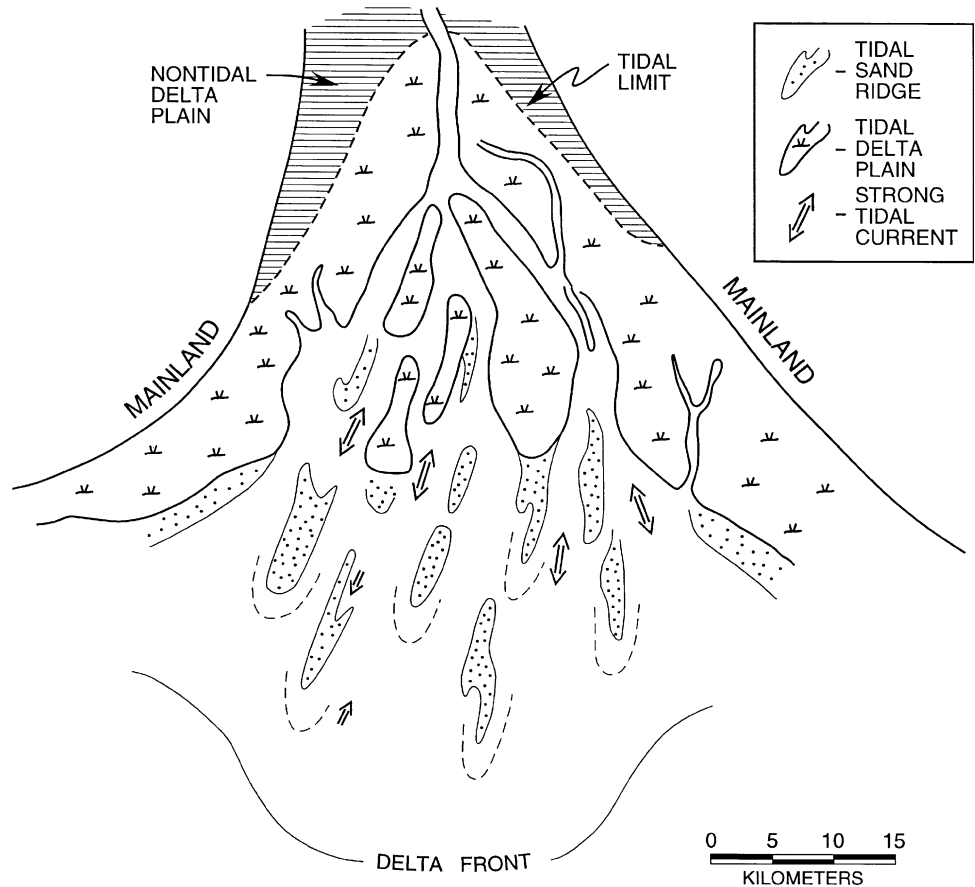
Tide-Dominated Coasts,

Fig. 2 Hypothetical regressive sequence for the intertidal zone of a prograding macrotidal (tide-dominated) shoreline in a non-deltaic setting



Tide-Dominated Coasts,

Fig. 3 Schematic sketch in plan view of a tide-dominated delta. Note the presence of tidal sand ridges at the offshore entrances to the delta complex



northwest Australia were the tidal range varies between 3.8 and 6.6 m. Tidal currents are oriented primarily in an onshore and offshore direction within an embayment which composes the major mouth of the Ord River. Linear sand ridges, which range in relief from 10 to 22 m and average 2 km in length, occur in the most seaward portion of the embayment. These sand ridges are formed and shaped primarily by tidal currents, which dominate over wave energy effects. A composite prograding stratigraphic column for this delta shows a fining upward sequence resulting from the progradation of muddy upper intertidal flats and marsh sediments over the sand bodies of the lower intertidal and shallow subtidal regions.

Tide-Dominated Estuaries

The concept of tide and wave dominance has also been applied to estuaries by Dalrymple et al. (1992). However, they state that estuaries are unlike many other coastal systems, because they are “geologically ephemeral.” If the rate of sediment supply is sufficient to eventually fill the lowstand valley within which the estuary is located, the filled valley then becomes a delta. The estuaries they term wave-dominated are composed of a sand body complex (barrier/tidal inlet) at the entrance, a muddy central basin, and a bayhead delta system. *Tide-dominated estuaries*, on the other hand, contains elongate sand bars and sandy tidal flats at the entrance and complex tidal channel/wetland habitats further inland.

Cross-References

- [Barrier Island Landforms](#)
- [Deltas](#)
- [Estuaries](#)
- [Tidal Environments](#)
- [Tides](#)
- [Wave-Tide-Dominated Coasts](#)
- [Wave-Dominated Coasts](#)

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Tides

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Introduction

Tides are the periodic motion of the waters of the sea caused by the changing gravitational effects of the moon and the sun as they change position relative to the rotating earth. The tides in the oceans are actually very long waves hundred or thousands of miles long. Although produced by astronomical forces, their behavior in the oceans and connected bays (and the size of resulting water level oscillations) is determined by hydrodynamics (i.e., the physics of the water movement).

The vertical rise and fall of the water surface is usually referred to as the *tide*, while the accompanying horizontal movement is referred to as the *tidal current*, with the tidal flow into a bay called the *flood* and the flow out of a bay called the *ebb*. For most areas of the earth the rise and fall, and flood and ebb, occur twice a day (referred to as a *semidiurnal* tide), but in some areas there may only be one *high water* and one *low water* per day (referred to as a *diurnal* tide). In many

areas, there are two high waters and two low waters per day, but one high water and/or one low water is a different height than the other (referred to as a *mixed* tide). The tide is only one phenomenon that produces variations in water level and currents. Such variations can also be caused by changes in the wind, atmospheric pressure, river discharge, and water density (due to changes in salinity and temperature), but they are not periodic like the astronomical tide and are not nearly as predictable, being associated with weather. Nontidal water level changes caused by changes in the wind and barometric pressure are usually referred to as *storm surges* (*q.v.*). The term *sea level* is generally used for longer-period, slower changes in water level. *Mean sea level* is the average of water level measurements over some time period (such as a day, a month, or a year), which averages out shorter-term oscillations like the tide.

Water level is the height of the water surface above some reference level, called a *datum*. A datum for a particular waterway is generally defined as an average height of a particular stage of the tide. For example, *chart datum* on a nautical chart in the United States is defined as the mean lower low water (MLLW) at each location. (*Lower low water* is the lower of the two low waters that occur each day, and MLLW is the average of all the lower low waters over some time period, usually at least a year). Depth soundings on a nautical chart are the depths below the chart datum, and the predicted tidal heights found in Tide Tables are the heights above the chart datum. Adding the two together gives the total water depth at that moment in time. These tidal datums also provide the legal definition of *marine boundaries*. MLLW, for example, is the dividing line between federal territorial seas and state submerged lands, and mean high water (MHW) is the dividing line between state tidelands and private uplands. Tidal datums at a particular tide gauge are referenced to the land through geodetic leveling to a number of *benchmarks*, which are brass markers driven into solid rock or other permanent structures. Tidal datums can change over decades if the land subsides (or rises due to glacial rebound) or if relative sea-level rises due to other effects.

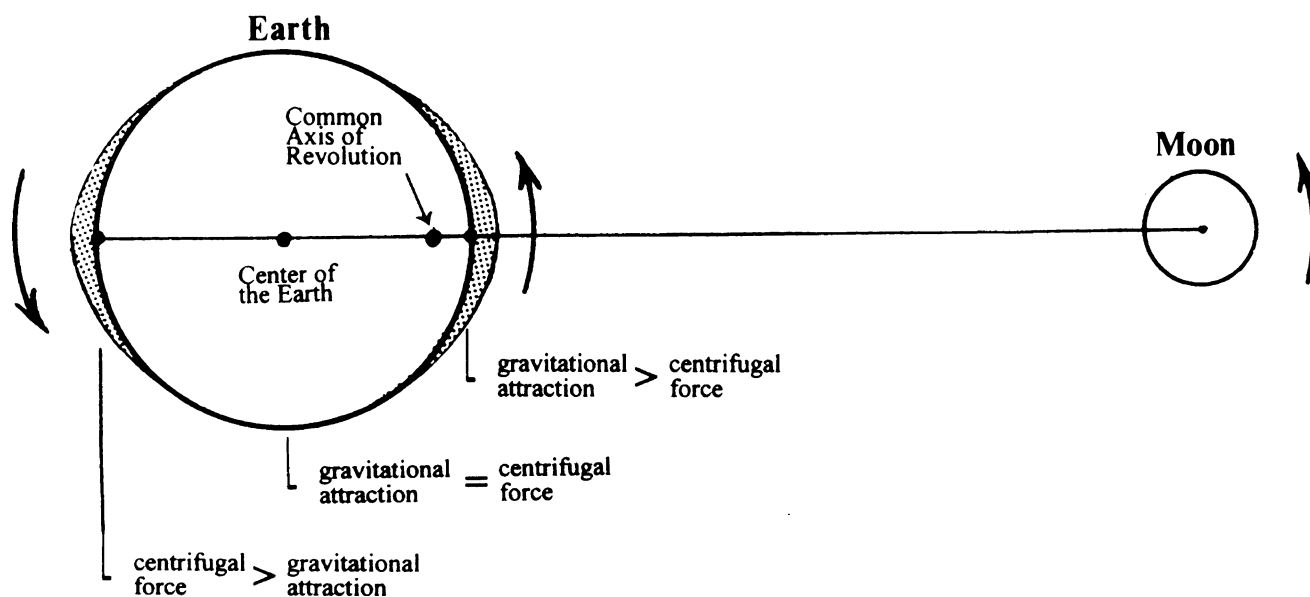
The tide dominates our thinking about changes in water level, not only because it usually causes the largest changes (except during storms), but also because it is very predictable (especially in comparison to how well we can predict the weather). After analyzing only a month's worth of water level measurements from a tide gauge, we can predict the tide quite accurately (for that location) for years into the future. This high predictability is due to the tide's periodic nature and our very precise knowledge of its astronomical forcing. The earth-moon orbit, the revolution of the earth around the sun, and the rotation of the earth on its axis involve periodic motions with fixed and precisely known time periods. Tidal energy is found at the same frequencies that describe these astronomical motions.

To fully understand and predict the tides one must understand both its astronomical forcing and the hydrodynamics of the oceans and bays. While it is the astronomical forcing of the tide that is the basis for the tide's predictability, it is the hydrodynamics of the tide that is responsible for the size of the *tidal range* (i.e., the height difference from low water to high water), the timing of high and low waters, and the *type of tide* (i.e., semidiurnal, mixed, or diurnal). It is the length, width, and depth of the bay or river (and of any adjoining waterways) that control the hydrodynamics. In shallow waterways, the hydrodynamics also transfers tidal energy to new frequencies, and distorts the shape of the tide curve away from perfect sine curves. These same shallow-water processes also lead to interactions between the tide and nontidal phenomena such as storm surge and river discharge.

The largest tidal ranges occur in shallow coastal waters, in particular, at the ends of certain bays and along coasts with very wide continental shelves. The increase in tidal range and tidal current speeds that one sees in the shallow waters of bays, rivers, and straits can go to dramatic extremes if the circumstances are right. Tidal ranges reach 15 m (50 ft) in Minas Basin in the Bay of Fundy. Tidal ranges greater than 12 m occur at the northern end of Cook Inlet near Anchorage in Alaska, in the Magellan Strait in Chile, in the Gulf of Cambay in India, along the Gulf of St. Malo portion of the French coast bordering the English Channel, in the Severn River in England, and along the open coast of southern Argentina. In a few rivers, a portion of the tide wave propagates up the river as a tumultuous wall of water, called a *tidal bore*. The largest tidal bores are found in the Tsientang River near Hanchow, China, and in the Amazon River, Brazil, where at certain times they can reach 7.5 m in height and travel up the river at a speed of 7 m/s. Smaller bores occur in the Meghna River in India, in the Peticodiac River at the end of the Bay of Fundy, in Turnagain Arm near Anchorage, and in the Severn River in England. Tidal current speeds greater than 7.5 m/s occur in Seymour Narrows, between Vancouver Island and the mainland of British Columbia, Canada. Tidal currents of 5 m/s are found in South Inian Pass in southeast Alaska and in Kanmon Strait, Japan. In some narrow or shallow straits, the tidal currents create dangerous whirlpools or *maelstroms*. Most famous is the whirlpool in the Strait of Messina (between Sicily and the southern tip of the Italian mainland), which Homer depicted in his *Odyssey* as the second of two monsters, Scylla and Charybdis, faced by Ulysses.

The Generation of Tides

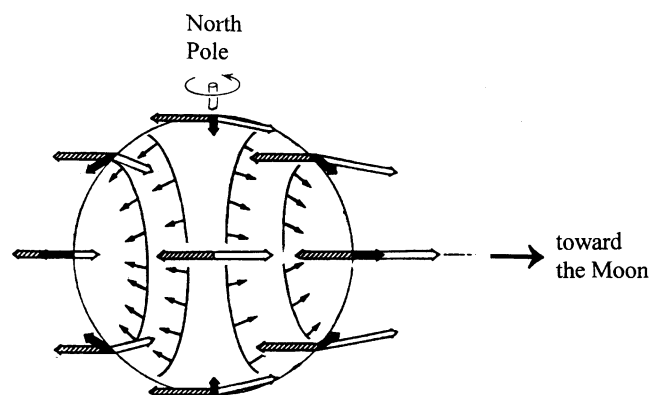
The tides are caused by both the moon and the sun, but the moon though smaller has roughly twice the effect because it is much closer to the earth than the sun. Although the moon



Tides, Fig. 1 The earth-moon system (viewed from above the North Pole) revolving around a common axis (inside the earth). The earth is shown with a hypothetical ocean covering the whole earth (with no

continents) and *two* bulges, resulting from the imbalances of gravitational and centrifugal forces

appears to orbit around the earth, the earth and moon both actually revolve around a common point, which, because the earth is much more massive than the moon, is inside the earth, but not at the earth's center (see Fig. 1). At the center of the earth there is a balance between the gravitational attraction (trying to pull the earth and moon together) and the centrifugal force (trying to push the earth and moon apart). At a location on the earth's surface closest to the moon, the gravitational attraction of the moon is greater than the centrifugal force. On the opposite side of the earth, facing away from the moon, the centrifugal force is greater than the moon's gravitational attraction. Figure 1 shows a hypothetical ocean (covering the whole earth with no continents) with *two* bulges, one facing the moon and one facing away from the moon, that result from the two imbalances of gravitational and centrifugal forces. However, if we look at the side of the earth facing the moon, the force vertically upward from the earth toward the moon is so small compared with the earth's gravitational force that it could not cause the bulges. If we move away from the equator to another point on the earth that is not directly under the moon, we see that the attractive force is still pointing toward the moon, but is no longer perfectly vertical relative to the earth (see Fig. 2). At this point, the force toward the moon can be separated in a vertical component of the force and a horizontal component, the latter one being parallel to the earth's surface. This horizontal force, though small, has nothing opposing it, and so it can move the water in the ocean. One can see from Fig. 2 that all the horizontal components shown tend to move the water into a bulge centered around



Tides, Fig. 2 The tide generating forces (the thick black arrows) on the earth resulting from the difference between gravitational attraction (the open arrows) and centrifugal force (the hatched arrows). The small black arrows are the *horizontal* components of the tide generating forces, which tend to move the water into the two bulges shown in Fig. 1

the point that is directly under the moon. Similarly, on the other side of the earth another bulge results.

One can easily envision the earth rotating under these bulges in this hypothetical ocean that covers the entire earth. In one complete rotation in 1 day there will be two high tides (when under a bulge) and two low tides (when halfway between bulges), and thus one entire tidal cycle would be completed in approximately half a day (actually 12.42 h, for reasons to be explained below). However, this is an extreme simplification (called the *equilibrium tide*) used merely to show how the tide generating forces change as the earth

rotates. Not only are the continents left out, but this assumes that the oceans respond instantly to the tide-generating force, which they do not.

Now consider the addition of continents and look at one of the oceans, with a bay connected to it. The tide-generating forces are too small to cause a tide directly in a small body of water like a bay. Only in a large ocean are the cumulative effects of the tide-generating forces throughout the ocean large enough to produce a tide. What is actually generated is a very long wave with a small amplitude, on the order of a half a meter or less (see Fig. 3). However, when this wave reaches the reduced depths of the continental shelf, there is a partial reflection of the wave, and the part of the wave that continues toward the coast is increased in amplitude. At the coast another reflection further increases the height of this long wave, now reaching at least a meter along most coasts. When the wave moves up into a bay there can be even more amplification depending on the depth, length, and width of the bay, with tidal ranges reaching 5, 10, or even 15 m for bays with the right dimensions.

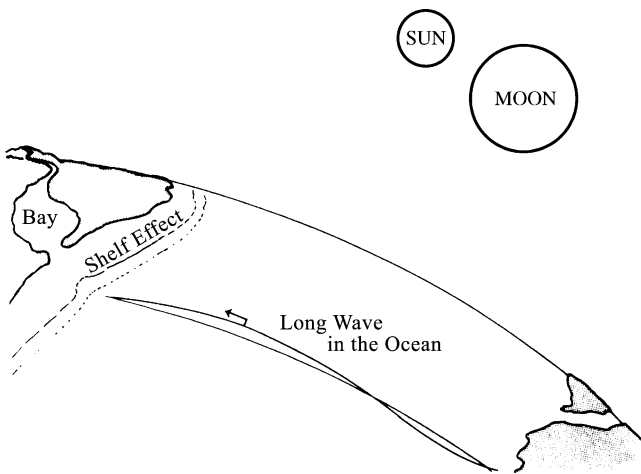
How large the tide range is depends on how close the natural period of free oscillation of the basin is to the period of the tide-generating force. If the natural period of the basin is same as the period of the tide-generating force, then the energy from the tidal forcing will be input in the same direction as the water is already moving and the resulting tide range will be larger. This is called *resonance*. The natural period of a basin, T_n , is approximately equal to $2L/(gD)^{1/2}$, where L is the length of the basin, D is the depth, and g is the acceleration due to gravity. The Atlantic Ocean is too wide for there to be resonance (its 19-h natural period being much longer than the 12.42-h tidal period). The

largest tide ranges in the world are in shallower basins with just the right length and depth combination to have natural periods close to the tidal period.

Astronomical Considerations

Because it is a forced oscillating system, the tide will oscillate with the periods determined by the relative motions of the earth, moon, and sun. There are many different periods involved due to the complex nature of the orbit of the moon around the earth and of the orbit of the earth around the sun. However, astronomers have very precisely determined all of these periods. To predict the tide at a specific location for any time in the future one must simply analyze water level data from that location to determine the amplitude and phase associated with each of the important tidal periods.

If the earth–moon orbit and the earth–sun orbit were circular and in the plane of the earth’s equator, there would only be two tidal frequencies, and only thus two semidiurnal constituents, M_2 and S_2 (defined below), would be needed to make tidal predictions. However, the orbital motions are more complicated. Distances between the moon and earth and between the earth and sun vary with time (the latter changes over a 20,942-year period and so is of no concern, except perhaps in paleoclimatology). Orbital planes are at angles relative to the earth’s equatorial plane and these angles also vary with time. All these motions modulate the tidal forces, so that tidal energy shows up at many more frequencies than at just M_2 and S_2 . The changing angles of the orbital planes also means that the moon and sun will not always be directly over the earth’s equator, thus making the two tidal bulges (on the opposite sides of the earth) asymmetric with respect to the axis of rotation, which introduces diurnal tidal frequencies. The fundamental periods in the motions of the earth, moon, and sun are shown in Table 1. Table 2 shows how



Tides, Fig. 3 The tide-generating forces caused by the moon and sun produce a very long wave of relatively small amplitude in the ocean. When this long wave reaches the continental shelf, then the coast, and finally propagates up a bay, it is amplified by an amount that depends on the length and depth of each of the basins

Tides, Table 1 Fundamental periods in the motions of the earth, moon, and sun. (After Parker et al. 1999)

Description	Period (mean solar units)	Frequency (1/period)
Sidereal day (one rotation wrt vernal equinox)	23.9344 h	Ω
Mean solar day (one rotation wrt to the sun)	24.0000 h	ω_s
Mean lunar day (one rotation wrt to the moon)	24.8412 h	ω_L
Period of lunar declination (tropical month)	27.3216 days	ω_1
Period of solar declination (tropical year)	365.2422 days	ω_2
Period of lunar perigee	8.847 years	ω_3
Period of lunar node	18.613 years	ω_4
Period of perihelion	20,940 years	ω_5

Tides, Table 2 Tidal constituents and their origins. (After Parker et al. 1999)

Symbol	Description	Period	Speed (/h)	Derived from	Coefficient C
Semidiurnal tides					
K_2^L	declinational to M_2	11.967 h	30.0821373	$2\omega_L + 2\omega_1 (=2\Omega)$	0.0768
K_2^S	declinational to S_2	11.967 h	30.0821373	$2\omega_S + 2\omega_2 (=2\Omega)$	0.0365
S_2^2	principal solar	12.000 h	30.0000000	$2\omega_S$	0.4299
M_2	principal lunar	12.421 h	28.9841042	$2\omega_L$	0.9081
N_2	elliptical to M_2	12.658 h	28.4397295	$2\omega_L - (\omega_1 - \omega_3)$	0.1739
L_2	elliptical to M_2	12.192 h	29.5284789	$2\omega_L + (\omega_1 - \omega_3)$	0.0257
Diurnal tides					
K_1^L	declinational to O_1	23.934 h	15.0410686	$(\omega_L - \omega_1) + 2\omega_1 (= \Omega)$	0.3623
K_1^S	declinational to P_1	23.934 h	15.0410686	$(\omega_S - \omega_2) + 2\omega_2 (= \Omega)$	0.1682
P_1^1	principal solar	24.066 h	14.9589314	$(\omega_S - \omega_2)$	0.1755
O_1	principal lunar	25.819 h	13.9430356	$(\omega_L - \omega_1)$	0.3769
Q_1	elliptical to O_1	26.868 h	13.3986609	$(\omega_L - \omega_1) - (\omega_1 - \omega_3)$	0.0722
Long-period tides					
Mf	declinational to M_0	13.661 days	1.0980331	$2\omega_1$	0.1564
Mm	elliptical to M_0	27.555 days	0.5443747	$(\omega_1 - \omega_3)$	0.0825
Ssa	declinational to S_0	182.621 days	0.0821373	$2\omega_2$	0.0729

The speed is another form of frequency. M_0 and S_0 represent constant lunar and constant solar forces. The coefficient, C, gives a global measure of each constituent's relative portion of the tidal potential (i.e., it ignores a latitudinal variation that is different for each species)

key tidal constituents are derived from these fundamental frequencies. (The *Doodson numbers* used in many tidal papers are a shorthand that indicates which of the six frequencies, ω_L , ω_1 through ω_5 from Table 1, are used to produce a particular constituent.) To make this a little clearer, we will describe the origins of a few of the more important tidal frequencies.

The moon orbits around the earth in the same approximate direction as the rotation of the earth, so that one lunar day (i.e., one complete rotation of the earth *with respect to the moon*) is 24.8412 h long ($1/\omega_L$ see Table 2). There are two tidal high water bulges on the earth, so the period of the largest semidiurnal lunar harmonic constituent, M_2 , is half a lunar day, or 12.4206 h.

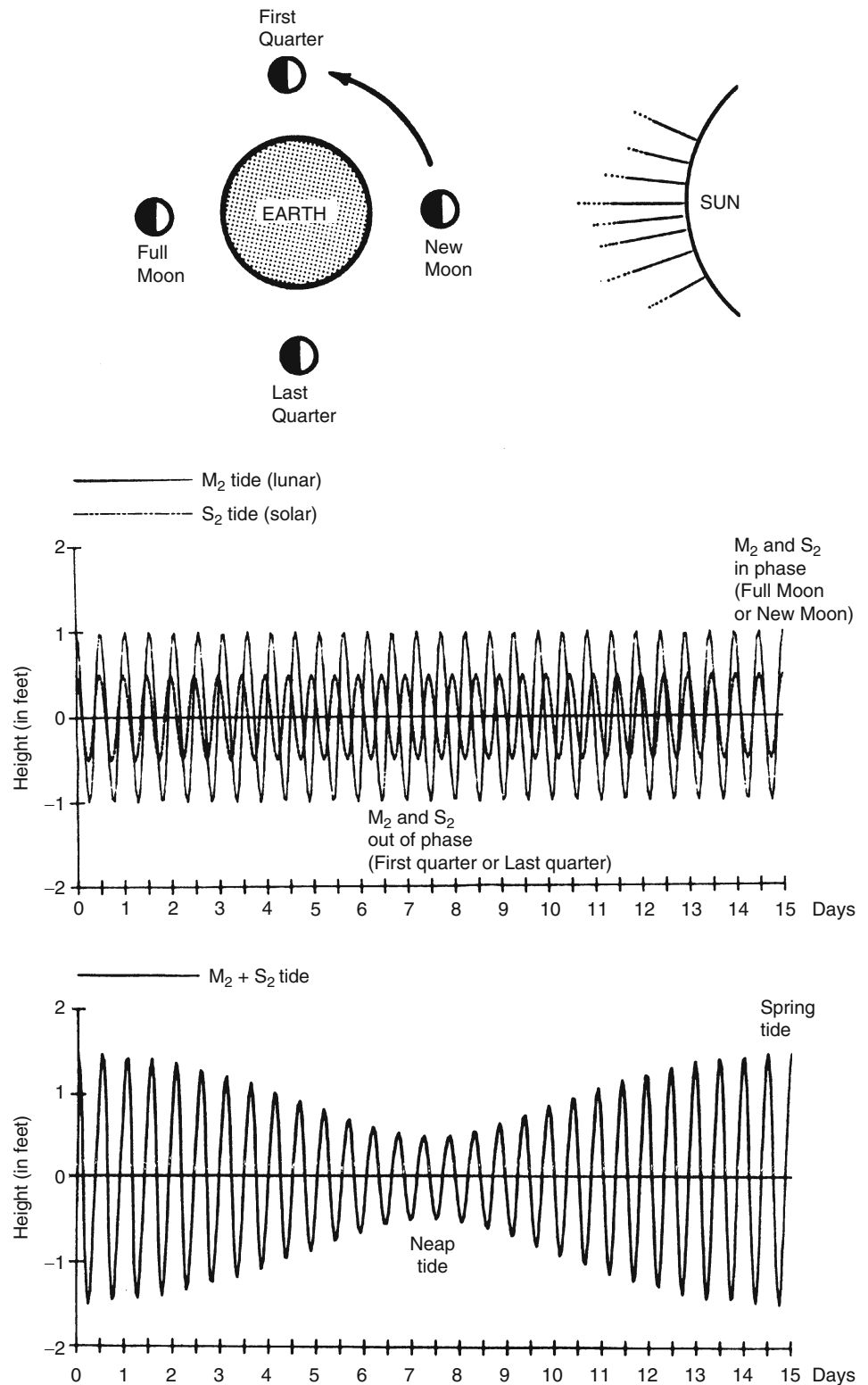
The earth turns under the sun exactly once every solar day, which leads to the main solar semidiurnal tidal constituent, S_2 , with a period of 12.0000 h. Because the sun is so much farther from earth than the moon (and the tidal force is inversely proportional to the cube of the distance), S_2 is much smaller in size than M_2 . When the moon and sun are in alignment (at new and full moons) their tidal forces work together to create *spring tides* with larger tidal ranges, while they work against each other at first and third quarters to create *neap tides* with smaller ranges (see Fig. 4).

The earth-moon orbit is elliptical, so that the distance between them varies over a 27.5546-day period ($1/[\omega_1 - \omega_3]$, see Table 2), from perigee (the moon closest to the earth, and so a stronger tidal force) to apogee (the moon farthest from the earth, and so a weaker tidal force) and back to perigee. This modulates the lunar tidal force. This modulation of M_2 shows up in a spectra (of water level or current data) as a line to the

left of the line for M_2 , this lower frequency line representing a second lunar harmonic constituent, N_2 , whose period is 12.6583 h. The stronger *perigean* tidal force will occur when M_2 and N_2 come into phase (leading to larger tidal ranges), while the weaker *apogean* tidal force will occur when M_2 and N_2 are exactly out of phase (leading to smaller tidal ranges). We can use Fig. 4 to illustrate this, if, in that figure, we replace S_2 with N_2 , spring tide with perigean tide, and neap tide with apogean tide. The difference is that with the M_2 plus S_2 case there really are two distinct effects being added, but in the case of the changing distance between the moon and earth, this directly varies the amplitude of the tide; and N_2 is merely a convenient way (in combination with M_2) to represent this variation of amplitude. There are several times a year when lunar perigee is reasonably close in time to new or full moon to produce the largest tidal ranges of the year, called *perigean spring tides*.

The plane of the moon's orbit around the earth is at an angle to the plane through the earth's equator. Thus, as the earth rotates under the moon, there will be times of the month when the moon is north of the equator (Northern Declination), over the equator (Equatorial Declination), and south of the equator (Southern Declination). When the moon is north or south of the equator, one of the tidal bulges is more north of the equator and one is more south, so that at a particular location on the earth there will either be only one high water per day (a diurnal tide), or, if there are two, they will be of different heights (the difference being the *diurnal inequality*) (see Fig. 5). The diurnal lunar tidal forces resulting from lunar declination are represented by two tidal constituents, O_1 and K_1 , with periods of 25.8193 and 23.9345 h, since they must

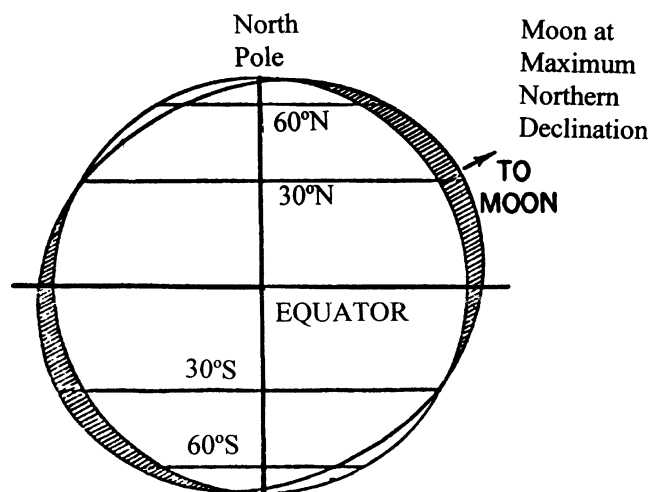
Tides, Fig. 4 The combined effect of the moon and sun varies throughout the month. When the moon and sun are working with each other (at full moon and new moon) one sees the largest tidal ranges (*spring tides*). At First Quarter and Last Quarter the moon and sun work against each other resulting in smaller tidal ranges (*neap tides*)



cancel each other out every 13.66 days ($1/2\omega_1$, see Table 2), at the times when the moon is over the equator. (The maximum angle between the plane of the moon's orbit and the earth's equator varies from 18.3° to 28.5° over a 18.6-year period; see below for further discussion of this *nodal cycle*.) The sum of

the O_1 and K frequencies is equal to the M_2 frequency, so that the time of the diurnal high water does not change with respect to the times of the two semidiurnal high waters.

The plane of the earth's orbit around the sun (called the *ecliptic*) is also at an angle to the plane through the earth's



Tides, Fig. 5 When the moon is at maximum declination north or south of the equator the tidal bulge also shifts north or south. When this happens certain locations on the earth would rotate under only one high water “bulge,” that is, there will be a diurnal component of the tidal forces. Hydrodynamics will determine which oceans and bays have significant diurnal tides

equator. Around December 21st the sun is furthest south of the equator (December solstice) and around June 21st it is furthest north of the equator (June solstice), the angle between the ecliptic and the equator reaching 23.5 in each case. December solstice marks the beginning of winter in the Northern Hemisphere and the beginning of summer in the Southern Hemisphere, and *vice versa* for June solstice. Around March 21st the sun is over the equator (vernal equinox) and again around September 21st (autumnal equinox). This movement of the sun north and south of the equator also leads to diurnal tidal constituents, in this case P_1 with a period of 24.0658 h ($1/(\omega_S - \omega_2)$, see Table 2), and another K_1 . Thus, the K_1 used for tidal prediction has both lunar and solar parts. P_1 and the solar part of K_1 cancel each other out every 182 days, at vernal and autumnal equinoxes.

As mentioned above, the angle between the plane through the moon's orbit and the plane through the equator varies over a 18.6-year period ($1/\omega_4$, see Table 2). This is referred to a *lunar nodal regression* because the intersection of the moon's orbital plane with the ecliptic, called the ascending lunar node, regresses backwards along the ecliptic over this 18.6-year period. Lunar distance also varies with time because the longitude of the lunar perigee rotates with an 8.85-year period ($1/\omega_3$ see Table 2). The spectral splitting due to these long-period effects can also be represented by harmonic constituents, called “satellite” constituents, but harmonic analyses using them must use data series that are 18.6 years long. Traditionally the nodal effects have been handled in a form that directly represents the modulation of each lunar tidal constituent. The amplitude of each modulation is called the *node factor*, f and the phase is called the *equilibrium*

argument, u . f and are regarded as constants for the period of analysis (or prediction) and are obtained from astronomically calculated tables (e.g., Tables 14 and 15 in Schureman 1958). The largest variation in f over a 18.6-year period is found in O_1 , which varies $\pm 18\%$, and in K_1 , which varies $\pm 11\%$. The variation of f for M_2 is 4%, but since M_2 is much larger than O_1 and K_1 in many locations, it may only be the $\pm 4\%$ variation with which one is usually concerned. There is no direct nodal effect on solar constituents such as S_2 or P_1 .

Tidal Analysis and Prediction

Knowing only the periods of these and other smaller tidal constituents, one can analyze a data series of water level observations from a particular location. The result of such a *harmonic analysis* is an *amplitude* and *phase* for each of these tidal constituents, which represent how large each effect is *at that particular location*, and when in time the peak of each effect will take place (such as relative to when the moon passes over (transits) that location). The hydrodynamics affect both the amplitude and phase (timing) of each tidal constituent, but we really do not need to know the details of how it happened, only that it did happen. Hydrodynamics in shallow-water areas will also produce additional tidal constituents not seen in deeper water (discussed in the next section) that can also be easily handled by the harmonic method. As long as the hydrodynamics stay the same (i.e., the depth of the basin stays the same), the tide predictions from these calculated harmonic constants will be accurate. For small bays, shoaling or dredging can change the hydrodynamics and thus change the tide range and times of high and low waters.

Harmonic analysis is still the most common method used for the analysis and prediction of tides and tidal currents. It has not changed much from the theory first developed by Lord Kelvin in 1867 and later refined by George Darwin, except that the Fourier analysis solution technique (e.g., Schureman 1958) has generally been replaced by a least-squares solution technique (e.g., Foreman and Henry 1989) that minimizes the squared differences between measurements and computed tidal predictions. Doodson (1921) carried out a complete harmonic development of the tide-generating potential, determining over 400 tidal constituents (most very small), which has been the standard reference work for tidal analysis and prediction. This was recalculated and updated 50 years later by Cartwright and Tayler (1971).

Harmonic analysis is used to produce the predictions at the reference stations in all nationally published Tide and Tidal Current Tables. Nonharmonic analysis, which is simply one of several methods of comparing the tide or tidal current at two locations, is used to calculate the time differences, height differences, velocity ratios, etc. for the thousands of secondary stations in these tables that are referred to the reference

stations. Several refinements to the harmonic methods have been developed in recent decades, as well as the nonharmonic/cross-spectral approach of the response method developed by Munk and Cartwright (1966). Whatever the method used, some principles will always apply. For example, the analysis of longer data time series will lead to more accurate predictions because more tidal frequencies can be resolved in a longer series.

Using the harmonic method the tide is represented by the sum of various tidal constituents, each representing one of these frequencies. The height of the tide at any time is typically represented by a formula such as (Schureman 1958):

$$h = H_0 + \sum f H \cos [at + (V_0 + u) - \kappa],$$

where

h = height of tide at any time t

H_0 = mean height of water level above the datum used for prediction

H = mean amplitude of any constituent A

f = node factor

a = speed of constituent A (i.e., its frequency)

t = time, reckoned from some initial epoch (such as the beginning of the year of predictions)

$(V_0 + u)$ = value of equilibrium argument of constituent A at $t = 0$.

K = epoch (phase) of constituent A.

The “speed” of a constituent is merely its frequency, but it has been traditionally given in terms of degrees per solar hour, where 360° is one complete cycle. Thus, M_2 , which has a period of 12.42 h and a frequency of 1.932 cycles per day, has a speed of $28.984104^\circ/\text{h}$.

For tidal current predictions the same equation is used twice, once for each of two orthogonal components (e.g., major and minor axes of flow), which when combined will give the speed and direction of the flow. The major and minor component for each tidal constituent can be combined to produce a *tidal constituent ellipse*, which shows what the constituent flow would be for each instant in a constituent cycle.

The more tidal constituents that can be calculated the more accurate the tidal prediction will be. The number of constituents that can be calculated depends on the length of the data series. The length of a data series needed to resolve two tidal constituents is inversely proportional to the difference in the frequencies. Table 3 lists the 37 typically most important tidal constituents, listed in order of length of time needed to resolve that constituent from a nearby larger constituent. One sees natural groupings near 15 days, 29 days, 6 months, and 1 year.

If one has only 15 days of data, for example, then a harmonic analysis will provide values for the major constituents M_2 , S_2 , K_1 , and O_1 plus a few higher harmonics and a couple of less important constituents. However, these calculated values will also include energy from the constituents that could not be separated out in only 15 days. For example, the M_2 will include energy from N_2 (which could have been resolved from M_2 if there had been 29 days of data). This N_2 contribution could make the M_2 value calculated from 15 days of data larger than it should be, or smaller than it should be, depending on when the data were measured. Likewise, K_1 will include the effects of P_1 (which could have been resolved from K_1 if there were 6 months of data). If we harmonically analyzed successive 15-day periods, we would see the amplitude of K_1 slowly vary over a 6-month period because of the influence of P_1 . The key to an accurate tidal prediction is determining amplitudes and phases for the most tidal constituents that can be calculated with a given length data time series.

To carry out a tide prediction in the era before the above equation could be programmed on a computer (as it is done routinely today), large machines were built with gears and pulleys connected by a wire to a pen. Each tidal constituent had a different size rotating gear and a pin and yoke system connected to a pulley (see Fig. 6). The pin and yoke system turned the rotating motion of the gear into a vertical up and down motion of the pulley, which moved the wire over it and thus moved the pen up and down on a roll of moving paper. The wire ran over a number of pulleys so all the constituent effects could be added together. The first tide predicting machine was a wooden model built for Kelvin in 1872, but later models were huge brass machines with dozens of finely made gears and pulleys. The first one built in the United States was by William Ferrel in 1885. Prior to the use of harmonic analysis there were other less sophisticated methods based on recognized relationships between the tides and the movements of the moon and sun. For example, for a particular place, high tide might occur a certain number of hours after the moon was directly overhead, and the highest (spring) tide might occur a certain number of days after full moon or after new moon. In many of the early maritime nations, tide prediction schemes were treasured family secrets passed on to the next generation. The two earliest tide tables discovered were for the tidal bore in the Tsientang River in China in 1056 and for London Bridge in England in the early 1200s.

Hydrodynamic Considerations in Coastal Waters

When the very long tide wave generated in the ocean reaches the shallower water of the continental shelf and the even shallower water of the bays and rivers, it is slowed up, amplified, modulated, and distorted by a number of

Tides, Table 3 Thirty-seven tidal constituents that can be calculated from a one-year series, listed in order of length of data time series needed to resolve each constituent from a nearby larger constituent, and size

Constant	Speed (°/h)	Origin	Days needed to separate	From	Amplitude at Trenton, NJ (feet)
M_2	28.984104	Lunar	—	—	3.547
M_4	57.968208	Shallow-water	0.5	M_2	0.517
M_6	86.952313	Shallow-water	0.5	M_4	0.266
M_8	115.936417	Shallow-water	0.5	M_6	0.120
K_1	15.041069	Lunisolar	1.1	M_2	0.349
S_6	90.000000	Shallow-water	4.9	M_6	0.005
S_4	60.000000	Shallow-water	7.4	M_4	0.005
O_1	13.943036	Lunar	13.7	K_1	0.288
MK_3	44.025173	Shallow-water	13.7	$2MK_3$	0.120
$2MK_3$	42.927140	Shallow-water	13.7	MK_3	0.116
OO_1	16.139102	Lunar	13.7	K_1	0.030
$2Q_1$	12.854286	Lunar	13.8	O_1	0.028
S_2	30.000000	Solar	14.8	M_2	0.461
MS_4	58.984104	Shallow-water	14.8	M_4	0.148
$2SM_2$	31.015896	Shallow-water	14.8	S_2	0.025
M_3	43.476156	Lunar	27.3	MK_3	0.034
M_1	14.492052	Lunar	27.3	K_1	0.027
N_2	28.439730	Lunar	27.6	M_2	0.553
MN_4	57.423834	Shallow-water	27.6	M_4	0.176
Mm	0.544375	Lunar ^a	27.6	Mm	0.124
Q_1	13.398661	Lunar	27.6	O_1	0.022
J_1	15.585443	Lunar	27.6	K_1	0.016
$2MN_2/L_2$	29.528479	Shallow-water/lunar	31.8	S_2	0.409
$2MS_2/\mu_2$	27.968208	Shallow-water/lunar	31.8	N_2	0.219
MSf	1.015896	Lunar ^a	182.6	Mf	0.186
Mf	1.098033	Lunar ^a	182.6	MSf	0.132
P_1	14.958931	Solar	182.6	K_1	0.110
K_2	30.082137	Lunisolar	182.6	S_2	0.094
ν_2	28.512583	Lunar	205.9	N_2	0.210
λ_2	29.455625	Lunar	205.9	$2MN_2$	0.102
$2NM_2/2N_2$	27.895355	Shallow-water/lunar	205.9	$2MS_2$	0.042
ρ_1	13.471514	Lunar	205.9	Q_1	0.013
Sa	0.041069	Solar ^a	365.2	Ssa	0.430
Ssa	0.082137	Solar	365.2	Sa	0.169
S_1	15.000000	Solar	365.2	K_1	0.062
T_2	29.958933	Solar	365.3	S_2	0.056
R_2	30.041067	Solar	365.3	S_2	0.028

Amplitudes from a 1981 analysis of water level data (from Trenton, NJ, on the Delaware River) are provided as an example (after Parker et al. 1999)

^aValues are determined predominantly by long-term meteorological effects and thus vary from year to year

hydrodynamic mechanisms. The long wave enters and propagates up a river as a *progressive wave*, that is, the crest of the wave (high water) moves progressively up the river, as does the trough of the wave (low water) (see Fig. 7). In such a progressive tide wave the flood current (the tidal current flowing up the river) is fastest at approximately the same time as high water, and the ebb current (the tidal current flowing down the river) is fastest at approximately the same time as low water. Slack water (the time of no current) occurs approximately halfway between the times of high water and low water.

If there is nothing in a river to impede or stop the tide wave (like a dam or rapids or a sudden decrease in width), it will continue to travel up the river until bottom friction wears it down. However, if the width of the river decreases as the tide wave moves upriver, then the tidal range will be increased, because the same energy is being forced through a smaller opening. If the depth of the river decreases there is a similar but less dramatic amplifying effect generally outweighed by the increased frictional energy loss.

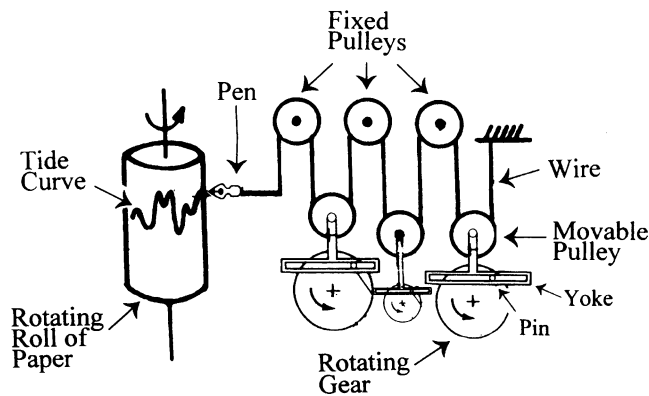
The greatest amplification of a tide wave usually occurs in a bay (or in a river with a dam). In this case, the tide wave is

reflected at the head of the bay and travels back down the waterway toward the ocean. This *reflected* wave is not observable by someone on the shore because it is superimposed on the next incoming tide wave that is propagating up the bay, and it is the combination of the two waves that one observes. The resulting combined wave is called a *standing wave*, because the high and low waters do *not* progress up the bay or river. The water simply moves up and down everywhere at the same time (see Fig. 8), with the greatest tidal range at the head of the bay. With a standing wave, the tidal range decreases as one moves from the head of the bay toward the ocean entrance, and, if the bay is long enough, reaches a minimum at one location (called a *node*) and then starts increasing again. This node occurs at one-fourth of a tidal wavelength from the head of the bay (see Fig. 8). (In a progressive wave high water comes half a wavelength before low water, so if a high water travels a distance equal to one-fourth of a tidal wavelength up the bay to the head, where it is reflected, and then travels one-fourth of a wavelength back

down the bay, it will have gone half a wavelength and so coincide with low water of the next incoming wave, and the two will cancel each other out at that location, producing a very small tidal range.) For a standing wave high waters occur at the same time everywhere on one side of the node, which is the same time as low waters occur on the other side of the node. The strongest tidal currents occur when water level is near mean tide level, approximately halfway between the times of high water and low water. At the times of high water and low water there is no flow (slack water). The water flows into the bay, stopping the inward flow at high water, reverses direction, flows out of the bay until low water, at which time it reverses again and starts flowing into the bay again.

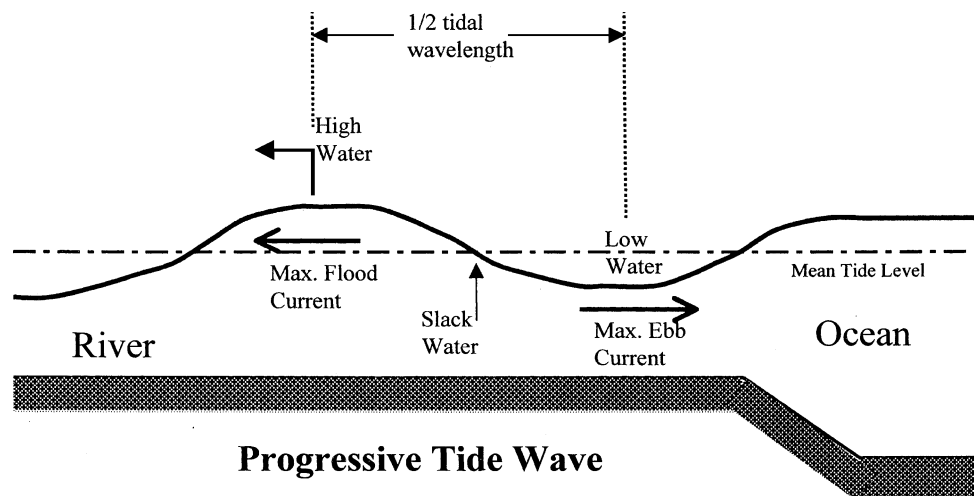
When length of a bay is exactly one-fourth of a tidal wavelength, then resonance occurs, which creates the largest tides possible. When the water in the bay is forced to move up and down by the tide at the entrance, it will freely oscillate (slosh up and down) with a natural period that depends directly on its length and inversely on the square root of its depth. If the basin has the right combination of length and depth so that the natural period is the same as the tidal period, then the oscillation inside the bay will be synchronized with the oscillation at the entrance due to the ocean tide. In other words, the next ocean tide will be raising the water level in the bay at the same time that it would already be rising due to its natural oscillation (stimulated by the previous ocean tide wave), so that both are working together, thus making the tidal range inside higher.

Most bays actually fall in between the extremes of pure progressive wave and pure standing wave described above, because bottom friction reduces the amplitude of the tide wave as it travels. Thus, the reflected wave will always be smaller than the incoming wave, especially near the bay entrance, and the combination of the two frictionally damped progressive waves will not be a pure standing wave. There

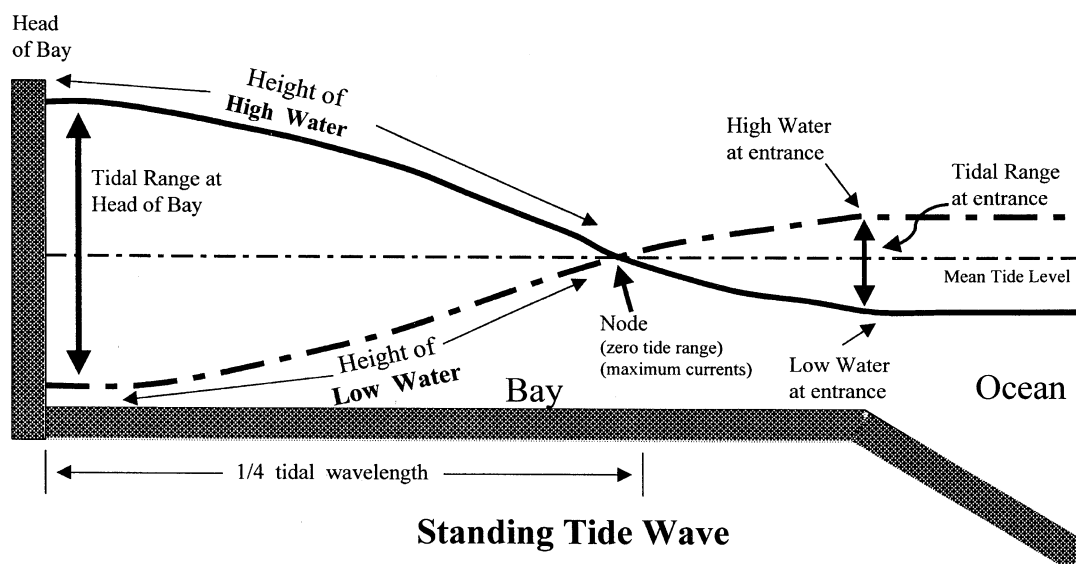


Tides, Fig. 6 A schematic of an early tide prediction machine. Each gear and pulley combination represented one tidal constituent. The wire running over every pulley summed the motions and moved a pen on a moving roll of paper to draw a tide curve

Tides, Fig. 7 The tide propagating up a river as a *progressive wave*. High water occurs later as one moves upstream. The tidal wavelength is typically on the order of hundreds of miles



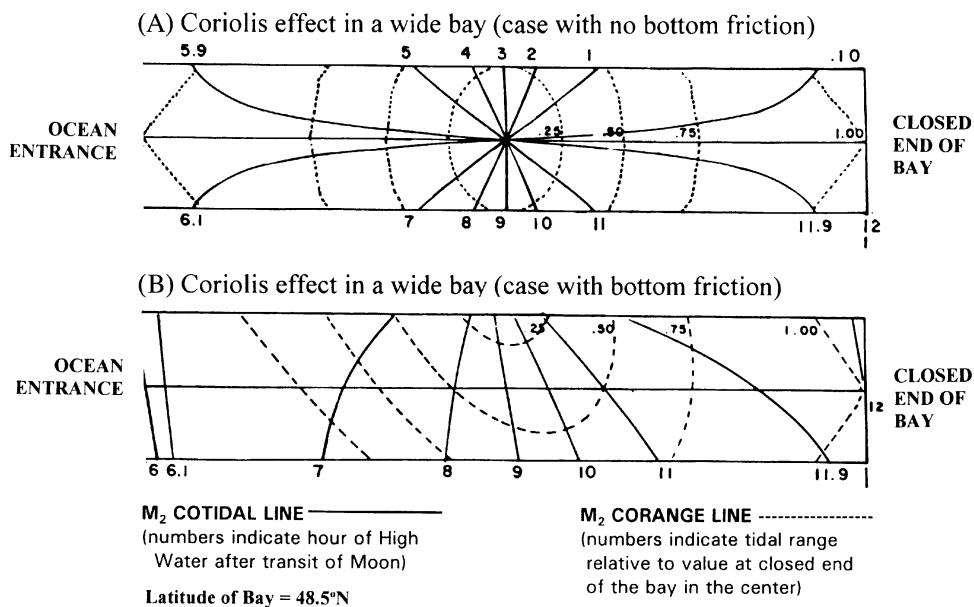
Progressive Tide Wave



Tides, Fig. 8 The tide in a bay as a *standing wave* (the water level is shown for two extremes, high water and low water). High water occurs at approximately the same time everywhere on one side of the node (the

point of zero tidal range). This is an idealized case assuming there is no bottom friction effect. With friction the tidal range at the node is not zero and the times of high water do progress slightly up the bay

Tides, Fig. 9 The effect of Coriolis force on M_2 tidal range (corange lines) and the times of high water (cotidal lines) for an idealized rectangular bay. The top figure assumes no bottom friction effects, and a single point of zero tidal range in the middle of the bay results. The bottom figure, which includes friction, is more realistic, and there is no point with zero tidal range

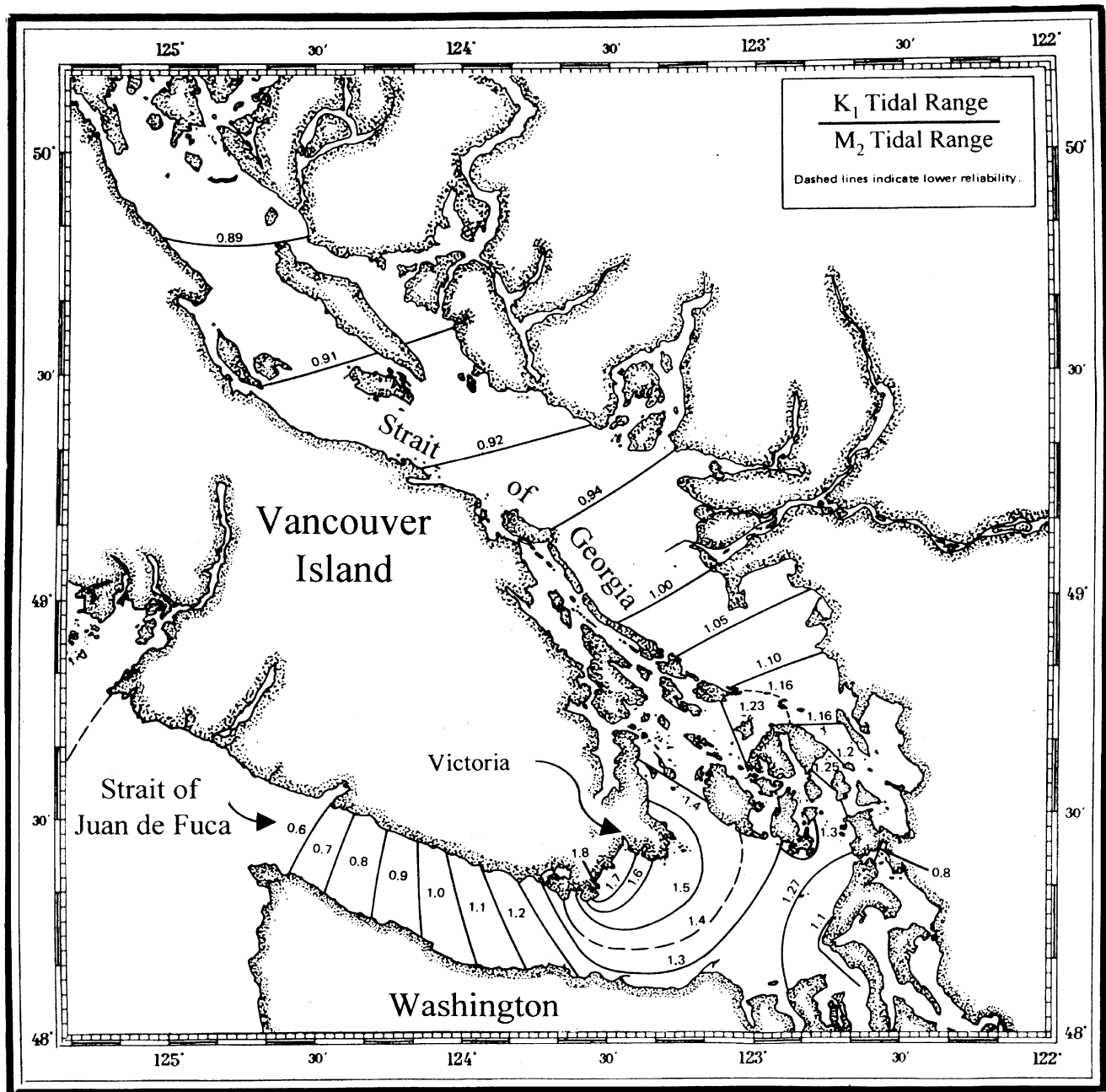


will be no point of zero tidal range, but only an area of minimum tidal range. There will be some progression of high waters and low waters up the bay, and maximum flood or ebb currents will not occur exactly half way between high water and low water. A basin one-fourth of a wavelength long will still produce the largest possible tidal range at the head of the bay, but friction keeps that tidal range much smaller than it would be without friction.

In some bays, the very high tidal range at the head of the bay is due to a combination of both a narrowing width and a near resonant situation (due to the right length and depth). The highest tidal ranges may involve several amplifications, the

bay being perhaps connected to a gulf with perhaps a wide continental shelf beyond that, with amplifications of the tide wave occurring in each basin. This is the case with the Bay of Fundy tides, the tide wave being already amplified by the Gulf of Maine and the continental shelf prior to entering the Bay of Fundy.

If a bay is wide enough one also sees larger tidal ranges on the right side of the bay (looking up the bay) due to the Coriolis effect. Figure 9a shows lines of locations with the same time of high water (cotidal lines) and lines of constant tidal range in an idealized rectangular basin for the case where the effect of bottom friction is ignored. A single point of zero



Tides, Fig. 10 The ratio of the largest diurnal component of the tide (K_1) to the largest semidiurnal component of the tide (M_2) along the length of the Strait of Georgia-Strait of Juan de Fuca. The highest ratio of diurnal range to semidiurnal range occurs near Victoria, British

Columbia, because that is the area of the semidiurnal node (minimum M_2 tidal range), which is one-fourth of a semidiurnal tidal wavelength from the northern end of the Strait of Georgia (see text)

tidal range occurs in the center of the bay. A more realistic case, including the damping effect of bottom friction, is shown in Fig. 9b, with the point of no tidal range disappearing onto land. (The effect of Coriolis can also be seen in Fig. 10, where the shape of the M_2/K_1 lines near Victoria are caused by the point of zero M_2 tidal range having moved to the left on land.)

Huge tidal ranges are not restricted to bays. If the continental shelf is the right combination of depth and width, a near resonant situation can also result. This is the reason for the 12 m tidal ranges along the coast of southern Argentina. The continental shelf there is over 1,000 km (600 miles) wide, and includes the Falkland Islands near the edge of the shelf (where the tidal range only reaches 2 m. The distance from the

Argentinean coast to the edge of the shelf is fairly close to one-fourth of a tidal wavelength for that depth of water. Essentially, that wide shelf has a natural period of oscillation that is fairly close to the tidal period.

The largest tidal currents in bays tend to be near the entrances. Maximum tidal current speeds are zero at the head of the bay (since there is no place for the water to flow). As one moves down the bay toward the ocean the maximum flood and ebb tidal current speeds increase, with the greatest speeds occurring at the entrance, or, if the bay is long enough, at the area of smallest tidal range (the nodal area). However, if the width of the bay decreases at any point, the current speeds will be increased in that narrow region (since the same volume of water is being forced to flow through a smaller cross section, it must flow faster). This can be especially dramatic if there is a sudden decrease in width and depth. The largest tidal currents are found in narrow straits in which the tides at either end have different ranges or times of high water. Where a strait suddenly becomes very narrow or where it bends, eddies and whirlpools can be formed as the result of the sheltering effect of the land and the inertia of the coastal flow.

The dimensions of a basin can also determine the size of the diurnal tidal signal compared with the usually dominant semidiurnal tidal signal. A particular bay could have a natural period of oscillation that is closer to the diurnal tidal period (approximately 24.84 h) than to the semidiurnal period, thus amplifying the diurnal forcing at the entrance to the bay more than the semidiurnal signal. Depending on the size of the diurnal signal at the entrance the result could be a mixed tide or a diurnal tide. At such locations (such as parts of the Gulf of Mexico) the tide will be diurnal near times of maximum lunar declination, but will be semidiurnal near times when the moon is over the equator.

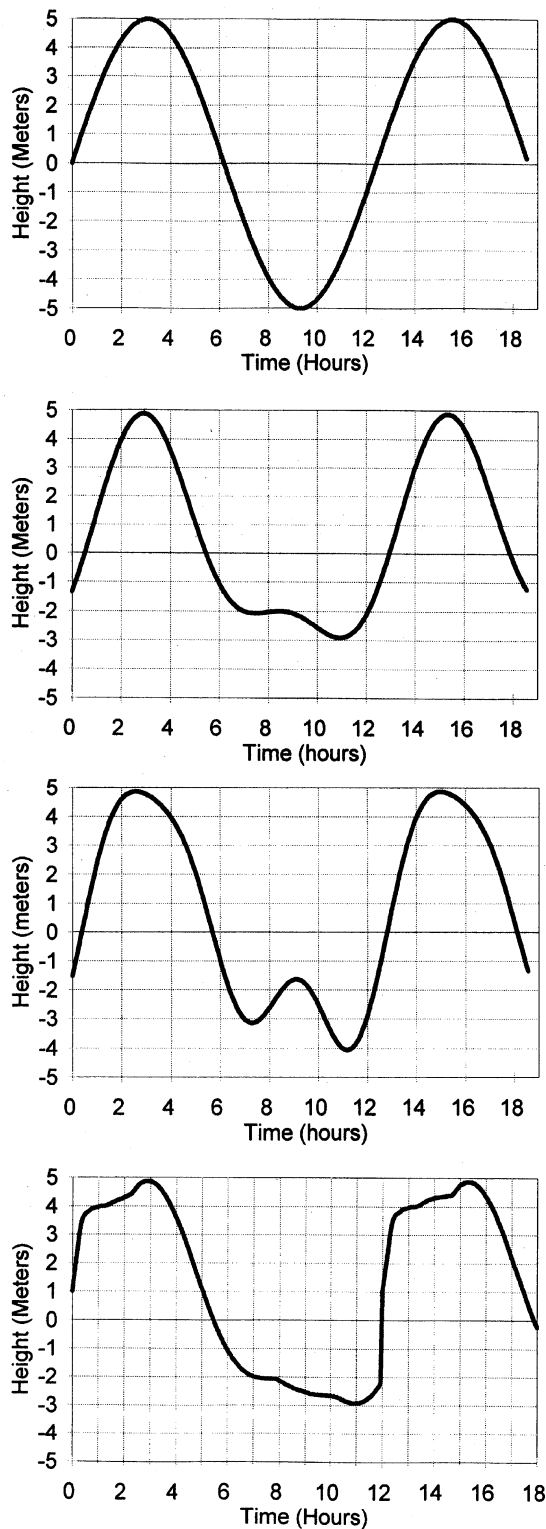
The wavelength, L , of a tide wave in a bay depends on the depth of the water, D , and on the tidal period, T , according to $L = T(gD)^{1/2}$ (if we ignore frictional effects). The shallower the bay the shorter the wavelength. The longer the tidal period the longer the wavelength. A diurnal tidal component has a wavelength twice as long as a semidiurnal tidal component since its period is twice as long. When a waterway is shallow enough and long enough so that more than one-fourth of a semidiurnal wavelength fits in the waterway, there will be a nodal area with a very small semidiurnal tidal range. This will be an area where the diurnal tide could dominate, since the diurnal tide would still be large at the semidiurnal nodal area (its the diurnal node being twice as far from the head of the bay). Thus near the head of the waterway the tide could be semidiurnal, but near the semidiurnal nodal area the tide could be mixed or even diurnal. This is the case near Victoria, British Columbia, at the southeastern end of Vancouver Island (see Fig. 10). At that location along the Strait of Georgia-

Strait of Juan de Fuca waterway, the semidiurnal tidal component decreases to a minimum, but the diurnal component does not, and so the tide becomes diurnal (while at the northern end of the Strait of Georgia the tide is semidiurnal).

Whether due to a basin size conducive to amplifying the diurnal signal or due to the existence of a semidiurnal nodal area (leaving the diurnal signal as the dominant one), there are numerous areas around the world with strong diurnal tides – places like Norton Sound in Alaska near the Bering Strait, and various (but not all) locations in the Philippines, New Guinea, and the islands of Indonesia. In southern China, at Beihai, and at Do Son, Vietnam, the diurnal signal is very dominant, with tidal ranges that reach 5 m and 3 m, respectively (near times of maximum southern declination of the moon); the tide remains diurnal even near times when the moon is over the equator.

The primary effect of shallow water on the tide that we have discussed so far is that it shortens the tidal wavelength down to the same order of magnitude as the lengths of bays and river basins, thus bringing the dynamic situation closer to resonance and increasing the tidal ranges. (Or, one can also look at it from the point of view of the shallower depths increasing the natural periods of these bays and rivers (which are very small basins compared to the ocean) to be closer to the tidal period.) However, very shallow water can have other effects on the tide, for example, distorting the shape of the tide wave, that is, making it very asymmetric, so that its rise and fall (and its flood and ebb) are no longer equal (see the second curve in Fig. 11). The tide can then no longer be described by a simple sine wave (the first curve in Fig. 11). In some cases such distortion leads to double high waters or double low waters (see third curve in Fig. 11). The extreme case of distortion is a tidal bore (the fourth curve in Fig. 11).

Shallow water distorts the tide through several mechanisms. The speed, C , at which a long tide wave travels depends on the depth of the water, D , according to the formula $C = (gD)^{1/2}$. When depth of the water is much greater than the tidal range, the speed of the crest of a tide wave and the speed of the trough are virtually the same, since the tide wave itself has only a very small effect on the total water depth. But in the shallow water where the depth is not much greater than the tidal range, the total water depth under the crest is significantly larger than the total water depth under the trough. In this case, the crest of the wave (high water) travels faster than the trough of the wave (low water). If the tide wave travels far enough the crest begins to catch up with the trough ahead of it (which is falling behind the crest ahead of it). The shape of the tide wave then begins to look like the second curve in Fig. 11, with a more rapid rise to high water and a slower fall to low water. In terms of harmonic constituents, this distortion transfers energy from M_2 into a constituent called M_4 , with half the period.



Tides, Fig. 11 Typical tide curves (i.e., tidal height plotted with respect to time), over one-and-a-half tidal cycles for an area with no shallow-water effects (top panel) and for three areas with different degrees of distortion caused by the shallow water. In the third panel a double low water occurs. The fourth panel shows the almost instantaneous rise in water level due to the passage of a tidal bore

Another shallow-water distorting mechanism is caused by bottom friction, which can have both asymmetric and symmetric effects. The asymmetric effect (similar to that just discussed and represented in Fig. 11) results because friction has a greater effect in shallow water than in deep-water (there being less water to have to slow down), and so it slows down the trough more than the crest, contributing to the distortion of the tide wave and the generation of M_4 . A symmetric effect results because energy loss due to friction is proportional to the square of the current speed. This means that there will be much more energy loss during times of maximum flood and maximum ebb than near slacks. This results in the generation of another higher harmonic, M_6 with a period of one-third that of M_2 . This effect, combined with the asymmetric effect, can lead to double high or low waters (see third curve in Fig. 11). Higher harmonic tidal constituents like M_4 and M_6 are referred to as *overtides*.

The above symmetric frictional effect also causes the interaction of two tidal constituents, for example, M_2 and N_2 . M_2 and N_2 go in and out of phase over a 27.6-day cycle (perigee to apogee to perigee). In this case, the greatest energy loss occurs when M_2 and N_2 are in phase and producing the strongest tidal currents, and the lowest energy loss occurs 13.8 days later with M_2 and N_2 are out of phase and producing the weakest tidal currents. The increased energy loss when M_2 and N_2 are in phase is greater than the decreased energy loss when they are out of phase, and the result is that each constituent will be smaller than if it existed without the other present. There is a 27.6-day modulation of this energy loss from M_2 and N_2 , which produces two new *compound* tidal constituents called $2MN_2$ and $2NM_2$. (Similarly, the above asymmetric mechanisms also cause interactions between constituents, producing higher frequency constituents such as MN_4 from M_2 and N_2 .)

Friction, of course, dissipates energy from the entire wave and slowly wears the entire wave down. However, if, as the wave propagates up the river, the river's width is decreasing significantly, this can keep the amplitude of the wave high in spite of the friction. Thus, the tide wave can continue to travel up a narrowing river, getting more and more distorted in shape. A further distortion can be caused by the river flow interacting with the tide (see below). In the extreme case, the distortion from all these effects can lead to the creation of a tidal bore (see fourth curve in Fig. 11).

In a river there will also be the river current itself (resulting from fresh water flowing downhill) added onto the tidal current, the result being a faster and longer lasting ebb current and a slower shorter flood current phase. Far up a river where the river flow is faster than the strongest tidal current, the flow of water will always be downstream, but the speed of flow will oscillate, flowing the fastest downstream at the time of

maximum ebb for the tidal current and flowing the slowest downstream at the time of maximum flood for the tidal current. This is a simple addition to the tide, but the river flow also interacts with the tide and distorts it through interaction caused by bottom friction. As just mentioned, energy loss due to friction is proportional to the square of the total current speed. When the river current, flowing in the same direction as the ebb current, creates a larger combined ebb current, there is a greatly increased energy loss. Likewise, when the river current flowing opposite to the flood current reduces the total speed, the energy loss is greatly reduced. This not only has an asymmetric effect that distorts the tide (causing a faster rise to high water, delaying the time of low water, and contributing to M_4), but it also further wears down the entire wave because the increased energy loss during ebb is larger than the decreased energy loss during flood.

Another type of shallow-water effect causes interactions between tide and storm surge (generated by the wind) that have periods longer than tidal periods. In this case, when the water level is raised by an onshore wind, that increases the water depth and changes the tidal dynamics, usually increasing the tidal range. When an offshore wind lowers the water level, decreasing the water depth, the result is usually a decreased tidal range. Knowing that river discharge and storm surge can modify the tide, it is important when harmonically analyzing water level data to make sure that these data were not taken only during such meteorological events. These shallow-water effects that distort and modulate the tide and cause interactions with storm surge and river discharge are called *nonlinear* effects because the mechanisms that produce these effects are represented by several *nonlinear* terms in the equations of motions used to model the tidal hydrodynamics.

We have shown how tidal prediction can be accomplished quite accurately by harmonic analysis of a water level data time series knowing only the astronomical frequencies involved, and ignoring hydrodynamics. Accurate inclusion of the additional tidal constituents produced by nonlinear shallow-water effects into the harmonic prediction method does involve some knowledge of hydrodynamics (especially when deciding the formula for each constituent's node factor). However, recently tidal prediction has begun to involve the use of sophisticated hydrodynamic numerical models, which have two important advantages over statistical techniques. First, such models can provide tide and tidal current predictions at locations where there are no water level data, thus providing predictions at hundreds or thousands of locations in an area, as opposed to the few locations where statistical methods can provide them (limited by needing data to analyze). Second, such models handle very nicely the nonlinear interaction between the tide and nontidal phenomena like storm surge and river discharge.

Cross-References

- [Altimeter Surveys, Coastal Tides, and Shelf Circulation](#)
- [Coastal Currents](#)
- [Meteorological Effects on Coasts](#)
- [Sea-Level Fluctuations over the Last Millennium](#)
- [Storm Surge](#)
- [Tidal Datums](#)
- [Tidal Environments](#)
- [Tidal Inlets](#)
- [Tidal Prism](#)
- [Tide Gauges](#)

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Time Series Modeling

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Fundamental Concepts of Coastal Time Series

There has been a considerable expansion in the coastal database in the past two decades. Discrete and continuous data are collected from the coast and its interacting hydrodynamic, biological, morphological, sedimentological, and other associated subsystems in order to make better decisions on coastal management and sustainable development. While the quality and length of the collected coastal data vary greatly, many datasets fall within the domain of time series analysis. Any time series can be considered as a time (or space) ordered sequence of realizations (or observations) of a variable of

interest. The set of observations generated sequentially in time could be either continuous or discrete. In statistical terms, a time series is a realization or sample function from a certain stochastic or random process. The time series to be analyzed can be considered as a particular realization, produced by the underlying probability mechanism of the system (i.e., coastal) under consideration. The observed values of a stochastic process are generally considered as a realization of the stochastic process. Time series analysis can consider a class of stochastic processes, called stationary processes, which assume that the process is in a particular state of statistical equilibrium (Box et al. 1994). In the analysis of stationary stochastic processes it is worthwhile to note that “stationary processes generally arise from any “stable” system which has achieved a “steady-state” mode of operation” (Priestley 1981, p. 14), whereas a nonstationary series is regarded as having properties which change with time.

Since in time series analysis, inferences can be made from a realization to the generating process, it is necessary to consider the statistical properties of the time series. Essentially, the properties of a time series can be obtained from a single realization over a time interval or based on several realizations at a particular time. The properties based on a time interval of a single realization are referred to as time-averaged properties. The properties from several realizations at a given time are considered as the ensemble properties. “Since different sections of a time series resemble each other only in their average properties, it is necessary to describe these series by probability laws or models” (Jenkins and Watts 1969, p. 2).

Coastal Time Series Modeling

Since stochastic processes deal with systems which develop in accordance with probabilistic laws, stochastic models can be used to gain insights on the spatial and temporal behavior of the coastal system (Lakhan 1982; Lakhan and Trenhaile 1989). Time series modeling of the many natural phenomena occurring in the coastal environment, which appear to behave in random or probabilistic ways, is essential for understanding not only the operating processes, but also for many coastal applications. Many examples of the use and applications of probabilistic and time series models in coastal studies have been provided by Guedes Soares (2000). Even the modeling of one sea-state parameter has many applications in coastal and offshore engineering. For example, modeling a time series of significant wave height at a location is a useful complement to long-term probabilistic models that describe the wave climate in different areas (Cunha and Guedes Soares 1999).

The complex nature of the coastal system makes it necessary to develop models which can explain both the deterministic and random features of the time series for any coastal process (Lakhan 1989). For example, sea-states can be visualized as

random and unpredictable. However, there is some constancy of a sea-state for a short duration of several minutes to fractions of minutes, and therefore it is possible to assume stationarity for such short wave records. Conversely, a wave record of long duration will not exhibit stationarity because the significant wave height and other statistics are known to vary with respect to time (Goda 2000). To characterize and predict the probabilistic nature of the sea surface a time series model can be developed to represent the sea surface elevation as a non-stationary stochastic process, with the sea-state considered to be a Gaussian, statistically stationary stochastic process in short time periods. In modeling stationary or nonstationary time series, coastal and allied researchers have utilized either the frequency domain or the time domain approach.

Approaches to Time Series Analysis

Given the fact that there are various terminology, theoretical and practical aspects of time series analysis, this short review will introduce only the fundamental concepts for coastal and allied researchers. Hannan et al. (1985) pointed out that since its inception the theory and practice of the analysis of time series has followed two lines. One of these proceeds from the Fourier transformation of the data and the other from a parametric representation of the temporal relationships. The two approaches to time series analysis, commonly referred to as the frequency domain approach (or spectral analysis approach) and the time domain approach, have been discussed in several books (e.g., Anderson 1975; Otnes and Enochson 1976; Gottman 1981; Kendall and Ord 1990; Wei 1990; Bendat and Piersol 1993; Harvey 1993; Hamilton 1994; Brockwell and Davis 1996; Pollock 1999; Shumway and Stoffer 2000). In brief, the frequency domain approach uses spectral functions to study the nonparametric decomposition of a time series into its different frequency components. The time domain approach concentrates on the use of parametric models to model some future value of a time series as a parametric function of the current and past values.

While the frequency and time domain approaches can be used to provide different insights into the nature of the actual time series it should, however, be pointed out that both approaches are mathematically equivalent. According to Gottman (1981), the two approaches are linked by the famous theorem called the Wiener-Khinchine Theorem which provides a shuttle between the frequency and time domains. The Wiener-Khinchine Theorem shows that there is a one-to-one relationship between the autocovariance function of a stationary process and its spectral density function (Pollock 1999). Knowing the correlation structure in the time domain corresponds to knowing the form of the spectrum in the frequency domain. Essentially, the autocorrelation function and the spectrum function form a Fourier transform pair (Kendall and Ord 1990).

Although the two approaches are complementary rather than competitive (Harvey 1993), there are situations when one approach is more appropriate to use than the other. According to Shumway and Stoffer (2000) the two approaches may produce similar answers for long series, but the comparative performance over short samples is better done in the time domain. Given this observation, brief remarks and selected coastal applications for both approaches will, therefore, be provided. Emphasis will then be placed on modeling in the time domain because there are well-established techniques for model selection, identification, and estimation.

The Frequency Domain Approach

Extensive discussions on the frequency domain, or spectral approach, to time series analysis can be found in several books (e.g., Jenkins and Watts 1969; Rayner 1971; Kanasewich 1973; Koopmans 1974; Brillinger 1975; Bloomfield 1976; Priestley 1981; Brillinger and Krishnaiah 1983; Brigham 1988; Brockwell and Davis 1996; Fuller 1996; Ramanathan 1998; Pollock 1999). According to Jenkins and Watts (1969, p. 16), “spectral analysis brings together two very important theoretical approaches, the statistical analysis of time series and the methods of Fourier analysis.” Since, in the frequency domain approach, Fourier transforms play a very important role (Brillinger and Krishnaiah 1983), some brief remarks will be made on Fourier’s methods which form the basis of all spectral analysis (Rayner 1971).

Time series spectral analysis can be traced to Jean Baptiste Joseph de Fourier (1768–1830) who made the claim in 1807 that an arbitrary function defined on a finite interval could be represented as a infinite summation of cosine and sine functions (see Lasser 1996). Many mathematicians have worked on the development of the techniques of Fourier analysis, and contemporary books (e.g., Köner 1988; Lasser 1996; Ramanathan 1998; Howell 2001) have presented various aspects of the mathematics of Fourier analysis. Without discussing the details of Fourier techniques it is worthwhile to note that one notable early investigation which focused on analyzing time series in the frequency domain was Schuster (1898) who employed the technique of periodogram analysis. The early underlying model expressed the series as a weighted sum of perfectly regular periodic components upon which a random component was superimposed. While much of the theory of spectral analysis of random processes focused on stationary processes, in recent years, “a new form of spectral analysis has been developed which, while not accommodating all nonstationary processes, does however enable us to treat a fairly large class of such processes in a unified theory which includes stationary processes as a special case” (Priestley 1981, p. 17).

The Fourier analysis of stochastic processes can provide a representation of an infinite sequence in terms of an infinity of trigonometric functions whose frequencies range continuously. The underlying stochastic process can be represented by the Fourier integral. This can be attained by describing the stochastic processes which generate the weighting functions. There are two weighting processes, associated respectively with the sine and cosine functions; and the function that defines their common variance is the so-called spectral distribution function whose derivative is the spectral density function or the “spectrum” (Pollock 1999). For practical purposes, the spectral density function can be referred to as the power spectrum. “Since the power spectrum is the Fourier cosine transform of the autocovariance function, knowledge of the autocovariance function is mathematically equivalent to knowledge of the spectrum, and vice-versa” (Box et al. 1994, p. 39). The power spectrum provides significantly more information about the time series than simply the total power. The estimated spectrum can be used to obtain insights about the mechanism that generated the data (Koopmans 1974).

The power spectrum plays a very important role in the analysis of coastal time series. By considering the probability distribution of the sea surface as nearly Gaussian, a good approximate description is provided by the covariance function, the Fourier transform of which is referred to as the wave spectrum. In coastal studies, the wave spectrum has physical meaning because it can be demonstrated to be the density function specifying the distribution of energy over wave components with different wave number vectors and frequencies. Its integral over all wave components is proportional to the total wave energy per unit area (see Komen et al. 1994). It should be noted that the components of the spectrum are the squares of the wave amplitude at each frequency, which are related to wave energy. Since the covariance function can be related directly to the energy spectrum or the wave spectrum, the covariance function is computed, and then its Fourier transform is calculated to obtain the power spectrum (Dean and Dalrymple 1984).

With the implementation of the Fast Fourier Transform (FFT) algorithm of Cooley and Tukey (1965), many researchers have followed the contributions of some significant early studies (e.g., Pierson and Marks 1952; Munk et al. 1959; Hassleman et al. 1963) and utilized spectral concepts for either the analysis or the modeling and prediction of ocean waves. Simplified discussions on the use of spectral techniques for the analysis and modeling of the spectrum of ocean waves have been provided by Dean and Dalrymple (1984), Tucker (1991), and Goda (2000). Timely reviews on the importance and applications of spectral concepts to the description and modeling of water waves have been presented by several authors, among them Cardone (1974), Cardone and Ross (1979), Komen et al. (1994), and Cardone and Resio (1998). Recent advances in the development of spectral wave modeling for various application purposes have been

presented by several researchers (e.g., Holthuijsen et al. 1993; Rivero et al. 1998; Booij et al. 1999; Ris et al. 1999; Monbaliu et al. 2000; Schneggenburger et al. 2000). In the frequency domain perspective, some researchers (e.g., Guedes Soares and Ferreira 1995; Lakhan 1998) have modeled parameters such as significant wave height. A Fourier representation described the periodic components of time series of significant wave height values from different coastal locations. In addition, Hegge and Masselink (1996) also demonstrated the usefulness of spectral techniques to model data from topographic profiles. The results of the spectral analysis represented the amount of variance of the time series as a function of frequency. With spectral modeling, coastal researchers are obtaining greater insights on the physical characteristics of the generating mechanisms of coastal time series.

The Time Domain Approach

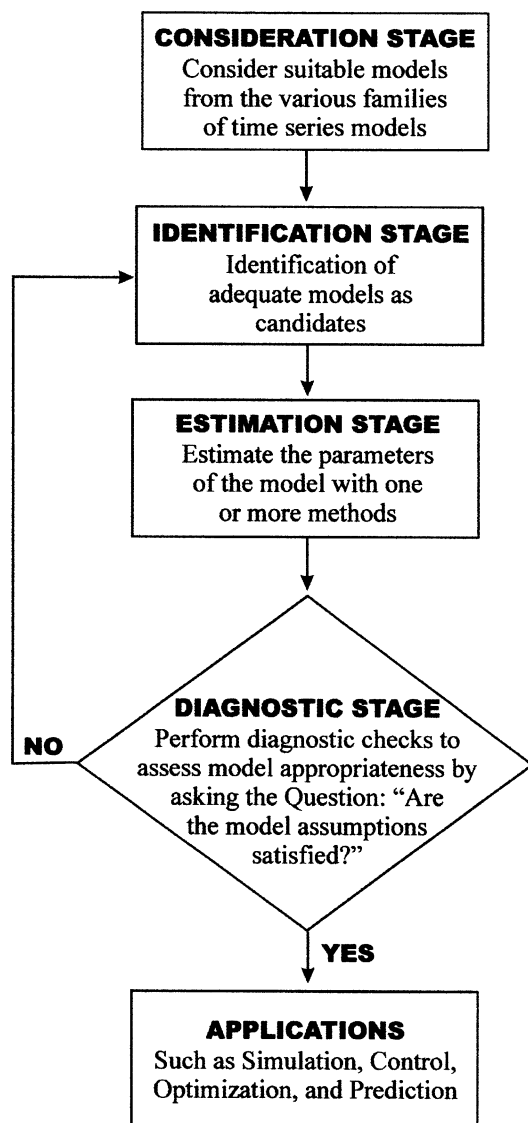
The time domain approach, which focuses on the contribution of parametric models for single series or models for two or more causally related series, can be traced to the classical theory of correlation. Details on modeling and forecasting in the time domain can be found in several books (e.g., Box and Jenkins 1970, 1976; Kendall 1973; Anderson 1975; Gottman 1981; Abraham and Ledolter 1983; Hoff 1983; Pandit and Wu 1983; Pankratz 1983; Vandaele 1983; Chatfield 1984; Montgomery et al. 1990; Bowermand and O'Connell 1993; Harvey 1993; Box et al. 1994; Hamilton 1994; Hipel and McLeod 1994; Armstrong 2001; Pourahmadi 2001). The time domain models that originated with Yule (1927) and Slutsky (1937) provided a strong foundation for time series analysis in the time domain. Yule (1927) pioneered the concept of autoregression and his autoregressive model was generalized by Walker (1931) to allow for dependence on more than two previous values. Wold (1938) followed the practice of Yule and Walker and plotted correlation coefficients against their lags. According to Pourahmadi (2001, p. 35), "the term correlogram was coined by Wold (1938, p. 7) as a substitute to the Schuster's periodogram and has been used effectively ever since as a means of identifying model or appropriate probabilistic description of time series data." The early autoregressive (AR) model of Yule (1927) and moving average (MA) models of Walker (1931) and Slutsky (1937) were not widely used because of the lack of appropriate methods for identifying, fitting, and checking these models (Jenkins 1979). AR and MA types of models were combined into the mixed autoregressive moving average (ARMA) model.

ARMA models have their foundation in the work of Box and Jenkins (1970, 1976) who utilized concepts from mathematical statistics and classical probability theory. Box and Jenkins also extended ARMA models to include certain types of nonstationary series, and proposed an entire family

of models, called autoregressive integrated moving average (ARIMA) models, which can be applied to practical problems in several disciplines. Besides ARMA and ARIMA models, Box and Jenkins (1976) presented other models, including transfer function noise (TFN) models and seasonal autoregressive integrated moving average (SARIMA) models. "The ARMA and ARIMA classes can provide very useful descriptions for a wide range of time series data and the Box-Jenkins approach (as it has become known) has been used extensively by time series practitioners in a wide variety of fields" (Priestley 1997, p. 16). In the field of coastal research, several autoregressive models have been utilized for data on sea-state parameters. For example, Lakhan (1981) and Lakhan and LaValle (1986) modeled the correlation structure of wave heights and wave periods, and then parameterized stochastic models with autocorrelated distributed variates. Spanos (1983) used ARMA processes to simulate individual waves in short-term periods. Scheffner and Borgman (1992) also simulated individual waves but accounted for long-term variability. Several other studies (e.g., Guedes Soares and Ferreira 1996; Cunha and Guedes Soares 1999; Guedes Soares and Cunha 2000) have modeled time series of sea-state parameters with autoregressive models. Besides sea-state parameters, researchers (e.g., Walton 1999; LaValle et al. 2001) have also modeled shoreline changes with time series techniques. The study by LaValle et al. (2001) utilized Box-Jenkins modeling procedures to identify models which best described a time series (1978–94) of beach and shoreline data. By following the Box-Jenkins model construction approach described below, it was found that a spatial model described the variation for beach net sediment flux, and a space-time autoregressive model provided the best fit for data on spatial-temporal variations of shoreline retreat. Here, it must be mentioned that ARMA models can be generalized to include spatial location (Cressie 1993).

Time Domain Modeling-the Box-Jenkins Approach

The success of the Box-Jenkins modeling approach in coastal and other studies can be attributed to the fact that Box-Jenkins models have several advantages over traditional models because there is a large class of models, there is a systematic approach to model identification, and the validity of models can be verified (Hoff 1983). While a publication of this kind cannot provide technical details on the construction of Box-Jenkins models, it is, nevertheless, worthwhile to note that no matter what type of stochastic model is to be fitted to a given dataset it is recommended that the identification, estimation, and diagnostic check stages of model construction be followed (Box and Jenkins 1976). By following Box and Jenkins, univariate models for any coastal time series can be constructed by using the iterative approach presented in Fig. 1.



Time Series Modeling, Fig. 1 Stages in the iterative approach to model construction. (Modified from Box et al. (1994, p. 17))

Consideration Stage

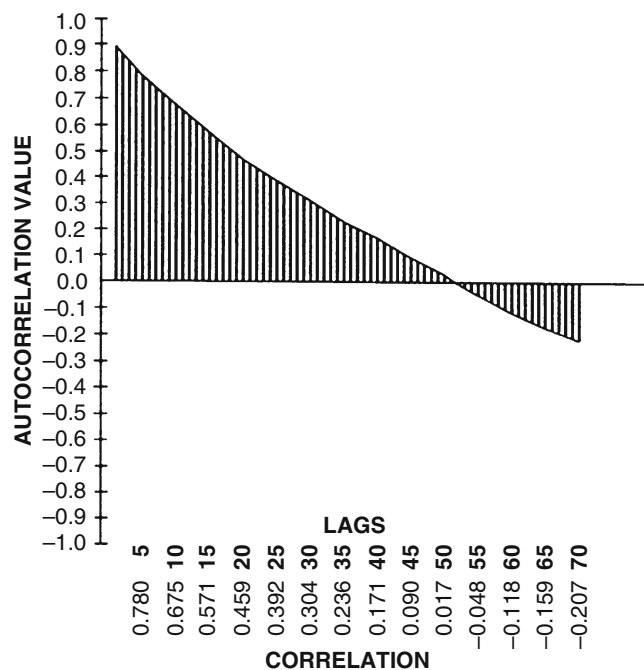
In the consideration stage, it is necessary to be cognizant of all standard time series models. Consideration must be given to the various families of stochastic models, which can be fitted to the coastal time series. It is of paramount importance to consider those models, which on the basis of theory, practical experience, understanding of the problem, and the published literature, have the potential to fit the observed data.

Identification Stage

At the beginning of the identification stage it is best to ascertain the subclasses of models that hold greater promise for adequately modeling the coastal time series. The first step is to obtain a graphical plot of the data because a plot of the time series can demonstrate some of the essential mathematical

characteristics of the data. From the plot, it can be determined whether the series contains a trend, outliers, seasonality, non-constant variances, and other non-normal and nonstationary phenomena. This knowledge allows for possible data transformations. Differencing and variance stabilizing transformations can be used. The normal procedure is to apply variance stabilizing transformations before taking differences because differencing may create some negative values. To stabilize the variance the Box-Cox power transformation can be used. For a series with nonconstant variance a logarithmic transformation can be employed (Wei 1990). In modeling time series of significant wave height data, Guedes Soares and Ferreira (1996) tried both the Box-Cox transformation and the logarithmic transformation. Following transformation to stationarity it is necessary to compute and examine the sample autocorrelation function (ACF) and the sample partial autocorrelation coefficient (PACF) of the original series to determine the need for further differencing.

After performing the necessary differencing, the ACF and PACF are computed for the properly transformed and differenced series. The autocorrelations and partial autocorrelations of a series are considered principal tools for identifying the correct parameters to include in a Box-Jenkins ARIMA model. Autocorrelations are statistical measures computed from the time series data. An autocorrelation measures how strongly time series values at a specified number of periods apart are correlated to each other over time. The number of periods apart is called the lag. The rule of thumb is that the maximum number of lags should not exceed one-fifth of the number of observations. The partial autocorrelation is similar to autocorrelation, except that when calculating it, the (auto)correlations with all the elements within the lag are partialled out (Box and Jenkins 1976). In univariate Box-Jenkins modeling, it is normal practice to produce correlograms depicting the autocorrelation coefficients plotted against the lag intervals. According to Harvey (1993), the correlogram is the basic tool of analysis in the time domain. An inspection of the correlogram provides important information as to whether the series exhibits a pattern of serial autocorrelation which can be modeled by a particular stochastic process or whether the series is random. Figure 2 is a correlogram of significant wave height data which shows that the data values are not independent of each other. The series is not random because the autocorrelations should be near zero for randomness. By understanding the association between the ACF and PACF and their corresponding processes, a tentative model can be identified. In the case of Fig. 2, with the autocorrelation function decaying exponentially, and knowing that the PACF has a distinct spike at lag 1, it is possible to specify an AR(1) model. It should, however, be stressed that several other models could fit the data. The primary goal is to identify a model with the smallest number of estimated parameters.



Time Series Modeling, Fig. 2 Correlogram of significant wave height. (From Lakhan 1998)

Parameter Estimation

Subsequent to identifying an adequate model for fitting to a particular series, the next step is to estimate the parameters in the model. Hoff (1983) outlined several objectives in estimating a model, among them obtaining fitted values that are nearly identical to the original series values, obtaining residuals that are not correlated to each other, and using a minimum number of parameters as necessary. A good model incorporates the smallest number of estimated parameters which are needed to fit the patterns in the data. To estimate parameters such as the mean of the series, AR parameters, MA, and other parameters for an identified ARMA model, several estimation procedures can be utilized. Details in estimation theory and estimation procedures can be found in several publications (e.g., Kruskal and Tanur 1978; Sachs 1984; Mendel 1987; Kotz and Johnson 1988; Box et al. 1994). For the time series modeling of coastal data, the method of moments, and the maximum likelihood method are widely used approaches to parameter estimation.

Diagnostic Checking

Once the parameters of the identified model have been estimated, the next phase is to perform diagnostic checks to determine whether certain assumptions about the model can be verified. Two assumptions to be checked are usually the normality of residuals of the model, and independence. To ascertain whether the residuals are white noise, the residuals from the estimated model are used to calculate the autocorrelation coefficients. Ideally, the residual autocorrelation function

for a properly constructed ARIMA model will have autocorrelation coefficients that are all statistically zero. If the residuals are autocorrelated they are not white noise, and this requires formulating another model with residuals that are consistent with the independence assumption (Pankratz 1983).

The portmanteau lack of fit test is usually applied for testing the independence of a time series. The portmanteau lack of fit test, originally proposed by Box and Pierce (1970), was improved by Ljung and Box (1978). Applications of the portmanteau test for physical time series can be found in several studies (e.g., Hipel and McLeod 1977; Salas et al. 1980; Hipel and McLeod 1994). The test statistic is the modified Q statistic which uses all the residual autocorrelations as a set to check the joint null hypothesis. The Q statistic approximately follows a chi-square distribution, with degrees of freedom equal to the total number of lags used minus the number of model parameters and their associated probability values. Recent studies (e.g., LaValle et al. 2000, 2001) have demonstrated that the Q statistic is appropriate for determining the goodness-of-fit of autoregressive models fitted to data on water levels and beach and shoreline changes.

Applications

When a model is accepted it could be used for various applications, among them filtering and control, simulation and optimization, and prediction. A model that fails one or more diagnostic checks is rejected. To construct a good model it becomes necessary to return to the identification stage, and repeat the iterative process of identification, estimation, and diagnostic checking.

Selection of the Best Model

In coastal time series analysis, it is possible to have several appropriate models that can be used to represent a given dataset. To solve the problem of choosing the best model from the various adequate models, several model determination procedures and model selection criteria have been proposed (e.g., Stone 1979; Hannan 1980). Selection criteria are different from the model identification methods discussed above because when there are several adequate models for a given dataset the selection criterion is normally based on summary statistics from residuals computed from a fitted model or on forecast errors calculated from the out-sample forecasts (Wei 1990).

Some well-known model selection criteria based on residuals are the Akaike Information Criterion (AIC) of Akaike (1974), Parzen Criterion for Autoregressive Transfer (CAT) Functions of Parzen (1977), and the Schwartz's Bayesian Criterion (SBC) of Schwartz (1978). Of the various selection criteria, the AIC is widely used in time series model fitting because it increases the speed, flexibility, accuracy, and simplicity involved in choosing the "best" model. In addition, the AIC is useful for application to many different kinds of time

series (Hipel 1981; Hipel and McLeod 1994), and facilitates the selection of a parsimonious model that, at the same time, provides a good statistical fit to the data being modeled.

Cross-References

- [Coastal Modeling](#)
- [Simple Beach and Surf Zone Models](#)
- [Surf Modeling](#)
- [Wave Climate](#)

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Tors

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The word “tor,” Celtic in origin, is used generally in the British Isles to denote a rather tall rock column (Cunningham 1968; Jackson 1997). Linton (1955) was the first to propose it as a scientific term in describing the tors at Dartmoor, Devonshire, England, now considered the type

Maurice Schwartz: deceased.

Tors, Fig. 1 The tor at Anse Source D'argent, La Digue Island, Seychelles, Indian Ocean; often considered to be the most beautiful beach in the world (Photo courtesy of New Adventures)



area for the feature (Palmer and Neilson 1962). Early hypotheses for the origin of tors invoked deep weathering along joint planes in granite, with subsequent removal of the loose material leaving exposed columns. Alternate possibilities outlined by Cunningham (1968) include differential erosion during scarp recession and relict subaerial prominences formed in the Tertiary. Since tors have been found worldwide, often in granite but also in other igneous, sedimentary, and metamorphic rocks, it is appropriate to consider Palmer and Neilson's (1962) pronouncement that "It is not possible to offer a definition that will encompass the many landforms to which the name "tor" has been given." With such wide distribution and varied lithology it is arguable that the most scenic among all of these are coastal tors, as can be seen in the Seychelles and Virgin Islands.

Seychelles

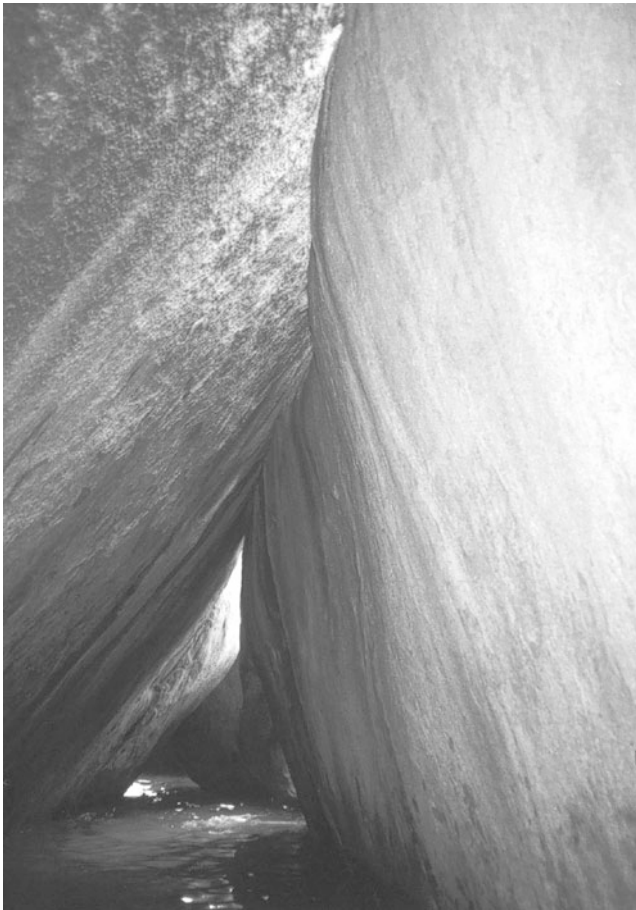
The Seychelles, along with Madagascar, were displaced in a northeasterly direction away from the African landmass during the early formation, in the Jurassic, of the Indian Ocean (Brathwaite 1984). As such, 42 of the 116 islands comprising the Seychelles Archipelago are the world's only mid-ocean islands composed of granitic rocks (Cilek 1978). Grey and pink amphibolitic granite, of late Precambrian age, is spectacularly displayed on the northern island of La Digue and at Mahe as discussed by Wagle and Hashimi (1990).

La Digue Island has long been known for its world-class resorts featuring tropical flora, white-sand beaches, swimming in crystal clear waters, snorkeling among coral reefs and magnificent scenic views. Anse source d'argent beach, located on the island, is the site of several tors, chief among them that is pictured in Fig. 1. The size of the pink tor can be judged when compared with the heads of the three swimmers seen in the mid-foreground. Graphic too is the weathered "fluting" described as typical of tors by Linton (1955) in his pioneering work, and by Brathwaite (1984) for a tor on Mahe.

Virgin Islands

The Virgin Islands, located in the northeastern Caribbean along the leading edge of the Caribbean plate, are composed of Mesozoic and lower Tertiary deformed island-arc terrane. Much of the northeastern British Virgin Islands region is underlain by the Virgin Gorda granitic pluton or batholith, which was intruded into the surrounding country rock in mid-to later Eocene time (Mattson et al. 1990). Weathered exposures in tonalitic rocks at the southern end of the island of Virgin Gorda reveal huge boulders now located upon the beaches (Weaver 1962) in a park system managed by the B.V.I. National Parks Trust.

The origin of these boulders has been described by Ratté (1986) in the classic Linton (1955) style for tors of deep weathering along joints followed by removal of the rotted material. A favorite with tourists, the site is called "The



Tors, Fig. 2 Grotto at the base of the tors at The Baths, Virgin Gorda, British Virgin Islands (Photo: M. Schwartz)

Baths,” not because of the underlying batholith as a geologist would imagine, but for the salt water pools in the grotto at the base of the tors (Fig. 2). Here one may walk, and crawl, along a trail between boulders that range up to 4 m in height.

Cross-References

- [Boulder Beaches](#)
- [Caribbean Islands, Coastal Ecology and Geomorphology](#)
- [Coastal Hoodoos](#)
- [Indian Ocean Islands, Coastal Ecology and Geomorphology](#)
- [Tourism, Criteria for Coastal Sites](#)
- [Weathering in the Coastal Zone](#)

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Tourism and Coastal Development

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²Tidal Delta Consulting, Seattle, WA, USA

Coastal tourism is a process involving tourists and the people and places they visit. It is more specifically defined as tourism brought to bear on the coastal environment and its natural and cultural resources. Most coastal zone tourism takes place along the shore and in the water immediately adjacent to the shoreline. Coastal tourism activities occur outdoors and indoors as recreation, sport and play, and as leisure and business (Miller and Ditton 1986). As with other human endeavors in the coastal zone associated with development, tourism is viewed positively by some for the opportunities it creates. Others condemn coastal tourism for its unacceptable consequences.

Coastal tourism destinations fall all along an urban–rural continuum (see “► [Demography of Coastal Populations](#),” *q.v.*). At one end of the scale are major cities and ports (Hong Kong, Venice, New York, Rio de Janeiro, and Sydney come to mind) known for their cultural, historical, and economic significance. At the other end of the continuum are the relatively isolated and pristine coastlines found around the world that are valued for their natural beauty, flora, and fauna. Of course, many coastal tourism destinations offer rich mixtures of cultural, historical, social, environmental, and other values to visitors.

Coastal tourism technologies of travel include both those which carry tourists from their homeland (e.g., airplanes, ships, cars, buses, and trains) and which are regarded by

travelers as mere means to the end of arriving at destinations, and those which transport tourists at coastal destinations but which become part of the touristic experience (e.g., cruise ships, personal watercraft, sailboats, dive boats, motorcycles, bicycles, and forms of animal transportation). Again, some transportation technologies can, depending on the circumstance, be important for being both convenient and for being interesting or pleasing.

In a manner of speaking, all tourism is a matter of supply and demand. With this perspective, coastal tourism is a business for those who make a living by developing accommodations and attractions, and by providing touristic and recreational products and services. Competing marketing programs of a multifaceted industry alert tourists and would-be travelers to coastal tourism amenities. Today, tourists travel to the coastal zone for parts of a day, for weekends, for short vacations, and for prolonged stays. Depending on the circumstances, they may travel alone, with family, or in groups. Some coastal tourism is organized for a special purpose such as ecotourism, adventure tourism, scientific tourism, and dive tourism. Coastal tourism accommodations range from small residences and camping sites rented out as opportunities arise, to single bed-and-breakfast and hotel rooms, to luxury suites in resort enclaves.

Many coastal tourism activities count as a business for those in the tourism industry and as an experience for tourists. Scuba diving, for example, provides an excellent example of how advances in technology have provided foundations for business and have facilitated touristic access to the marine environment. Other coastal activities that have a business aspect (involving, for example, guides and instructors, or special equipment) include recreational and sport fishing, boating, sailing and parasailing, and whale and bird watching. Then too, there are many forms of coastal tourism – swimming and body surfing, snorkeling, beach-combing, hiking and rock climbing, sketching and painting, photographing, sightseeing – that are “free,” or for which costs to providers are recovered indirectly through taxes, or are incorporated in standard hotel or accommodation billing practices. In recent years, windsurfing, body-boarding, wake-boarding, kite-surfing in addition to surfing (*q.v.*), have reached new levels of popularity in the coastal zone.

At the same time that coastal tourism fosters economic relationships between industry producers and tourist consumers, the process has shown itself to be an enormously potent force in transforming the natural environment and the lives of people who are neither part of the business of tourism nor a member of the community of tourists.

Coastal tourism is inherently controversial. The coastal zone is a scarce resource prized not only by those who engage in and profit by tourism, but also by those with personal residences near the sea, and those who find employment in fishing, aquaculture (*q.v.*), maritime shipping, nuclear energy,

and national defense, among other industries. Congestion and competition in the coastal zone frames the characterization and the resolution of tourism issues. Coastal tourism problems and opportunities are therefore properly debated as “multiple-use” or “multiple-value” conflicts.

Origins of Tourism

Although the early Greeks and Romans were known to enjoy the seashore for leisure purposes, coastal tourism has its roots in the Grand Tour traditions beginning with the Renaissance. As an educational institution, the Grand Tour offered young men first-hand exposure to European courts, customs, and prominent cities and ports. Not surprisingly, the Grand Tourists and their “bear-leader” tutors mixed education with pleasure. By the start of the eighteenth century, these tourists had begun to develop an aesthetic vocabulary that allowed them to more fully appreciate not only the “beautiful” in nature, but the “picturesque” and the “sublime” as well. Especially popular with the Grand Tourists were seascape paintings by Claude Lorrain (1600–82), Salvator Rosa (1615–73), and Gaspard Poussin (1615–75), depicting storms, shipwrecks, harbors, rocky coastlines, and ruins.

The mid-eighteenth century European “discovery” of the seashore for spa and medicinal purposes in England gave rise to early forms of coastal tourism. In the first half of the nineteenth century, coastal resorts saw a faster rate of population increase than manufacturing towns and by the mid-nineteenth century the medicinal beach was replaced by the pleasure beach. In 1841, the London to Brighton railway was opened, and in the same year Thomas Cook began a legendary career by promoting his first group excursion (Manning-Sanders 1951; Hern 1967; Corbin 1994). Since that time, beaches, atolls, islands, and harbors around the world have supported coastal tourism (see “► [Beach Use and Behaviors](#),” *q.v.*).

Magnitude of Coastal Tourism

Although there are no standardized practices for reporting tourism statistics within the coastal zone, it is not difficult to see how tourism has a major coastal aspect. More than 70% of the earth is covered by water, and only several dozen out of well over 200 nations in the world lack coastlines (Miller and Auyong 1991a).

International Trends

World Tourism Organization statistics confirm that tourism is the world’s largest industry as measured by the number of people involved and by economic impacts. From 1945 to 2000, international arrivals increased from 25 million to

a record 699 million (WTO 1996, 2001). Between 1970 and 1990, tourism grew by nearly 300% (United Nations Environment Programme 1992).

By the year 2020, it is estimated that international tourist arrivals will reach over 1.56 billion. Statistical estimations for total tourist arrivals by region show that in 2020 the top three receiving regions will be Europe (717 million tourists), east Asia and the Pacific (397 million tourists), and the Americas (282 million tourists), followed by Africa, the Middle East, and south Asia (WTO 2001).

The World Travel and Tourism Council (WTTC 1995) reports that between 1980 and 1989, economic expenditures on international travel (excluding transportation) doubled to \$209 billion, rising one-third faster than the world gross national product. Additionally, travel and tourism generated an estimated \$3.4 trillion in gross output in 1995 – creating employment for 211 million people, and producing nearly 11% of the world gross domestic product. This growth reflects investments of \$694 billion in new facilities and equipment, and contributions of more than \$637 billion to global tax revenues (WTTC 1995).

In 2000, tourism generated total international receipts of US\$ 476 billion (WTO 2001). Of the world's top 15 tourism destination countries in 2000, 12 were countries having coastlines (WTO 2001). In 2000, the cruise ship industry expected to host over 6.5 million passengers which would represent a 1,200% increase in number of passengers since 1966 (Godsman 2000).

Sun, beautiful beaches, and warm ocean waters have become standard vacation requirements for many tourists. Of those visiting the Caribbean 49% do so for the beaches, while 28% are primarily interested in sightseeing, and 17% in water sports (Waters 2001). The Pacific region (which includes Australia) has enjoyed a healthy annual growth rate of nearly 4% since the mid-1990s, though arrivals represent only 1.4% of the world's inbound travelers. Of the Pacific territories, French Polynesia is the most dependent on travel with 78% of the nation's GDP coming from tourism. Tourism in the African region has been growing at a faster rate than for many other parts of the world (Waters 2001).

US Trends

The Travel Industry Association of America (TIA 2001) reports that tourism is the nation's largest services export industry, the third largest retail industry, and one of America's largest employers. TIA (2001; see also WTTC 1995) has tabulated that travel and tourism in the United States alone has an impact exceeding \$541 billion a year in expenditures which includes spending by US resident and international travelers within the United States on travel-related expenses (i.e., transportation, lodging, meals, entertainment, and

recreation, as well as international, fares on US flag air carriers). This generates more than 17.5 million jobs, and fuels the largest trade surplus of any industry, totaling nearly \$25 billion in 1999. Between 1986 and 1996, international visitation to the United States grew by 78% and expenditures by foreign visitors grew by 223%. It is estimated that more than 90% of foreign-tourist spending occurs in coastal states (US Travel and Tourism Administration 1994).

US coastlines are popular sites for tourism and recreational activities. Coastal beaches, wetlands, fisheries, aesthetic landscapes, and the human-designed facilities and attractions in the touristic hinterland combine in an endless list of inviting opportunities for visitors, local residents, and entrepreneurs. The major recreational elements of coastal tourism are visiting beaches, swimming, snorkeling and scuba diving, boating, fishing, surfing, and wildlife watching.

It is important to note that coastal tourism and recreation activities often overlap and are not always confined to the marine and coastal environment. For example, diving, fishing, and whale watching are often done while boating, surfing, swimming, and bird watching are usually done while visiting beaches and coastlines; and not all recreational boats are used exclusively in marine and coastal waters.

In the United States, beaches are the leading foreign and domestic tourist destinations (Houston 1996). In 1995, coastal states made up 11 of the top 15 destinations for overseas travelers visiting the United States (Waters 1997). A 1999 survey that measured travelers' satisfaction with their visits revealed that Hawaii, Alaska, California, and Florida – all coastal states – were the top four “most liked” destinations in the United States (Volgenau 2000).

Tourism and recreation are highly significant economic activities in the US coastal zone. By one estimate, approximately 180 million people visit the coast for recreational purposes, and 85% of tourist-related revenues are generated by coastal states (Houston 1996). Overall, beach tourism and recreation have been estimated to contribute \$170 billion annually to the US economy (Houston 1995). Coastal states receive 85% of all tourist-related revenues in the United States (Houston 1995). Coastal districts (defined in terms of state congressional districts) received more than \$185 billion in tourism expenditures in 1997 (TIA 1998). In addition, it has been estimated that US beaches and marine waters support 28.3 million jobs (Environmental Protection Agency 1995).

Comprehensive and time-series statistics measuring employment, and the economic and social value of coastal tourism and recreation in the United States are not available. Quantitative and reliable data measuring involvement in specific coastal recreation and tourism activities in the United States are limited (and often proprietary). Nonetheless, many small and unconnected studies have been conducted on specific tourism topics and destinations in the coastal zone.

Several boating and fishing statistics provide some idea of the economic and social importance of coastal tourism. In 1998, according to the US Coast Guard, registered boats numbered 12.3 million, with 10 coastal and Great Lakes states (Michigan, California, Florida, Minnesota, Texas, Wisconsin, New York, Ohio, South Carolina, and Illinois) accounting for nearly half of them (National Marine Manufacturers Association 2000). In 1999, 77.8 million people participated in recreational boating and recreational boaters spent nearly \$23 million on related products and services (National Marine Manufacturers Association 2000). Between 1991 and 1996 the number of Americans (age 16 and older) who participated in recreational saltwater fishing increased 5.6% from 8.9 million to 9.4 million (Cordell et al. 1997).

Coastal Tourism Systems

Coastal tourism systems involve interactions between people and place in destinations that include small communities and villages, self-contained resorts, and cosmopolitan cities. From a sociological perspective, coastal tourism systems have three kinds of actors: (1) *tourism brokers*, (2) *tourism locals*, and (3) *tourists* (Miller and Auyong 1998a, b).

A “broker-local-tourist” (BLT) model of a coastal tourism system is displayed in Fig. 1 (see Miller and Auyong 1991a, 1998a, b). Tourism brokers consist of persons who in one way or another pay professional attention to tourism. Main sub-categories include (1) private sector brokers who are part of the tourism industry, (2) public sector brokers at various levels of government who study, regulate, and plan tourism, and (3) social movement brokers in nongovernmental, non-profit, and environmental organizations who address tourism issues. Tourism brokers of these and other types do not necessarily agree on the kind of tourism that is “best” for coastal tourism systems. Indeed, broker–broker conflict is as common as cooperation. Tourism locals consist of persons

who reside in the general region a coastal tourism destination, but do not derive an income from tourism or engage in its management and regulation. Finally, tourists consist of persons of domestic and international origin who travel for relatively short periods of time for business, recreation, and educational purposes before returning home.

Motivation

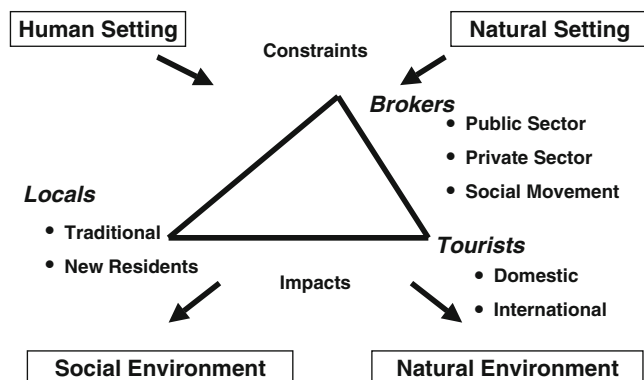
From the times polite society planned their Grand Tour itineraries through Europe to Rome and other Italian destinations in the eighteenth and nineteenth centuries, all who have participated in or witnessed the growth of tourism have pondered the motivations of those fortunate enough to travel. While there are many psychological, social psychological, and social concepts and frameworks for accounting for tourism, only several are mentioned here.

First and looking to the motives of tourists, Miller and Ditton (1986, p. 11) suggest that the fundamental promise of travel “lies in its promise of *contrast*.” In elaboration, these authors show that individual trips and vacations allow opportunities for contrast or personal change along three dimensions. *Recreational tourism* as engaged in by the athlete or escapist has a restorative purpose, and provides for change in the physiological or emotional state of the tourist. *Educational tourism* as pursued by the student has a philosophical purpose and provides a basis for change in the intellectual and artistic understanding of the tourist. *Instrumental tourism* as involving entrepreneurs, reformers, and pilgrims exhibits an economic, political, or religious purpose and leads to change in business, network, or moral opportunities available to the tourist. With this framework, a trip by one tourist to, say, a South Pacific island might be experienced as highly recreational, mildly educational, and not at all instrumental. Those accompanying such a tourist could, of course, experience the trip with different weightings along the three dimensions of touristic contrast.

A second way of considering the motivation of tourists emphasizes their intention to experience a psychological state of *challenge* that Csikszentmihalyi (1975, 1990) terms *flow* or *optimal experience*. When in a state of flow – as one might be when surfing, sailing, scuba diving, or even engaging in stimulating conversation – the tourist has found a fine match between his or her abilities and the physical, intellectual, or social challenge at hand. Flow, then, is a state of mind between boredom and anxiety. More fully:

[f]low denotes the wholistic sensation present when we act with total involvement. It is the kind of feeling after which one nostalgically says: ‘That was fun,’ or ‘That was enjoyable.’ (Csikszentmihalyi 1975, p. 43)

According to Csikszentmihalyi (1975) the flow experience is engaged in for its own sake and is marked by (1) a merging of action and awareness, (2) a centering of attention on a limited stimulus field, (3) a feeling variously described as “loss of



Tourism and Coastal Development, Fig. 1 Broker-local-tourist (BLT) model of coastal Tourism. (Adapted from Miller and Auyong 1991a, p. 75)

ego,” “self-forgetfulness,” “loss of selfconsciousness,” and even “transcendence of individuality,” and “fusion with the world,” (4) a feeling of control over one’s actions and the environment, (5) coherent, noncontradictory demands for action, and clear unambiguous feedback, and (6) its autotelic [from Greek *auto* = self and *telos* = goal, purpose] nature.

It is often remarked that people who travel together gradually develop a kind of touristic solidarity. By seeing and doing the same things, by sharing emotions and reactions, by facing a common set of logistic obstacles, and even by jointly creating a set of “story lines” with which they might talk about a trip with others, tourists are brought together through the small and multiple secular rituals of travel. In acknowledging this ritual potential, a third perspective on touristic motivation stresses the passionate *commitment* that some tourists exude in performing their favorite coastal touristic activity.

In a series of sociological studies of amateurs, volunteers, and hobbyists in sports, science, and the arts, Stebbins (1992, p. 3) noted intense levels of personal involvement and high levels of technical competence, and coined the term *serious leisure* to describe commitment that was tantamount to professionalism:

[S]erious leisure can be defined as the systematic pursuit of an amateur, hobbyist, or volunteer activity that is sufficiently substantial and interesting for the participant to find a career there in the acquisition and expression of its special skills and knowledge.

In the realm of coastal tourism, tourists who pursue serious leisure are omnipresent as evidenced by scuba divers, sailors, whale watchers, amateur naturalists and marine conservationists, and the like.

System Dynamics

Coastal tourism systems change in size and character over time. To understand and ultimately to predict these changes, and also to plan for desired societal and environmental outcomes, analysis must focus on the behavior of components of the system. In this regard, two processes merit attention.

First, population dynamics of the BLT model should be monitored. It is not unusual for individuals in the system to change statuses. This can occur as, for example, tourists who visit a coastal destination decide to stay and take on a residence, either as a broker of some kind (e.g., as a scuba dive shop entrepreneur or restaurant owner), or as a local (e.g., as a lawyer or teacher). Other transformations in status take place as locals change occupations and become private sector brokers by engaging in a tourism business, or become public sector tourism brokers by finding government employment that concerns tourism. Of course, locals and brokers take on the role of a tourist when they vacation on travel of their own.

Second, power dynamics of the BLT model should be taken into account. Tourism is often examined as a product

of the aggregate decisions of individual tourists. The relationship forged between the tourist and the local is accordingly depicted as socioeconomic in nature; tourists and locals interact as “guests” and “hosts” or as consumers and producers. Where power relationships are perceived to exist (e.g., as between First World tourists and Third and Fourth World locals), it is argued that these reveal the colonial and imperialistic leverage tourists have over those whom they visit. From this perspective, tourism systems are controlled and determined – often in unfortunate ways – by the behavior of tourists.

While there certainly are many instances in which tourists have exercised their influence to selfish and inappropriate advantage in coastal tourism settings, a narrow concentration on the power of tourists can result in analysts missing the power of tourism brokers. As Cheong and Miller (2000) have pointed out, tourists are frequently vulnerable to the power and control of brokers and locals. This is the case when, for example, tourists abide by laws and regulations of public sector brokers, and when they follow the advice and instructions of private sector brokers such as tour guides and travel agents.

Tourism Development

Coastal tourism development in the coastal zone has become a constant since the end of World War II. Well-known examples are found on the coastlines and islands of Europe, North and South America, Africa, and Asia (Miller and Auyong 1991b, 1998a, b; Conlin and Baum 1995; Lockhart and Drakakis-Smith 1997). Tourism development necessarily leads to changes in society and the environment of some kind. While conclusions about the “appropriateness,” “success,” “inappropriateness,” or “failure” of coastal tourism development projects vary to a degree with the political and economic orientations, aesthetic standards, and environmental philosophies of analysts and observers, there is no question about the power of tourism development to quickly effect dramatic change.

From a societal point of view, tourism development promises better quality of life. In theory, poverty is alleviated through the creation of new jobs. Personal income and taxes derived from tourism then fosters better health, education, and other social services. In practice, these goals are only sometimes met. In many cases, failures of political institutions have led to unfair distributions of tourism-generated revenues and to problems of environmental justice.

As noted above, the tourism process provides incentives for locals to become tourism brokers. The lives, then, of both locals and new brokers are changed by coastal development. In some instances changes in the community that are derivative of tourism are undeniably positive. In other cases, the

effect is negative. In a study of tourism in a Mexican coastal community, McGoodwin (1986) has identified a *tourist impact syndrome* which identifies the possible cultural costs to tourism system locals as including: (1) loss of political and economic autonomy (including loss of real property), (2) loss of folklore and related cultural institutions, (3) social disorganization (including radical changes in value orientations and in norms regarding social relations; heightened desire for material objects; changes in norms regarding work, sexual behavior, and drug use; promotion of illusory life aspirations; and loss of parental control and of respect for elders), and (4) hostility towards tourists (e.g., thievery, hustling, verbal aggression).

From an environmental perspective, tourism development is often seen to promise degradation of ecosystems (see “► [Human Impact on Coasts](#),” *q.v.*). This, of course, is unavoidable with the building of airports, ports, road systems, hotels, resorts, and other facilities. This said, tourism development can also provide financial support for the protection of the marine environment and endangered species, as for example, in the creation of underwater and marine parks (*q.v.*) and protected areas.

Over the last decades, there has been growing recognition of the social and environmental trade-offs of tourism and also of the unintended consequences and economic externalities of tourism development (e.g., see Mathieson and Wall 1982; Edwards 1988; Pearce 1989; Clark 1996; Orams 1999). With this, coastal tourism development is increasingly designed, debated, and evaluated against the ideal of *sustainable development*. Two prominent statements on this important concept follow:

Economic growth always brings risk of environmental damage, as it puts increased pressure on environmental resources. But policy makers guided by the concept of *sustainable development* will necessarily work to assure that growing economies remain firmly attached to their ecological roots and that these roots are protected and nurtured so that they may support growth over the long term. (World Commission on Environment and Development 1987, p. 40)

(*Sustainable development* means) improving the capacity to convert a constant level of physical resource use to the increased satisfaction of human needs. (World Conservation Union, the United Nations Environment Programme, and the World Wide Fund for Nature 1990, p. 10)

Toward the Resolution of Coastal Tourism Issues

Coastal tourism has been seen to be responsible for both positive and negative impacts to the natural and social environment (see Fig. 1). The impacts of coastal tourism on the social environment involve social, cultural, political, and economic issues. On the positive side, coastal tourism can foster community pride, improved quality of life and new job

opportunities; on the negative side, coastal tourism can lead to problems of overcrowding, social displacement, and crime. Impacts on the natural environment are often biological, physical, and ecological in nature. Increased protection and conservation of many areas and species has been a positive result of coastal tourism; nevertheless problems of erosion, pollution, and loss of species diversity occur far too frequently. It should not surprise that many coastal tourism issues simultaneously affect the social and natural environment.

The viability of coastal tourism systems and the natural environment in which they occur is very much dependent on human behavior. The resolution of coastal tourism issues can arise from the work of tourism system brokers and also from the individual decisions of locals and tourists.

Of the multiple ways available to society to control human conduct, three broker-driven mechanisms are prominent in the coastal tourism context. These mechanisms are *tourism management*, *tourism planning*, and tourism education.

Tourism management, planning, and education are crucial to the sustainable evolution of a touristic destination. It is therefore imperative that each are administered in such a way as to provide for the social and economic needs of the community, while at the same time ensuring that environmentally sensitive areas and ecologically important habitats are identified and excluded from tourism pressure. It is also recognized that tourism management, planning, and education are necessarily not only for scientific purposes and to conserve the environment for the benefit of residents, but also for the protection of long-term investments in tourism infrastructure, attractions, facilities, services, and marketing programs. It deserves to be noted that coastal tourism management, planning, and education programs are often designed and implemented by the same agencies and organizations. This overlap is often desirable and is found in some instances of larger efforts of government to promote integrated coastal zone management (see “► [Coastal Management Practices](#),” *q.v.*; see also Clark 1996; Cicin-Sain and Knecht 1998).

The manner in which a country, region, or community chooses to conduct touristic management, planning, and education activities is framed by societal (political, economic, etc.) and environmental (geography, natural resource, etc.) constraints. In the long run, the wisest course of action is to balance environmental, business, management, and social concerns so that tourism development is recognized as a potentially dangerous, but also potentially valuable and responsible course of action.

Coastal Tourism Management

Very generally, management concerns the actions of an executive decision-making entity in accordance with overarching goals of the larger enterprise in which it is housed. Although resorts, hotels, restaurants, transportation businesses, and

many other firms in the coastal tourism sector do make decisions in strategic and professional ways, management, as the term is employed here, points to work engaged in or sponsored by public sector brokers to address problems and opportunities of coastal tourism.

Throughout the world, coastal tourism is managed by regulatory entities in accordance with the structure and procedures of the prevailing political system. In the United States, coastal tourism management is undertaken by federal agencies, and by regional, state, and local authorities at other levels of government. These executive entities rely on the two other branches of government for guidance. Thus, legislatures provide the mandate for tourism management in the design of laws, and the judicial branch of government interprets law as it applies to regulatory actions and the behavior of constituencies and tourists.

In the United States the growing importance of coastal tourism to the Nation was communicated to the general public with publication of *Year of the Ocean: Discussion Papers* in 1998. In a major chapter in this document, "Coastal Recreation and Tourism" is defined as embracing:

the full range of tourism, leisure, and recreationally oriented activities that take place in the coastal zone and the offshore waters. These include coastal tourism development (hotels, resorts, restaurants, food industry, vacation homes, second homes, *etc.*), and the infrastructure supporting coastal development (retail businesses, marinas, fishing tackle stores, dive shops, fishing piers, recreational boating harbors, beaches, recreational fishing facilities, and the like). Also included is ecotourism and recreational activities such as recreational boating, cruises, swimming, recreational fishing, snorkeling, and diving. Coastal tourism and recreation . . . likewise includes the public and private programs affecting all the aforementioned activities. (US Federal Agencies 1998, p. F-2)

In stressing the need to coordinate federal coastal tourism policies and programs, the chapter discusses governmental management and planning, management of clean water, and healthy coastal ecosystems, management of coastal hazards, and beach restoration programs.

At the federal level, the United States does not have a "Department of Tourism" with regulatory authority. Instead, touristic, recreational, and leisure activities are regulated by a host of executive agencies in accordance with a suite of legal mandates. Prominent in the control of coastal tourism and recreation are the National Park Service, the National Marine Fisheries Service, the Fish and Wildlife Service, and the US Army Corps of Engineers (Table 1). These agencies oversee many programs that conserve and protect natural resources and the environment while also fostering public access to the shore and to the marine environment.

At the state-level and at the local-level of government, coastal tourism is managed through the regulatory efforts of a variety of departments (e.g., departments of fish and wildlife, and park departments) with mandates that resemble those

Tourism and Coastal Development, Table 1 US federal statutes and actions influencing coastal tourism (selected)

Antiquities Act (1906)
National Park Service Organic Act (1916)
Fish and Wildlife Coordination Act (1934)
The Wilderness Act (1964)
National Sea Grant College Program Act (1966)
National Historic Preservation Act (1966)
Executive Reorganization Plan Number 4 and Executive Order 11564 establishing the National Oceanic and Atmospheric Administration (1970)
Coastal Zone Management Act (1972)
National Marine Sanctuaries Act (1972)
Fishery Conservation and Management Act (1976; renamed Magnuson-Stevens Fishery Conservation and Management Act)
Archaeological Resources Protection Act (1979)
Fish and Wildlife Conservation Act (1980)
Presidential Proclamation 5030 establishing a 200-mile Exclusive Economic Zone (1983)
The Recreational Boating Safety Act (1986)
Abandoned Shipwreck Act (1987)
Executive Order 13158 establishing a national system of Marine Protected Areas (2000)

of federal counterparts. It is also common for city governments to cooperate with chambers of commerce in the development of infrastructural and business practice standards.

In practice, then, coastal tourism management is conducted by public sector brokers at all levels of government, by private sector brokers in businesses, and by some NGOs and environmental and social movement brokers. An array of tourism management tools (e.g., licensing regulations, zoning rules, tourist quotas, time and areas restrictions, and carrying capacity and limits of acceptable change regulations) have been used successfully throughout the Pacific, in the Caribbean, in the Atlantic and elsewhere (see Pearce 1989; Miller and Auyong 1991b, 1998a, b; Conlin and Baum 1995; Lockhart and Drakakis-Smith 1997; Orams 1999). The concepts of *zoning and carrying capacity* deserve further attention due to their particular applicability to the management of tourism in coastal areas.

Two very significant parks utilize zoning as a means of tourism management. The Great Barrier Reef Marine Park in Australia is a multiple-use protected area. With zoning, conflicting uses are physically separated. The range of protection in the park varies from virtually no protection to zones where human activity is conditionally permitted. The adoption of this zoning scheme allows the park authority, in association with interested members of the public and with other agencies, to develop and apply a tourism strategy for the entire Great Barrier Reef Marine Park. Zoning ensures that the Reef will not become overpopulated with tourist and other structures, but also allows for careful development in areas which are suitable for that purpose. The Galapagos Islands

National Park in Ecuador also employs zoning strategies. The park is effectively managed with intensive use, extensive use, and scientific use (off limits to all but a few visitors) zones.

Carrying capacity (*q.v.*) regulations illustrate the “precautionary principal” method of natural resource management and are highly regarded by practitioners of tourism management. Coastal tourism managers who seek to determine the appropriate level of use that can be sustained by the natural resources of an area are well aware that carrying capacities and use-intensity limits of tourism destinations are dynamic, and depend greatly on the biological and ecological processes of natural resources.

Coastal area carrying capacity can be evaluated in four ways (Sowman 1987). Physical carrying capacity is concerned with the maximum number of “use units” (e.g., people, vehicles, boats) which can be physically accommodated in an area. Economic carrying capacity relates to situations where a resource is simultaneously utilized for outdoor recreation and economic activity. Ecological carrying capacity (sometimes referred to also as physical, bio-physical, or environmental carrying capacity) is concerned with the maximum level of recreational use that can be accommodated by an area or an ecosystem before an unacceptable or irreversible decline in ecological values occur. Social carrying capacity (also referred to as perceptual, psychological, or behavioral capacity) is concerned with the visitor’s perception of the presence (or absence) of others simultaneously utilizing the resource of an area. This concept is concerned with the effect of crowding on the enjoyment and appreciation of the recreation site or experience.

The *limits of acceptable change* (LAC) framework developed by Stankey et al. (1985) enables managers to move beyond calculation of carrying capacity figures to address actions needed for management goals. This approach concentrates on establishing measurable limits to human-induced changes in the natural and social setting of parks and protected areas, and on identifying appropriate management strategies to maintain and/or restore desired conditions (Stankey et al. 1985). Knowledge of the natural (physical, biological) setting is combined with knowledge of the human (social, political) setting in order to define appropriate future conditions.

The LAC method employs nine steps as follows: (1) identification of area concerns and issues, (2) definition and description of opportunity classes, (3) selection of indicators of resource and social conditions, (4) inventory of resource and social conditions, (5) specification of standards for resource and social indicators, (6) identification of alternative opportunity class allocations, (7) identification of management actions for each alternative, (8) evaluation and selection of an alternative, and (9) implementation of actions and monitoring of conditions. To date, the LAC system has proved to be a valuable tourism management tool in several

wilderness areas in the United States and has direct applications to coastal areas as well.

Coastal Tourism Planning

Planning, broadly conceived, entails the consideration of a range of actions likely to contribute to the attainment of organizational goals. In some instances, overarching goals are well known in advance and planning professionals concentrate on the means that will ensure these ends. In other situations, the determination of goals requires prolonged deliberation.

Coastal tourism planning is often integrated with other resource analyses in the development of coastal area or region. Planners take into account not only visitation rates and statistics, but also the fact that tourists increasingly insist that destinations be high-quality and pollutionfree, as well as inherently interesting. Therefore, it is in both the public and private sector brokers’ interest to implement a planning strategy for tourism. The goals and policies of government agencies and businesses are, however, frequently different from one other and may even be in direct conflict. To minimize and even prevent disruptions and loss of time, communication between tourism brokers is crucial. Planning also leads to equitable distributions of coastal tourism benefits.

The success or failure of a tourism project frequently hinges on the conditions of natural amenities in the surrounding environment. This is especially true for tropical environments found, for example, on Pacific and Caribbean islands. Parks and natural resource areas, scenic vistas, archaeological and historic sites, and coral reefs are all touted tourism attractions. Marketing strategies for coastal, marine, and island tourism especially promote destinations for being close to white sand beaches. However, development of permanent structures for tourism near beaches often exacerbates beach erosion, property damage, and requires construction of shore protection structures. If touristic facilities are to be sited near beaches, proper planning is essential for the protection of the coastal zone and private property.

In many numerous coastal and island states located in the Mediterranean, Caribbean, and the Pacific where tourism is a major economic force, major national-level departments of government shape coastal tourism through the design of investment incentives and international joint venture opportunities. In nations such as Mexico and Costa Rica, these activities are linked to the preparation of strategic tourism plans.

In the United States – and with notable exceptions such as those provided by the National Park Service – very little coastal tourism planning takes place in the federal government. At the state-level, it is commonplace for departments of tourism to promote tourism. While many states have experienced great success in attracting tourists with advertising strategies, most state departments of tourism have yet to

augment the marketing of tourism with the monitoring and assessment of coastal tourism's effects on the environment and quality of life. At the local-level, many city governments have utilized their planning departments to recommend approaches to issues having to do with public use of the coastline and natural resources, the revitalization of waterfronts, and zoning appropriate to resort and marina development.

Within the private sector, coastal tourism planning is an established professional specialty. Firms of all sizes develop coastal tourism plans tailored by expert consultants to the needs of developer clients. Increasingly, social movement brokers are being seen to engage in professional coastal tourism planning.

Coastal tourism planning falls into two main categories, depending on whether the project in question is driven by a preservation or a development ethic. Preservation goals predominate in the planning of recreational areas, in national park and marine protected area planning, and in planning that is part of natural resource management. The development framework has found application in seaside resort and theme park planning, in condominium time-share planning, and in varieties of coastal city planning. There are many examples worldwide of coastal tourism zones, replete with both preservation and development projects, that extend from major cities. The Costa Brava in Spain, the French Riviera, the Yucatan Peninsula, the East Coast of Australia, and the coastlines of the United States and many Polynesian islands illustrate mixed planning.

Landscape architecture and urban planning are important in shaping coastal tourism. Both fields have public and private applications, and design and planning aspects. Both tailor products and services to preservation and development goals and, accordingly, address biophysical and social and economic objectives.

Landscape architects design parks and gardens, resort and hotel facilities, marinas and waterfronts, plazas and squares, and transportation corridors providing access to coastal touristic destinations. Urban planners design circulation facilities, city districts and spaces, and produce master planning and site design products.

Planning activities and products of landscape architects include resource management plans, environmental analyses, and multidisciplinary feasibility studies, and needs assessment and community structure plans. In overlapping ways, urban planners produce tourism policy plans, functional plans, and environmental assessments.

Because coastal tourism planning efforts are attuned to local conditions, constituencies, and financial constraints, there is no single planning process for guaranteeing success. This said, most professional planning endeavors share a general structure. Grenier et al. (1993) suggest a three-phase tourism planning process. With this, a first "Frontend

Planning" phase encompasses scoping (entailing a statement of project philosophy, pre-assessments of key issues and themes, and formulation of objectives) and research (involving data collection and analyses supporting cultural, institutional, and environmental profiles; site reconnaissance; eco-determinant mapping; and analyses of constraints and opportunities). A second "Project Planning" phase is focused on refinement of project objectives, design and evaluation of alternative development plan concepts, and selection and approval of the preferred development plan concept. A third and final "Project Management" phase concerns activities of implementation, monitoring and evaluation, and refinement.

In summary, coastal tourism planning has been fostered by public sector brokers at all governmental levels, by consultants among other private sector brokers, and by an impressive range of nongovernmental and environmental organizations in roles these have taken on as social movement tourism brokers. Coastal tourism planning practitioners have developed an array of planning methodologies (e.g., comprehensive land-use planning, integrated coastal zone planning, and strategic and special use planning) and have utilized these throughout the world, in many instances by cooperating with tourism brokers with management expertise (see Gunn 1988; Pearce 1989; Miller and Auyong 1991a, 1998a, b; Conlin and Baum 1995; Lockhart and Drakakis-Smith 1997; Orams 1999; Hadley 2001).

Coastal Tourism Education

The two mechanisms for the control of human behavior in coastal tourism systems discussed above – management and planning – are similar to one another in that the tourism experts who analyze coastal tourism situations channel their recommendations upward to regulatory and planning authorities. These tourism brokers then implement policies and plans downward, influencing tourism businesses, tourists, and locals.

A third mechanism concerns coastal tourism education and communication. Although education about coastal science and environmental issues is effectively transmitted in classrooms, discussion here focuses on education and outreach in nontraditional settings and how people learn through the experience of being tourists or learn in the course of daily life. Guided tours, museums, brochures, public lectures, newspapers, and signage are but a few of the devices that figure importantly in the educational processes linked to coastal tourism.

In a manner of speaking, tourism education contrasts with management and planning in that the first clients of analysts are not managers and planners in positions of authority, but tourists and locals. Whereas managers achieve goals through policies and regulations and planners depend on plans, coastal tourism educators succeed when people take personal initiative to change their own behavior because they have been

taught something. Tourism education, then, is a process in which analyst brokers direct their ideas outward to people involved in tourism. Tourism educators and communicators do not evaluate success or failure at attaining their goals with studies of “enforcement” or “compliance.” This is so because successful education motivates individuals by persuasion, not coercion.

By definition and referring to Fig. 1, coastal tourism educators are public sector, private sector, and social movement brokers. These brokers design products and strategies to educate people and through this to change human behavior in coastal tourism systems. While educator brokers seek to impart their message to tourists and to locals, they also educate one another as, for example, when a nongovernmental organization (NGO) educates public sector and private sector brokers.

Efforts to resolve problems and opportunities of coastal tourism through education are steadily growing throughout the world. Tourism brokers who are advancing this promising agenda are benefiting from the work of educators who have focused on environmental and sustainability issues. Monroe (1999) has characterized successful environmental education and communication projects as having features that allow for: (1) empowerment of local communities and use of their expertise, (2) attention to scientific, social, economic, political, and cultural topics, (3) identification of a variety of stakeholders and integration of them into the process, (4) advancement of an environmental ethic as well as assistance to residents in developing decision-making skills, (5) development of a gender component, (6) flexibility in project design (including realistic timetables), and (7) project evaluation.

Coastal tourism brokers (e.g., those in government or in NGOs) that provide international aid in developing and poverty-plagued states have also benefited from the cross-cultural advice of Brazilian educator and philosopher, Paulo Freire. Freire has contended that the education process has for too long been regarded as a “delivery service” from the scientific and technological elite of the Western World to those suffering in the Third World. Freire’s (1999, p. 61, emphasis added) solution lies in education projects that emphasize collaborations between experts and clients at all stages of the process:

Through dialogue, the teacher-of-the-students and students-of-the-teacher cease to exist and a new term emerges: teacher-student with student-teachers. The teacher is no longer merely the one-who-teaches, but one who is himself taught in dialogue with the students, who in turn while being taught also teach. They become jointly responsible for a process in which all learn.

Few would disagree with the proposition that coastal tourism education has great potential to enhance the quality of tourism for tourists and locals, and to also protect the environment through responsible human conduct. The importance of

education (and of overlapping fields such as communication, journalism, and environmental and science reporting by the media) is recognized by virtually all marine scientists and researchers (see Pearce 1989; Miller and Auyong 1991b, 1998a, b; Conlin and Baum 1995; Lockhart and Drakakis-Smith 1997; Orams 1999). Still, many opportunities to integrate coastal tourism education with the mechanisms of management and planning have been missed.

Challenges Ahead

Over the last several centuries, the world’s coastlines have been substantially transformed to support recreational and touristic pursuits. In some cases, coastal tourism dominates the skyline. In others, tourism is one of many industries. As coastlines become more populated and accessible, it is ever more clear that, however, beneficial coastal tourism is to the tourist, it is neither a panacea that will invigorate any local economy, nor a pollution that will necessarily ruin environments and corrupt cultural traditions and values. Coastal tourism is a process amenable to management, planning, and education. Sustainable coastal tourism obliges humanity to have respect for other life forms and the environment, while it affords opportunities for people to learn, recreate, and reach their potential as individuals through travel.

Because the stakes are high and because mistakes can be virtually irreversible, societal resolution of pressing tourism and coastal development issues requires imagination as well as sustained scientific and policy attention. Work to be done falls in the areas of research, and tourism broker and individual responsibility.

Tourism Research

Researchers in government, academe, and in the private and social movement sectors constitute a first group of practitioners whose work induces positive change in coastal tourism systems. Fundamental questions about physical, biological, ecological, social, cultural, economic, demographic, and political processes of coastal tourism are posed and answered in assessments, impact statements, profiles, and other products of natural, biological, and social scientists. With reference to the condition of the environment and society, the possibilities of coastal tourism development raise not only the question “What is?” but also questions about “What is ethical?,” “What is fair?,” and “What is beautiful?” As a result, analyses by professionals with backgrounds in the humanities and arts have proven to be useful in complementing those of scientists.

The need to formally study tourism is recognized more than ever in academe. Tourism research methods are under continual development in such fields as public affairs, business and marketing, architecture, urban planning and design,

political science, sociology, geography, cultural anthropology, marine affairs, and environmental studies (e.g., see Gunn 1979; Murphy 1985; Ritchie and Goeldner 1987; McIntosh and Goeldner 1990).

Tourism Broker Responsibilities

A second professional group made up of the different types of coastal tourism brokers will be counted upon heavily in the future to cooperate with one another. This can occur, for example, when investors and developers in the private sector coordinate goals and activities with those of government agencies and NGOs to make sustainable tourism a reality. Another kind of cooperation calls for tourism brokers to work effectively with government, business, and non-governmental organizations in other economic sectors. Better understandings of tourism-fishery interactions, tourism-aquaculture interactions, and tourism-ocean shipping interactions can lead to an improvement on single-sector governance with partially (or, under ideal conditions, fully) integrated coastal management.

In the aftermath of the terrorism attack on the World Trade Center in 2001, the responsibilities of tourism brokers have been enlarged. Brokers now must function not only as stewards of the coastal environment, businessmen, and representatives of constituencies, but also as protectors of residential and traveling publics.

Uncertainties generated by the terrorism of 2001 will change the ways in which coastal tourism is conducted in the United States and elsewhere. It has long been known that too much tourism can be bad by when it leads to degraded ecosystems and undesirable changes in quality of life. Now it is apparent that too little tourism can put entire coastal economies at risk. Declines in coastal tourism can create serious social problems in the same way declines in fishery resources can threaten livelihoods. Ultimately, coastal tourism and recreation destinations negatively affected by security-related changes in itineraries will become sustainable only to the extent brokers make tourism safe.

Individual Responsibilities

The discussion above has concerned the proactive roles researchers and brokers can play in promoting sustainable tourism and coastal development. To this must be added a comment about the personal responsibilities of tourists and locals to contribute toward sustainability in the coastal tourism systems which they visit or in which they live.

To a certain extent, the social role of the ethical tourist can be formulated to resemble that of the good citizen. Good citizens learn from their families and schools to reach their potential in society while knowing how to behave in socially appropriate ways. Using this template, tourists would be expected to behave in ecologically and culturally appropriate ways in the course of their domestic and international travel.

Ecotourism and ethnic tourism are two forms of tourism that have emerged to stress this self-conscious orientation. Many tourism brokers in business, in government, and in non-governmental organizations are now promoting the development of “best practices” and “tourism guidelines” to this end.

It is obvious that there are many benefits of coastal travel that accrue to the tourist. These are found in recreational, aesthetic, and educational activities. In exchange, the ethical tourist will strive to behave in a culturally and environmentally responsible manner. As this occurs, locals are given an added incentive to orient their conduct to the same ends. Improvements in the behavior of tourists and locals toward one another and toward the coastal environment will assist tourism providers and managers as they do their part to monitor and control tourism, and improve the tourism experience for all involved.

Concluding Remarks

Coastal tourism has demonstrated its considerable power to influence the fundamental configurations of coastlines and the social structures these support. Coastal tourism is sometimes found to be unfortunate in every respect. Coastal tourism can, however, be designed to improve the lives of tourists and those who are part of the tourism industry, conserve natural resources and protect the environment, and not offend locals. For this to occur, coastal tourism brokers – in government, business, and non-governmental organizations – will need to cooperate to insure that tourism is sustainable and safe.

It will also be necessary for tourists and locals to adopt “best practices” that underwrite cross-cultural communication and respect for the environment. In the eyes of many, it is time for all to abide by a coastal system *tourism ethic*. Such an ethic might reasonably incorporate Aldo Leopold’s (1949, pp. 224–225) famous caution about natural resource use based solely on economic self-interest:

[a] thing is right when it tends to preserve the integrity, stability, and beauty of the biotic community. It is wrong when it tends otherwise.

Through the implementation of responsible management, planning, and education policies – together with the diffusion of a tourism ethic – tourism and coastal development can be shaped to reflect the best tendencies of humanity.

Cross-References

- [Aquaculture](#)
- [Beach Use and Behaviors](#)
- [Carrying Capacity in Coastal Areas](#)
- [Coastal Management Practices](#)

- [Demography of Coastal Populations](#)
- [Economics of Beaches](#)
- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Marine Parks](#)
- [Surfing](#)
- [Tourism, Criteria for Coastal Sites](#)

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Tourism, Criteria for Coastal Sites

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Historically, seaside resorts date back to Roman times with a string of resorts along the Campanian littoral on the northern shore of the Bay of Naples (Turner and Ash 1976, p. 24). The modern seaside resorts had their origin in England and their early growth was attributed to the therapeutic value of sea-water drinking and bathing. From the mid-19th century, the era of railway saw the rapid growth of more seaside resorts, spreading to western Europe (Romeril 1984; Walton 1997). Partly as the result of the railway ending at about right angles to the coast, the European coastal resort has a basic T-shaped morphology, although factors varied. Usually, coastal resort morphology is dependent on site characteristics, tourist elements, and other urban functions (Pearce 1995, pp. 136–140; Nordstrom 2000, pp. 10–13).

This entry is concerned with such physical site factors as the “emphasis . . . on aspects of the physical implications of site selection such as coastal erosion rather than resort form . . . underlies the potential which this approach has to complement the more traditional resort morphologies” (Pearce 1995, p. 137). This approach is increasingly significant, as developing countries seek out appropriate beaches and islands for tourism development. On a worldwide scale, coastal tourism (see *Tourism and Coastal Development*) in the traditional Mediterranean area is complemented by the Caribbean area, South Pacific area, and Southeast Asia. Islands, especially *small islands (q.v.)*, are actively sought for tourism development. To the usual three “Ss” (sun, sea, and sand) for coastal tourism, one could add two more “Ss” (sunrise and sunset) if one were on a small island.

Coastal Site Criteria

Coastal tourism development involves the identification of a suitable site, developing the site by clearing, and providing access, accommodation, recreation facilities, and services for

the tourists (Ahmad 1982). Site criteria for coastal tourism are more than just association with white sand beaches and coral reefs. An analysis of the site criteria includes a wide variety of physical and other factors such as land-use, ownership, etc. The objective is to assess the site opportunities and constraints which have a significant implication for the resort entrance, backdrop, views, beaches and swimming areas, buildable area, vegetation, boating tours, fishing and diving opportunities, etc. In addition, various site planning considerations and development standards, including architectural, landscaping, and engineering design, have to be considered (Inskeep 1991, pp. 303–335).

Planners and architects usually acknowledge that two powerful forces make coastal tourism distinctive and have a bearing on site criteria. The first is the special amalgam where land meets the sea and the coastal site is a junction of landscape and seascape. The second is the linear character of the coast in which the natural resources for tourism are arranged in a linear fashion (Gunn 1988, p. 87).

The linearity of the coast presents special challenges for planning and tourism design. Generally, four zones with different characteristics can be identified with the more important criteria and implications for tourism development given in Table 1 (Gunn 1988, p. 88; Mieczkowski 1990, pp. 243–246). On the seaward side is the marine zone or neretic zone which is the ecological zone from the continental shelf to the beach. This area contains the marine life, reefs, and sandbars and is suitable for a number of marine-based activities. The beach consisting of the foreshore and back-shore is the most important of the four zones. It is used for many activities, especially if it is sufficiently wide and sandy. More specific requirements identified for beach resort development in this zone include beach protection and beach capacities (Baud-Bovy and Lawson 1998, pp. 71–72). The third zone is the shoreland which includes the dunes and is the area for camping, hiking, and other accommodation, food, shopping, and other service businesses. It serves as the visual linkage between land and sea. The most landward zone is the vicinage or hinterland which provides the setting for tourist business and vacation homes. It is often the zone of population and supporting services and is enhanced by variations in topography and vegetation. Its nearness and access to the sea is more important than the visual linkage.

An example of the application of the linear zones is the planning of development for southern Thai beaches facing the Andaman Sea (Tourism Authority of Thailand 1989). Three zones for development with 12 categories of landuse have been identified. The first two zones have specific widths. The beachfront area extends 300 m away from the beach; the interior area is 700 m wide and the hinterland is landward of the interior area.

As the coast is a basic component in coastal tourism, any potential resort site can be assessed by an initial

Tourism, Criteria for Coastal Sites, Table 1 Major criteria in coastal zones for tourism development

Zone ^a	Factor	Comments
Sub-tidal to offshore (Marine or neretic zone)	•Climate •Waves •Tides •Currents •Water temperature; clarity of water •Biodiversity, e.g., marine life, corals, seaweeds	•Physical conditions determine type, extent and seasonality of many recreational activities, e.g., swimming, water skiing, surfing, sailing, boating, travel to nearby island •Biodiversity presents additional attractions for recreational use, e.g., snorkeling, scuba diving, fishing •Free of pollution
Intertidal-nearshore (Beach)	•Beaches: width, gradient, material size, color •Risks from tidal movements •Potential erosion •Public access •Shore platforms: width, access •Wetlands: extent, access	•Physical properties of beach and coast influence type and extent of recreational activities •Beach capacity as useful management tool •Soft protection measures are preferred if need arises •Environmental guidelines, especially for wetlands
Backshore (Shoreland)	•Area •Views •Geomorphology (cliffs, dunes, wetlands) •Coastal vegetation •Microclimate •Scope for improvement •Access, e.g., roads	•Location of various tourist accommodation and service businesses •Proper setback, conditions for use •Preserve view (visual linkage is important) •Maximize specific advantages of geomorphological features •Dunes normally preserved as defense line and for selective uses •Good views from cliffs, headlands •Phased development; minimize degradation •Improvement, e.g., drainage
Onshore (Hinterland or vicinage)	•Topography •Vegetation •Existing development, e.g., population, supporting services	•Provides setting or backdrop •Separate planning zone •Access to sea is important

^aTerms in parentheses are used by Gunn 1988 and Mieczkowski 1990

Sources: Compiled from Baud-Bovy and Lawson 1998; Gunn 1988; Mieczkowski 1990; Viles and Spencer 1995; Wong 1991

understanding of its coastal geomorphology. As coasts differ widely, each type of site has implications for coastal tourism. For example, the potential of a specific coastal type or landform, such as, coral reef, coastal dune, sand spit, river-mouth barrier, rocky headland, can be known. Also, specific advantages and disadvantages of each landform, coastal type, or ecosystem have their influence on the choice of resort sites and also influence the pattern and development of resorts. Coastal geomorphology also helps to reduce negative impacts, protect valuable habitats and provide valuable information for subsequent alteration to the coastal environment, for example, changes to the drainage and water bodies (Wong 1999, 2000).

More important and often underestimated, geomorphology also takes note of the seasonal factor which can have a marked impact on the coast. For example, in Southeast Asia and Indian Ocean islands, a strong seasonality prevails in the coastal environment. Beaches undergo accretion/erosion cycles and during the onshore wind season, wave action can reach a higher level or further inland. With reversals in wind, wave action and currents, sand movement varies in the offshore–onshore direction and alongshore direction. The seasonal site features include beach morphological changes, a backshore with two berms, nearshore topographic changes, changing stream mouths, flooding, shifts in beach vegetation belts, etc. In particular, the analysis of the seasonal factor

provides some idea of the potential hazards in various coastal zones. The seasonality arising from waves, wind, and tides, and the potential erosion and pollution are among the coastal hazards considered in environmental planning for site development (Beer 1990, pp. 63–64).

Besides geomorphological criteria associated with tourist sites, other criteria in planning and development of resorts also have a strong physical base. These include adequate access, *setbacks* (*q.v.*), and EIA (environment impact assessment) before construction and carrying capacity. Pearce and Kirk (1986) have suggested specific types of carrying capacity that are closely associated with the linear zones of the coast. One recent development has been the application of a single criterion in the form of an ecolabel to assess the suitability of resort beaches. An example is the Blue Flag award for European beach resorts that focuses on *water quality* (*q.v.*), beach management, and safety (see *life saving and beach safety*) (UNEP/WTO/FEEE 1996).

Coastal Site Classification

Of various types of *classification of coasts* (*q.v.*) (Bird 2000) none is suitable for determining site criteria for coastal tourism. Classifications with an emphasis on coastal processes can be useful for resort sites focusing on recreation, such as

surfing (q.v.). For example, Bird (1993) identified various types of surf reflecting geomorphological and oceanic factors. Since the mid-1980s, the beach morphodynamics model with its identification and explanation of beach hazards, such as steep beaches and rip currents, is particularly useful for resort beaches in Australia (Short 1999).

Generally, physical or morphological types of coastal classification are more useful for recreation/tourism planning, development, and management. For recreational purposes in the temperate countries, coastal landforms are classified as sand and shingle beaches, tidal forms (mudflats, salt marshes), estuaries, cliffs, and shore platforms (Pickering 1996). Defert (1966 in Mieczkowski 1990, pp. 247–248) provided one of the earliest classifications for coastal resort development, in which four types of coasts were identified:

1. *Oceanic: continuous and linear*: Straight oceanic beaches with tourism facilities following straight sandy coasts.
2. *Oceanic: discontinuous and concentrated*: Bays alternate with promontories and peninsulas with tourism located in the bays.
3. *Mediterranean: continuous and linear*: Wide and gentle beaches prevail as a result of the absence of tides or very small tides.
4. *Mediterranean: discontinuous and concentrated*: This consists of two types (1) wide bays bordered by promontories or capes, and (2) coves with small beaches.

In a situation where the site is clearly restricted or limited to one coastal type, it is possible to have further categorization in terms of criteria other than physical. For example, in the Ko Samui/Surat Thani region located in the Gulf of Thailand, the beaches are further identified for tourism development as follows: (1) conservation beaches, where tourism activities are not allowed; (2) nature-oriented development beaches where activities and services are permitted to a certain extent; and (3) progressive development beaches where activities and services can be developed to meet international standard (Tourism Authority of Thailand 1985).

For the east coast of Peninsular Malaysia which is exposed to the northeast monsoon, resort sites are identified for tourism development with a strong consideration given to the seasonal factor. Four major types of sites are identified on the mainland coasts (Tourist Development Corporation 1979).

1. Beach front site. This is backed by relatively flat land, sandy soils, and coconut groves and exposed to the sea.
2. Site oriented to both ocean and river or brackish lagoon. The topography is generally flat or rolling gently with coconuts and other vegetation.
3. Site situated adjacent to a substantial headland promontory with sufficient flat land for development. The beach may

be interrupted by large boulders. Hill and weather patterns influence architectural design and site plan.

4. Site on hillside or hilltop with panoramic views located on headlands adjacent to good beaches. Weather and wind can be either greater or sheltered depending on the position of the development area. More constraints are placed on resort design.

These four types of sites can also be found on the islands which may provide a change of wind and sun orientation not possible with the mainland coastal site. However, access to islands is an important factor in construction and operation.

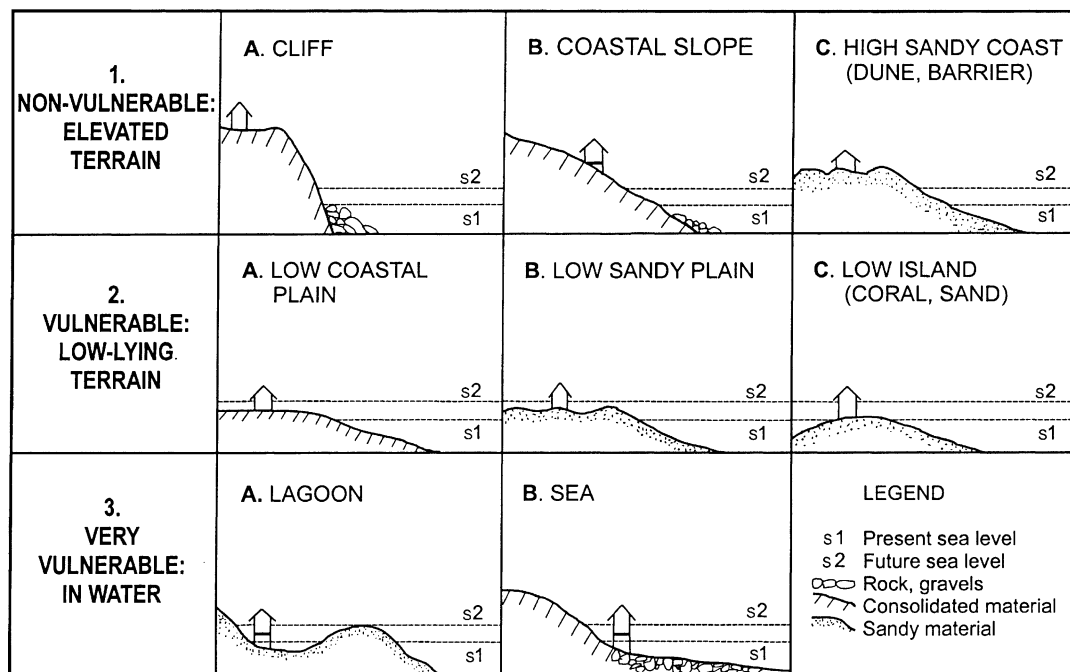
Based mainly on field examples from Southeast Asia and Indian Ocean islands a classification of resort sites is proposed for tropical coasts (Table 2). Where possible, the type of coast or coastal form is followed by an identification of specific suitable sites. Mainland *sandy coasts* (q.v.) are the most important for resort development, with various suitable sites depending on the planform and degree of protection offered by rock headlands. Small islands, especially *coral reef islands* (q.v.), are also favored for resort sites but they have fragile environments. Mangrove coasts hold a potential although the vegetation is being cleared cut down for other uses. Dunes are limited on tropical coasts and lagoons are actively used for local fishing. Stream mouths can be integrated into the resort site but requires adequate setback from possible flooding. Except for selective headlands, rock coasts (see *rocky coasts*) are underutilized for resorts. High rock coasts offer good views but require careful design and engineering. Low limestone coasts can be altered variously for resort sites, as evident on Mactan Island, where the modification can be limited or minimal (e.g., short seawalls, stone bunds), localized (e.g., groins, breakwaters, artificial islands) or effective (e.g., artificial beaches, selective excavation of rock to form small bays) (Wong 1999).

With a *sea-level rise* (q.v.) in the future, many resorts on low-lying sandy plains or beach-ridge plains will be threatened as the coastline is cut back by the rising sea. For small islands dependent on tourism, the situation can be serious. Perhaps the most badly hit would be the island resorts of the Maldives, where the height of cays are within a couple of meters of the sea level and the entire island resort industry can be wiped out (Domroes 2001). Mauritius has projected its loss of tourist beaches and in the major tourist area of Flic-en-Flac, an estimated 26,000 m² of beaches could be flooded by a meter sea-level rise (Mauritius National Climate Committee 1998, p. 45). The response to the threat of rising sea level can be adaptive strategies classified as (managed) retreat, accommodation, and protection. Although various hard structural and soft structural options are considered for protection, new methodologies are being sought to estimate coastal vulnerability and resilience to the sealevel rise. For some low islands in the Pacific, coastal types can be used to assess vulnerability

Tourism, Criteria for Coastal Sites, Table 2 Types of resort sites on tropical coasts

Coastal type	Form/feature	Resort sites
A. Rock coast	1. Cliff 2. High headland	1. Good view; generally exposed; adequate setback necessary 2. Good view; stability is crucial; can be sheltered or exposed. Requires compact design solutions
B. Mainland beach Coast	1. Linear 2. Crescentic bay 3. J-shaped bay (zetaform) 4. U-shaped bay 5. Cuspate foreland 6. Spit 7. River mouth associated with above	1. Large area available; can be exposed 2. Can be at head of bay but has higher wave energy; decreasing wave energy towards the limbs of bay 3. Sheltered in upcoast curve; increasing exposure to downcoast straight sector 4. Can be at head of bay; often decreasing sand toward the limbs of bay 5. Only on large foreland; need to determine stability of convexity 6. Only on large stable spits; river can be integrated into design; adequate setback from channel to avoid flooding 7. To be avoided because of rapid changes and seasonal closure of mouths, especially of small streams
C. Barrier coast	With/without dunes	Can be on barrier; but behind active dunes. Maintain seaward line of dunes as buffer zone. Lagoon can be integrated into design
D. Small island	As in A and B, where applicable.	Wider choice on many types of beaches on sheltered side. Limited beaches on exposed side. Access is important, especially during seasonal weather
E. Coral island	1. With/without lagoon 2. Cay	1. Adequate setback required avoid destruction of reefs 2. Adequate setback required. Strong seasonal changes; no dredging or structures interfering with sand movement
F. Mangrove coast		On piles to minimize impact on ecosystem. Tidal range can be critical factor in accessibility and jetty length
G. Developed coast	1. Original sandy coast 2. Original low rock coast	1. Sufficient setback from protection measures. Beach nourishment required 2. Selective removal or rock to create bays, artificial beaches. Beach nourishment required. Also raised sandy platform with coastal protection

Sources: Revised from Wong 1990, 1991, 1999, 2000

**Tourism, Criteria for Coastal Sites, Fig. 1** Types of resort sites affected by a sea level rise. (Redrawn from Teh 2000).

if further information is available. For example, while the upper shore can be of sand or shingle or a mixture of both, the lower shore can be a conglomerate platform, ramp out-crop, beachrock, mangroves, or seawall. Such details would

be useful for assessing the vulnerability of a resort coast (Mimura and Harasawa 2000).

Except for some island resorts, the vulnerability of resort sites to a sealevel rise has not been fully appreciated. In

Malaysia, Teh (2000) has classified coastal resort sites according to their vulnerability to inundation arising from a sea-level rise. The sites can be (1) non-vulnerable (on a coastal slope or a high beach-ridge plain), (2) vulnerable (on a low coastal plain, low beach-ridge plain, or low island of coral, sand, or mud), and (3) highly vulnerable (in a lagoon or built over the sea). An additional coastal type (cliff) is added to this classification to cover other coastal types in Southeast Asia and the Indian Ocean islands (Fig. 1). Compared with the classification of coasts for resort sites, this classification emphasizes the sea level relative to the two-dimensional coastal profile and much of the variety of coastal types and landforms is consequently made redundant. Nevertheless, it is an initial step in providing some idea of the vulnerability of resort sites to a sea-level rise.

Conclusion

Although many factors have to be considered in coastal sites for tourism, geomorphology remains basic as long as the coast is a necessary resource for tourism. Site analysis has to assess various types of coasts and its various zones. As coastal tourism caters to a widening demand, coastal sites for resorts are also being assessed for their suitability for other demands such as golf courses, marinas, oyster culture, and other related types of development. In the future, resort sites are likely to extend beyond the usual sandy coast and rock headland and include more adaptive use of the mangrove coast which has a potential for coastal ecotourism. Coastal sites are also likely to incorporate better technology not only for resort construction and infrastructure but also for coastal protection against erosion and beach management.

Cross-References

- Bay Beaches
- Changing Sea Levels
- Clifed Coasts
- Coral Reef Islands
- Headland-Bay Beach
- Holocene Coastal Geomorphology
- Human Impact on Coasts
- Indian Ocean Islands, Coastal Ecology and Geomorphology
- Lifesaving and Beach Safety
- Natural Hazards
- Pacific Ocean Islands, Coastal Geomorphology
- Rating Beaches
- Rock Coast Processes
- Sandy Coasts
- Sea-Level and Climate Change

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Tourist Beaches

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Definition

Tourist beaches are a socioecological resource located mainly on coastal destinations, which is developed and managed for the primary purpose of attracting tourists interested in sun, sea, and sand activities.

Fundamental Concepts from Tourism Theory

The most common definition provided by UNWTO states that tourism is “traveling to and staying in places outside their usual environment for not more than one consecutive year for leisure, business and other purposes” (UN and UNWTO 2010). This definition is open to interpretations, as the distance travelled and the minimal duration are not mentioned. Similarly, the definition provided by Leiper (1990) regards “tourists” as people away from home to the extent that their behavior is motivated by leisure-related factors. Therefore, the functional definition of “tourism beaches” blurs tourism, recreation, leisure, and day visiting to such an extent that one is indistinguishable from the others as they all share the same resources, use the same facilities, compete for the same consumer, exert similar impacts on the beach environment, and do not specify overnight stay. The only distinction in the definition of a tourist beach is that the activity takes place within the beach boundaries.

In general terms, “tourism beaches” can be defined through the concept of “tourist attraction.” Established definitions of a visitor attraction include Hu and Wall (2005: 619)

“a permanent resource, either natural or human-made, which is developed and managed for the primary purpose of attracting visitors”. The authors stress additional characteristics of an attraction that must be open to the public, without prior booking, and should be capable of attracting day visitors or tourists who stay overnight in the area as well as local residents, and finally, the attraction must be under a single management. All these characteristics in theory apply to urban, resort, and protected area beaches open to the public, although in practice, many beaches are not managed by the authorities officially responsible for their management.

An attraction is any place that consists of artificial as well as natural features that geographically draws people to provide them with a recreational experience and is composed of various components including tourism activities, local scenery, and services (Gunn 1988). Ritchie and Zinns (1978) expand that classification including natural beauty and climate; culture and social characteristics; sport, recreation, and educational facilities; shopping and commercial facilities; infrastructure; price levels; attitudes toward tourists; and accessibility. Tourism beaches can, but do not have to, include a range of supporting facilities and services that are located outside of the limits of the beach area but have a direct influence on the experience of beachgoers such as accommodation, transportation, food, entertainment, and recreation which are required by visitors. Although tourist beaches can be classified in terms of characteristics that apply to any tourist attraction, some of them are more specific to the beach and water environment (e.g., bathing, swimming, water sports, and appreciation of the coastal/marine ecosystem). In addition, the importance given to specific characteristics of tourist beaches as an attraction is different than that given to nonbeach attractions, it varies from culture to culture and it depends on the type of the beach and type of tourist it attracts (Botero et al. 2013). These characteristics compose the image that is the most important aspect of a tourist attraction from a marketing point of view (Lew 1987).

Tourism beaches can manifest at a variety of scales. Depending on the visitor's perception, they can draw people locally, regionally, nationally, and internationally. “Particular sights, sites, objects, or events will often be the focus of experiences for tourists and local residents, but for the former group, they are an object of tourist attraction and for the latter an object of leisure-related experience in the home locality” (Leiper 1990: 375). Focusing only on the beach as an attraction to tourists, various beach tourism models can be identified, according to the facilities offered and the profile of beachgoers (Fig. 1). The classical model is represented by mass tourism and the so-called 3S tourism (“sun, sea, and sand”), which is characterized by low beach user densities (<5 m²/user) and high level of tourism infrastructure. However, during the last few decades, new models had emerged, stemming from the desire of tourists to pursue more exclusive



Tourist Beaches, Fig. 1 Examples of beach tourism models in Colombia: Mass tourism: (a) *Bocagrande*, (b) *El Rodadero*; Sport tourism: (c) *Guachaca*; Cultural tourism: (d) *Cabo de La Vela*

or sustainable experiences. As a result, each day it is more common to discover luxury beach clubs with extremely high prices in order to use their facilities (>500 euros/day for renting an umbrella), and, beaches are focusing on specific sports such as wave-surfing or kitesurfing. Some remote beaches in natural areas have several restrictions on the majority of tourist activities. In consequence, innovation of facilities and services on beaches is allowing new “beach tourism complexes,” such as beach resorts, beach hostels, and varied types of beach clubs (Fig. 2).

Social and Economic Importance of Beaches

Beaches are commonly considered as other coastal ecosystems, framed within biotic and abiotic relations. However, with the advancement of the years and the enrichment of scientific methods with new paradigms, this perspective of beaches as merely ecosystems has changed to a wider, more

complex understanding: the beach as a socioecological system (Botero et al. 2014). Nowadays, it is recognized that beaches have a high interest for society in terms of economic and cultural assets. Hundreds of millions of tourists cross the world looking for sandy beaches, from very natural places to all-inclusive resorts, but in all cases expending money through several tourist brokers (see ► “[Tourism and Coastal Development](#)”). At the same time, the natural and cultural heritage is promoted as an attraction at a certain location, but impacted with tourism development. Therefore, an institutional framework and mechanism may be implemented to balance the interest to attract tourists and prevent degradation over the heritage that attracts them.

From this socioecological perspective, beaches have several functions for society. Nowadays, the main social function of beaches is linked with their capacity to receive hundreds and even thousands of visitors looking for leisure activities in a pleasant place. This recreational function allows local inhabitants and foreign visitors to enjoy the coast,



Tourist Beaches, Fig. 2 Examples of beach tourism complexes: Beach resorts: (a) Guardalavaca (Cuba), (b) Huatulco (Mexico); Beach hostels: (c) Dibulla (Colombia); Beach club: (d) Cozumel (Mexico)

consolidating a sense of identity with some specific beaches that are often visited by them. As a result, it is common that civil society organizations defend one or various beaches as their “common heritage” because they consider these coastal sectors a part of their daily reality. These two social functions, leisure and identity, are closely linked with the right for public access to beaches, which are under legal protection in the majority of countries around the world. Beaches are social places used by local people to meet their partners, interact with nature, enjoy open views, admire the coastal scenery, or do exercise either by walking or by swimming. Finally, beaches have a deep function to satisfy human needs of affection, understanding, participation, leisure, and identity, within a framework of the Human Scale Development (Max-Neef 1993).

In addition to the importance for social interactions, beaches have an increasing economic importance. Tourism became one of the largest and fastest-growing economic sectors in the world, rising from 25 million tourists in 1950 to 1,186 million in 2015 (UNWTO 2016); within this trend, the coastal areas have been a strong draw (UNWTO 2013). Nevertheless, the economic importance of tourism on beaches is more than merely macroeconomic data. Firstly, the sand area of beaches has a very high monetary value, as has been demonstrated by several authors, such as Victor Yepes, who

estimates the profits generated by beaches in the Valencia region (Spain) to be as high as 700 euros per square meter per year (Yepes Piqueras 2004) or James R. Houston who estimates the annual federal revenues from beach tourism in the United States to be 25 billion dollars (Houston 2013). Secondly, beaches have an ecosystem value represented in those services offered to human beings, as described by the Millennium Ecosystem Assessment (MEA 2003) and applied by several studies (e.g., Castaño-Isaza et al. 2015); this ecosystem value is often measured through economic valuation of nonmarket resource methods, such as Willingness To Pay or Travel Cost. Thirdly, the cultural value of beaches has an increased interest recently, based on their material and immaterial elements, such as scenery features, historical events, and ceremonial rites of local and indigenous communities. These three kinds of values are continuously interacting among them, although the monetary value is highly dependent on the ecosystem and cultural values, raising the obligation to seek a sustainable tourism of sun, sand, and sea products.

Services and Facilities in Tourist Beaches

Tourist beaches must be managed in a way that visitors comfortably enjoy their stay and reduce their environmental

impact to a minimum. The list of services and facilities to implement at a beach greatly depend on the kind of tourists who frequent the place and their dynamic preferences (Botero et al. 2013). Therefore, it is recommended to start from a general framework applied to the majority of beaches, and afterward customize each element to the particular requirements of users. In this way, Williams (2011) established the so-called “Big Fives,” related to the same elements that define preferences of almost any beachgoer in the world: Water Quality, Litter, Facilities, Scenery, and Safety. Therefore, a tourist beach manager must lead his actions to guarantee public and private/commercial services that satisfy the comfort and safety of visitors, and to provide enough facilities and infrastructure to support the tourist activity.

Main public services to maintain at a tourist beach are related to cleaning and safety. The former is in charge of maintaining the beach litter-free, with a focus on sand surfaces. Cleaning the beaches is a service usually done by local municipalities, although it should be reinforced with the tourism brokers located on the beach (e.g., restaurants, stores, shade tent/parasol providers). This cleaning service could be done through manual or mechanical methods, although it is recommended to mix both techniques (see “Cleaning beaches”). On the other side, safety service is composed of three elements: lifesaving, first-aid assistance, and hazard signage. In several cases, voluntary organizations such as the International Red Cross or the Surf Riders are in charge of lifeguard and medical assistance. In other cases, municipalities or resorts contract private lifeguards and medical

personnel on their own (see ► “Lifesaving and Beach Safety”).

Private services on the beach strongly depend on market dynamics. In a broader sense, services such as food and beverages, sports, spa, and ceremonies, among others are offered on a tourist beach. Almost all tourist beaches in the world have restaurants or bars in which visitors can buy a range of products from snacks and refreshment beverages to lunch menus, sometimes organized according to zoning and the carrying capacity of the beach. Moreover, peddlers are very common in many countries, selling a variety of articles from fresh food to clothes and toys. Similarly, a service often offered on beaches is equipment rental for nautical sports or beach games, which usually require facilities directly on the beach and exclusive areas on sand and water to practice. Another commercial service offered on several beaches (e.g., South East Asia, Europe, Latin America) is a tent installed with one or more cots to bring massages and stimulating treatments to clients, taking advantage of the relaxed-friendly atmosphere of the beach. The last commercial service to mention is related to religious and family ceremonies, which each day are becoming more popular all around the world, using specific sectors of the beach for weddings, rites, and parties (Fig. 3). It should be mentioned that this kind of service is not free of conflicts with other beach users and uses, as a consequence of exclusion of beachgoers from the ceremony or party.

Facilities and infrastructure on beaches are diverse (Fig. 4). Initially, any tourist beach must have sanitary facilities, such as toilets or showers, which are open for free or at a low cost

Tourist Beaches, Fig. 3 Private groin for religious ceremonies (Nuevo Vallarta, México)





Tourist Beaches, Fig. 4 Examples of facilities for tourist beaches in Spain ((a) Toilettes, (b) Lifesaving towers, (c) Beach bar; (d) Floatable wheelchair)

for visitors who need to use them during their stay on the beach. If pets and other kinds of animals are allowed on the beach, a dog's area is highly recommended. At beaches with an interest in universal accessibility, it is relevant to install ramps, wood-paths, and signage for disabled people, in addition to elements such as floatable wheelchairs. On urban beaches with a high number of visitors and built environment, the presence of pedestrian crossings, sidewalks, and signage are also necessary to allow a safe experience for tourists and locals. In terms of safety, lifeguard towers are the most common facilities, although they can vary from a small chair on the sand to high towers to oversee the whole beach. Finally, there are several private facilities on beaches, which are related to food and beverages (bars, restaurants), recreation (sport areas, children's parks, a library) or sunbathing (chairs, umbrellas); all of them occupy a place on the beach and must be carefully planned and designed to reduce conflicts with other users on the beach.

Differences Between Tourist, Popular and Natural Beaches

The vast majority of beaches nowadays are used for tourist purposes, but there is little scientific knowledge about socio-ecological interactions that take place on them. In general terms, all beaches are classified as tourist or natural, although the latter receive a few kinds of visitors (e.g., park rangers or scientists). However, between these two wide categories of beaches, a new type is emerging adding to the general typology: the "popular beaches." These kinds of beaches are those usually visited by local residents of the city or village in which the beach is located, and they are aimed just for recreational purposes. From a superficial view, these beaches are the same as tourist ones, with plenty of people taking sunbaths and swimming, but on a deeper level of analysis, the beachgoers' motivation and activities could be different, which also implies a need for specific management strategies. As a result,

there is furious competition every day (e.g., space on the beach, kind of services, prizes) between local residents who recognize the beach as their frequent public space for leisure, and tourists who travelled hundreds or thousands of kilometers attracted by this coastal location.

In order to characterize this phenomenon, some differential factors could be defined. Initially, the intensity of use has a different pattern, while local residents visit beaches on weekends, early in the morning or at the end of the day, tourists use the beach all-day long and their visits are concentrated in summer seasons, whether it is a workday or not. Secondly, local residents visit almost the same beach or few beaches throughout their lives, increasing a sense of identity and care for this coastal space. On the contrary, many tourists visit the beach to experience novelty, being less common that they return to the same place. Thirdly, the purpose of visiting for local residents is linked to social interaction with their peers, which are their friends, relatives, or colleagues; meanwhile, for tourists the main purpose is to discover a new place, new experiences, and even new people, who perhaps they will not see again in their lifetimes. Nevertheless, those visitors with a second house on coastal areas are a particular case, between locals and tourists, because they return to the same areas every season, meet the same people and have an identity with some beaches, but they rarely visit the beach other than on these dates.

As a consequence, beach management is deeply affected according to the type of beach, and strongly reinforced by neoliberal policies, which focus all attention on economic issues, and despise natural and cultural values. Therefore, management actions at tourist beaches should have economic support from the local and national governments, with centralized offices which coordinate actions and promote advantageous agreements with the private sector; on the opposite end, popular beaches have disperse management actions, done by several and uncoordinated agencies, commonly scarce of funding and technical teams. This unequal scenario is partially in balance with the role of civil society, represented by Non-Governmental Organizations, scientists, and local groups, which defend the protection of natural and cultural values of their beloved beaches with few resources but persistent actions; notwithstanding, it already happens in very small proportions at popular beaches in the world. Finally, as an example of a very wide problem pointed out by thousands of articles, actions for adapting to coastal erosion are highly concentrated on tourist beaches, motivated by the economic importance of these areas; meanwhile, other beaches outside the eyes of tourist planners are the only interest of local residents worried for their houses or frequent free-time places.

Last, but not least, some common characteristics of tourist and popular beaches must be pointed out in order to avoid misunderstandings that confuse future management. Firstly, the environmental quality status of both type of beaches used

to be the same, which means that a tourist beach is neither cleaner or nor healthier than a popular one because it depends more on the environmental attitude of beachgoers, which don't depend strictly on the origin of them. Secondly, tourist or popular beaches are not dependent on the vicinity to urban areas. There are consolidated tourist beaches within cities, such as Copacabana in Rio de Janeiro or Bristol in Mar del Plata, and thousands of resort beaches in remote areas, such as Cayo Coco in Cuba or Albarella in Italy; something similar occurs with popular beaches, which are located in a residential neighborhood or in a small village in the countryside. Thirdly, beachgoers' activities are almost the same in both types of beaches. In general terms, sunbathing, swimming, and sand games are done by any kind of visitor of the beach, with no regard to if they are local or foreign, with the only difference being the frequency and the duration of the stay highlighted below. Finally, basic services and facilities should and must be the same in tourist and popular beaches. Cleaning, signaling, and lifesaving services, as well as sanitary or accessibility facilities, are the minimum requirements on a beach to receive visitors, whether the main use is done by tourists or local residents.

Summary

Beaches have relevant social and economic functions within coastal destinations, even in high urbanized cities as well as in remote villages. As a consequence of the very fast growth of the tourism sector around the world, beaches are commonly considered a "tourist product" and their essence as socio-ecological systems is being largely forgotten. Therefore, "tourist beach" definition brings the concept of two other kinds of beaches: natural or popular. Stemming from the theory of tourism and beach management, the concept of a tourist beach is discussed through comparison with the emerged concept of "popular beaches" in order to suggest appropriate factors to differentiate them and promote better decision making.

Cross-References

- [Beach Awards and Certifications](#)
- [Beach Management Tools](#)
- [Beach Use and Behaviors](#)
- [Carrying Capacity in Coastal Areas](#)
- [Coastal Management Practices](#)
- [Coastal Scenery](#)
- [Economics of Beaches](#)
- [Rating Beaches](#)
- [Tourism and Coastal Development](#)
- [Tourism, Criteria for Coastal Sites](#)

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Tracers and Coarse Sediment

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Definition

Tracers are essentially sediment particles that can be easily identified within a large mass of sediment grains having different characteristics. The concept is widely used in sedimentary petrography, where particular mineral assemblages are inherited from the characteristics of the provenance basin. Another approach is to use sediment from the environment to be studied and tag it using an artificial agent (e.g., paint, radioactivity, RFID transponders).

Introduction

Since the early stages of research on sediment transport, it became evident that natural and artificial tracers had a high potential, both for qualitative and quantitative assessments. The first examples of tracing experiments using artificial materials date back to the beginning of the twentieth century, with the experiments of Richardson at Chesil Beach (UK) in 1902 and the reports of the Royal Commission on Coast Erosion published in 1907 (in Kidson and Carr 1971). Radioactive tracers were also at first very popular, but they never became widely used, since the marking technique is complex and the method has a considerable environmental impact. Although the last 30 years have seen the development of alternative methodologies for measuring sediment transport (see entry on Beach and Nearshore Instrumentation), it can be noted that tracers (especially fluorescent and RFID types) are the only technique that can be applied at a broad range of temporal and spatial scales.

Historical and Technical Development of the Tracer Technique

Fluorescent Tracers

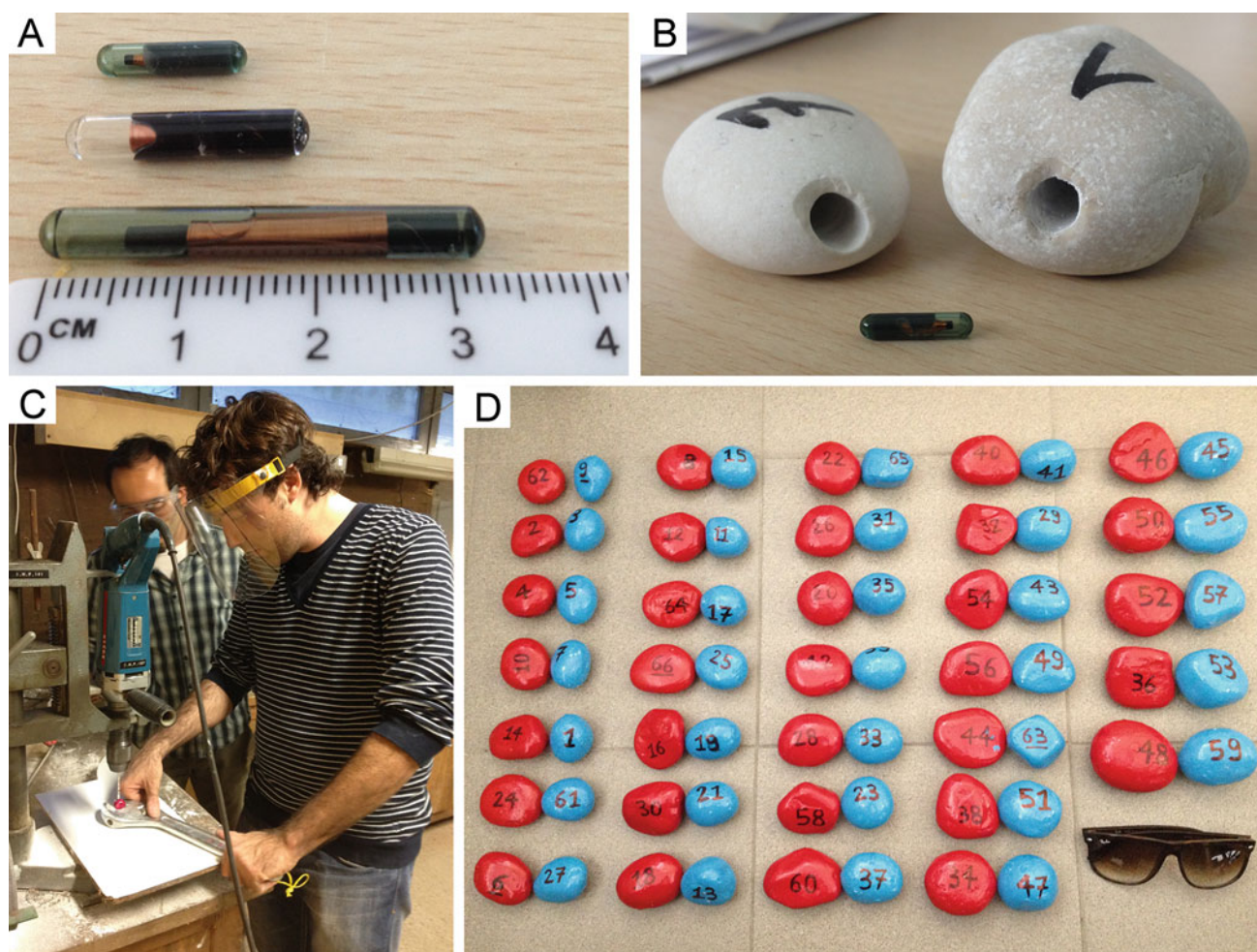
From the 1950s onwards, a group of investigators in the former Soviet Union started to use sands marked with fluorescent paints on a large scale (Zenkovich 1960). The methodology consisted in marking sand grains with a fine film of a colloid (agar-agar) containing fluorescent material. The film was resistant to water, chemical reactions, and mechanical abrasion. The marked sand was then dried and injected on the beach. Thereafter, at regular time intervals (e.g., from minutes to days), samples were collected at fixed distances from the injection point. In the following years, there was a boom in the use of fluorescent tracers for the study of sediment transport and investigators in the United States started to experiment with this technique. Yasso (1965) studied selective transport of different grain sizes in New Jersey; he used different colors of fluorescent paint, using a mixture of acrylic lacquer and beetle resin. In the 1970s, the application of tracers to measure longshore drift (see entry on Sediment Processes at Littoral Drift Barriers) became widespread. Empirical measurement of longshore transport to calibrate mathematical predictors for engineering studies was needed. Komar and Inman (1970) carried out several experiments on the western coast of the United States to build a database that eventually lead to their famous transport predictor. One of the largest experiments ever carried out on beaches is that of Chapman and Smith (1977) on the Gold Coast of Australia. Fifty tons of sand were marked and continuously injected using a dredge for about three weeks. Because the sand

had a high degree of natural luminescence, the use of fluorescent paints was ruled out and a blue dye was used. The fluorescent marking technique became again of interest to many European researchers in the 1990s, with the development of experiments in France, Belgium, Portugal, and Italy (e.g., Ciavola et al. 1998; Balouin et al. 2006; Silva et al. 2007). These experiments introduced several improvements to the older methodology, in particular regarding the automatic detection of marked sand grains and the study of tracer advection in three dimensions.

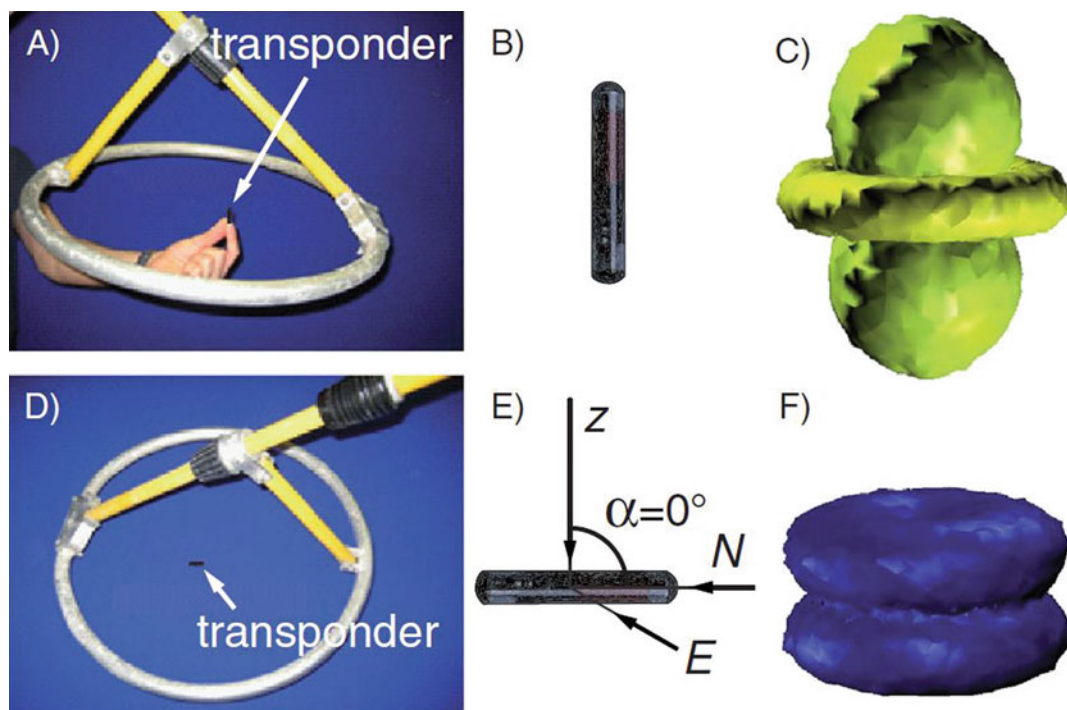
Radio Frequency Identification (RFID) Tracers

Coarse-grained beaches (see entries on Gravel Beaches, Boulder Beaches and Reflective Beaches) have recently become of notable interest to the coastal science community. Several tracer experiments have been performed during the last 10 years, searching for a constant improvement of the tagging technique as well the performance of equipment to better detect coarse particles. Coarse sediments were initially

marked with paint, labels, or with magnetic tags and then searched using a metal detector. The adoption by Allan et al. (2006) of Passive Integrated Technology (PIT) tags already used on rivers, along with RFID (Radio Frequency Identification) antennas (transmitters) and readers (receivers) started a brand new research wave within the tracing community. In the following years, the RFID technology was frequently adopted in tracing studies of coarse sediment and experienced an improved paired with drop in production cost of the tags. Up-to-date RFID technique consists of two parts: a transponder (or tag) which is the item inserted inside the sediment particle (Fig. 1a, b) and a reader (Fig. 5a) which is both the transmitter and the receiver and is the device that actually finds the marked sediment. RFID transponders can be passive or active, usually the passive ones are preferred for sediment tracing purposes because they are smaller (they do not need any battery) and have longer operational life (Chapuis et al. 2014). The power to activate them is supplied by the electromagnetic field generated by the reader's antenna (Benelli et al.



Tracers and Coarse Sediment, Fig. 1 (a) Cylinder glass transponders of different sizes; (b) Examples of drilled pebbles; (c) Drilling operations by means of a vertical driller; (d) Examples of tagged pebbles after sealing, painting and labeling



Tracers and Coarse Sediment, Fig. 2 Theoretical detection zones (by Chapuis et al. (2014), modified from Aquartis (2011)) according to the tag orientation compared to the loop antenna. (a), (b), and (c)

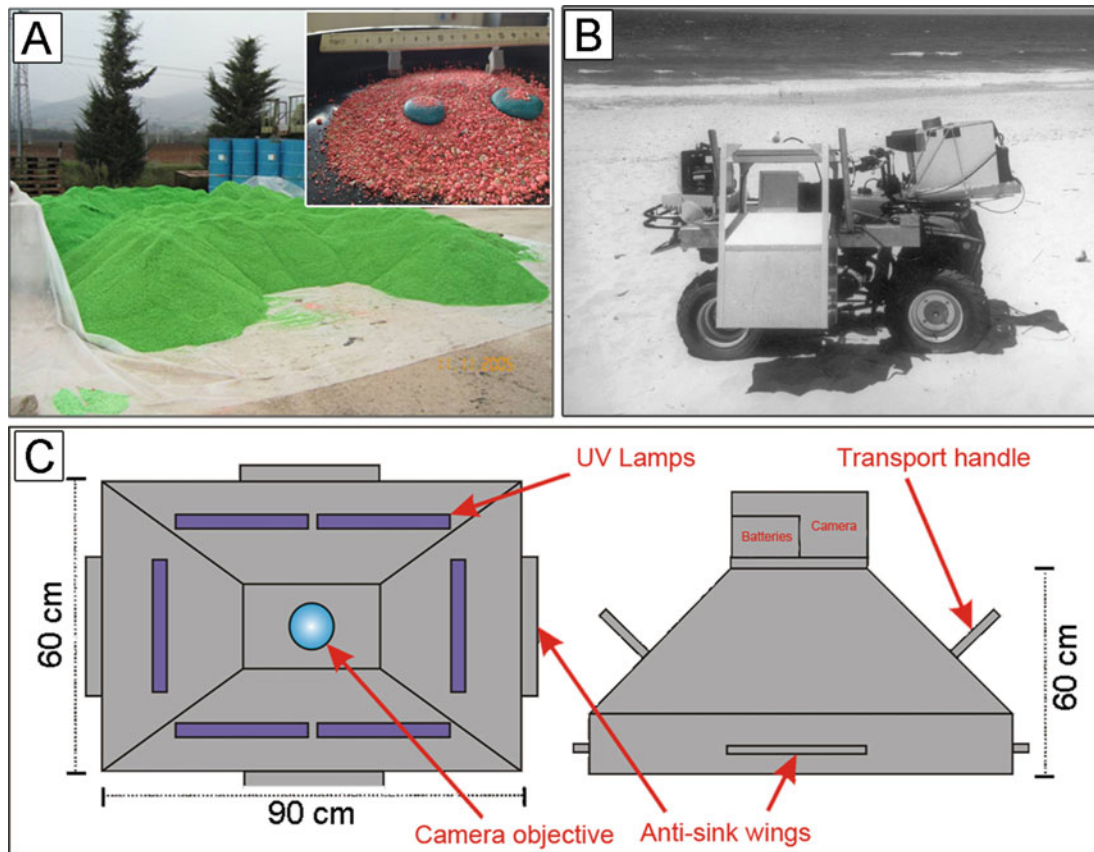
transponder is displayed vertically ($\alpha = 90^\circ$). (d), (e), and (f) transponder is displayed horizontally ($\alpha = 0^\circ$)

2012) and the detection zone of each tag is a 3D shape that varies depending on the transponder orientation related to the antenna position (Fig. 2). Furthermore, the transponder contains a small memory (i.e., semiconductor chip) with an alphanumeric code which allows the univocal identification. Transponders also differ from each other in terms of size, shape (cylinders and discs are more frequently used), and material (i.e., ABS, PVC, EPR, nylon, glass, etc.). Generally, larger transponders have also larger read ranges, but smaller sizes are crucial for studies on finer sediment sizes. Transponders can also be divided in terms of operational frequency in low frequency (LF), high frequency (HF), ultra-high frequency (UHF), and microwave: usually LF tags (generally between 135 and 125 kHz) are preferred for coastal or river studies since they can work also underwater (Benelli et al. 2012; Chapuis et al. 2014). The reading range, which is the maximum linear extent of the detection zone, varies depending on the type of tag (material, size, and shape), tag orientation (perpendicular or parallel to the reader antenna), type of antenna (size diameter), and media where the marked sediment is inserted into (salt water, still water, or air) but is poorly modified by burial of other sediments. In order to use the RFID technique on microtidal beaches, Bertoni et al. (2010) improved the RFID antenna to also work underwater. Reviews of the technical performances of passive LF RFID tags can be found in Benelli et al. (2012) and Chapuis et al. (2014).

Essential Concepts for Applying the Fluorescent and RFID Tracer Methods

Preparation of the Tracers and Assumptions

The fluorescent tracer method must fulfill some basic assumptions: the marked sands should have a hydraulic behavior and particle characteristics comparable to the unmarked ones; advection of the tracers should be prevalent over diffusion and dispersion once injection takes place; the transport system must be in equilibrium when sampling is carried out. Madsen (1987) presents an exhaustive mathematical description of the terms advection, diffusion, and dispersion. Different paint colors can be used to tag various grain fractions to study differential transport (Fig. 3a) and the sensitivity of the technique is on the order of 1 ppm. An essential factor is the quantity of tracer sand to be produced. For large quantities (e.g., 1,000 kg or more), industrial preparation is probably the only feasible method. If sand is collected from the local beach, a composite sample could be obtained by mixing equal subsamples collected on the lower-, mid-, and upper-shoreface. It is important to carry out grain size analyses of each subenvironment and of the composite distribution, to compare the effects of mixing. Care should be taken to avoid local sedimentary effects due, for example, to the presence of large-scale bedforms or moribund beach cusps. The sand is then washed with freshwater to remove the salt and then dried. Regarding the marking procedure, a number of



Tracers and Coarse Sediment, Fig. 3 (a) Examples of fluorescent tracers; (b) The SADAM system of Pinto et al. (1994) for in situ mapping of surface dispersal of fluorescent grains. It consists of an ATV, a dark

camera with UV light, and computer with image processing software; (c) An example of design of a dark camera with UV lamps

different types of resins and paints can be used. The main requirement for the paint is to be resistant to chemical and mechanical abrasion caused by seawater. Initially, substances such as seaweed glue, bone-glue, gum, and starch were used, but many authors later switched to acrylic paints, which offer better resistance to abrasion. In recent experiments, fluorescent paint soluble in toluene has been widely used. The next step is to dry the sand, minimizing the effects of aggregation. As far as the paint is maintained as thin as possible, aggregation should generally be negligible, especially if the sand is rapidly dried in the open air under sunlight. Even in the case of aggregate formation, if the sand is sieved with appropriate meshes after tagging, the mode of the populations should remain the same.

RFID tracers are prepared by drilling each pebble to create a suitable hole to accommodate the tag inside, then the hole has to be sealed with a waterproof resin. Further, one can consider to paint and label the marked pebbles for a quicker recognition on the beach (Fig. 1). Pebbles to be tagged can be randomly collected from the beach and selected according to the particle characteristic needed to be studied (shape, size, weight, etc.) or according to the typical population located on the different parts of the beach (storm berm, fair-weather

berm, swash zone, step zone; see entry on Gravel Beaches). The only limitation is the size, which has to be coarse enough to be drilled. The smaller transponder currently available is a cylinder glass tag with a 3 mm diameter and 13 mm long: this is suitable for a particle size ranging from -4 to -4.5 phi (16–24 mm, medium gravel a.k.a. pebbles). The tagging operation can reduce the particle weight about only 2 or 3 g (Grottoli et al. 2015) that is negligible considering the size of pebbles. At the end of the preparation, each marked sediment needs to be weighted and measured along the three axes in order to quantify, after recovery, the physical modifications occurred (e.g., abrasion). For both methods (fluorescent and RFID), the best balance between handling during tagging and percentage of recovery during the experiment should be met.

Injection of the Tracers into the Transport System

Injection of the sand into the transport system takes place according to the type of study being carried out. It is possible to adopt three different techniques to analyze tracer advection. The Time Integrated Method (TIM) has an Eulerian nature, whereby a known quantity of tracer is released at a point and variations in tracer concentration are monitored throughout time at a location down-drift. The Continuous Injection

Method (CIM) is similar to the previous one, with the exception that injection of the tracer is continuous at a known rate. The Spatial Integration Method (SIM) involves a Lagrangian approach, since it allows monitoring of tracer movement both in space and in time. Other advantages of the SIM over the previous methods are that the velocity of transport is computed referring it to the centroid (or center of mass) of the tracer cloud and that it is possible to calculate with confidence a recovery rate. In the case of continuous injection on the submerged beach outside the breakers, dumping from a vessel is a possible method, while in the case of working in the breakers this may be unfeasible in terms of safety of the boat. For this reason, researchers using the TIM and CIM methods preferred injecting the tracer from several points along a transect orthogonal to the beach (Yasso 1965; Boon 1970). Other experimentalists, mainly using the SIM method, placed the tracer at low tide into a shallow trench dug on the beach face, after washing it with liquid soap to avoid grain floating. There is no standard for the size and depth of the trench: the most important requirement is that all the tracer grains are removed almost instantaneously as soon as the site is covered by water. A delay in removal or burial of the tracer at the trench site could invalidate the basic assumptions of the method. RFID tracers can be basically injected in the system following two methods: in clusters on specific locations of the beach or individually on geomorphic features (i.e., storm berms, fair-weather berm, step) along cross-shore profiles. Given their larger size and weight, they need to be injected with a small hand pressure in a stable position among the surrounding sediments in order to avoid any overrated influence of gravity.

Tracer Sampling and Detection

Despite the fact that many authors only worked at surface level, it is preferable to collect shallow cores to assess the three-dimensional advection of the fluorescent tracer. PVC cores or other simple coring techniques are suitable for this purpose. Inman et al. (1980) collected their samples according to the type of analysis to be performed: the first technique was employing a grab sampler, especially designed to collect only the first two centimeters of the surface layer and was used to carry out a spatially distributed sampling to apply the SIM. The second method was using corers that were collecting samples along a cross-shore profile at constant time intervals, for the TIM method. Duane and James (1980) adopted a similar approach for the CIM technique. Samples were collected using a hand-held sloop, built on broad runners, which was penetrating the sand column no more than 1 cm, to avoid sampling at depths where the sand concentration was already in equilibrium. In case it is decided to assess surface distribution, to increase the significance of the results from the cores, it is possible to use automatic detection techniques like the

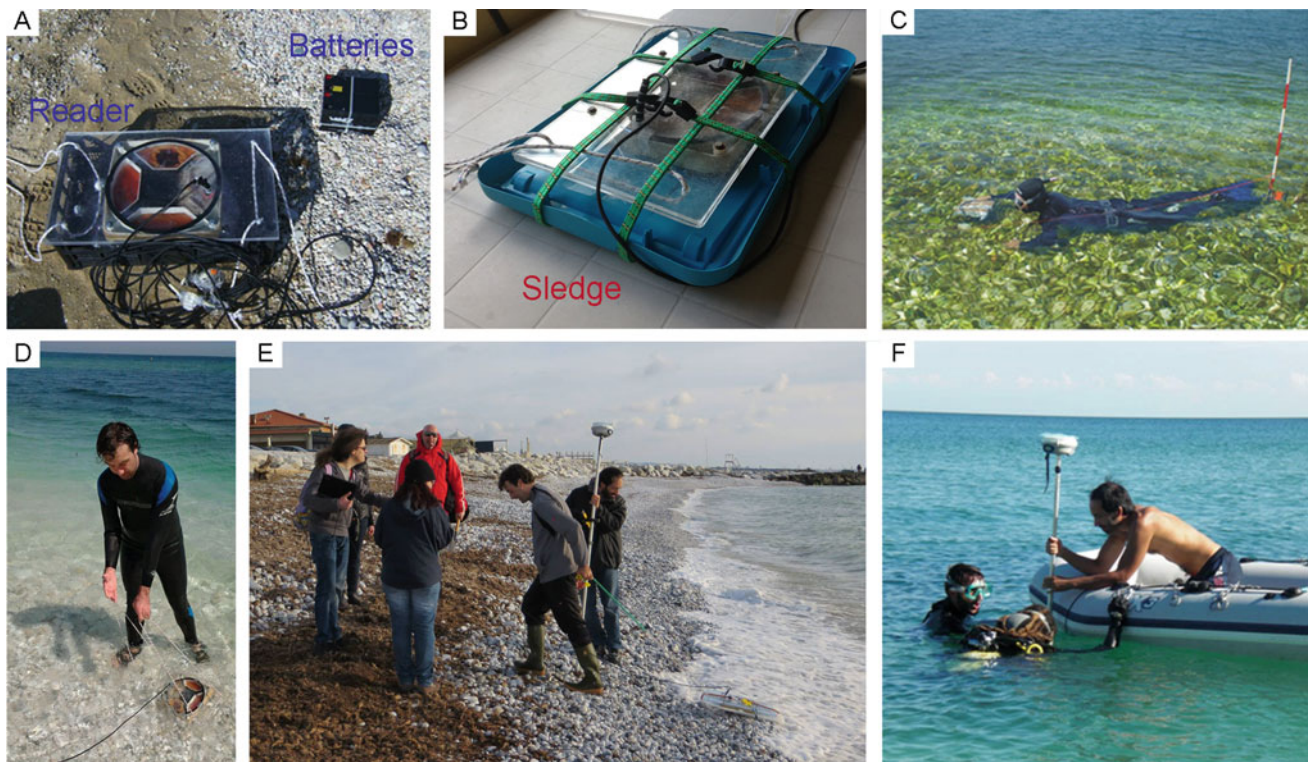
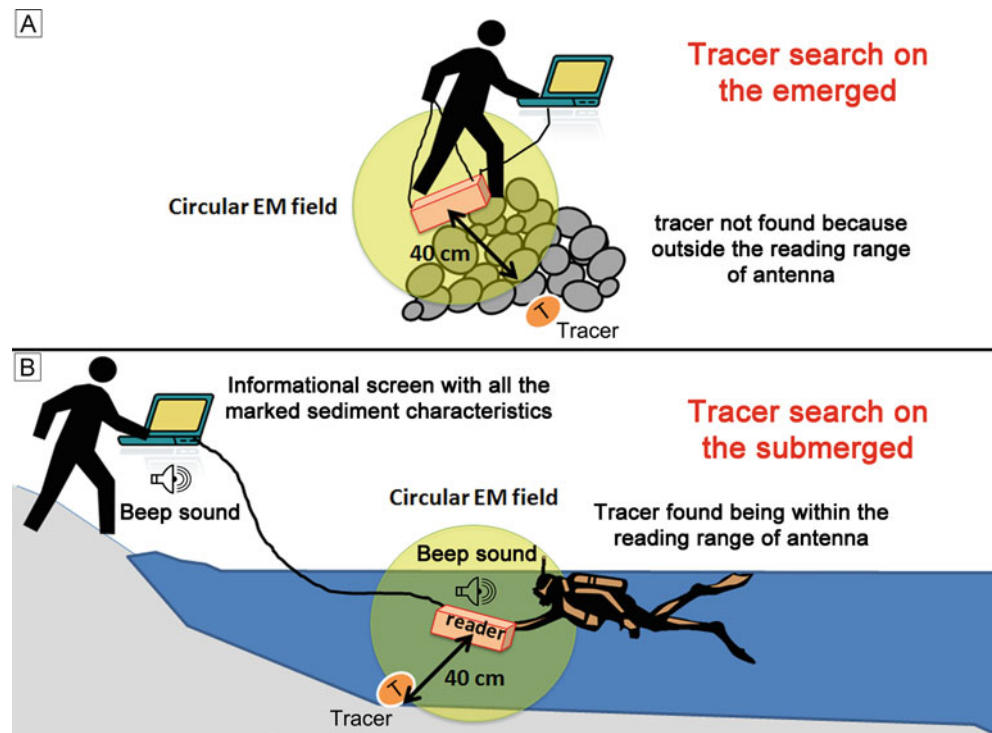
system of Pinto et al. (1994), where a computer system is installed on an ATV (All Terrain Vehicle), to perform rapid assessments using image analysis techniques (Fig. 3b).

Counting of the fluorescent grains in the samples under a UV lamp (Fig. 3c) can be tedious and involve several days of work. Whenever the number of tracer grains in a sample is high, automatic detection or indirect detection methods can be helpful, once the considerable calibration difficulties are overcome. In the study of Chapman and Smith (1977), where the dye was not fluorescent, counting took place following two different methodologies: photography and subsequent enlargement, affixing the samples to long strips of pressure-sensitive adhesive tape and passing the strips under a binocular microscope at an enlargement of 30–40 times to detect the grains. In both cases, the counting was undertaken using an automatic image analyzer.

The spectrofluorimetric method of Farinato and Kraus (1981) was instead based on a more original approach: the technique consisted in washing the sand samples with a solvent able to dissolve the resin and the paint covering the grains. The fluorescence of the effluent can then be measured using a spectrophotometer: the measured intensity is proportional to the number of tagged grains, the solution transmission, and the uniformity of the paint cover. The calibration of the intensity coefficient versus the number of marked grains per gram has to be undertaken according to the degree of marking, the grain size distribution, and the chemical properties of the solvent and of the resin. For the method to work, all sand grains should be almost equi-dimensional: certainly in the case of a uni-modal, well-sorted sand that can be true, but as sorting becomes worse, the method loses significance. Besides, the technique assumes that the paint cover of the grains is uniform. With the recent advances in digital photography and image analysis algorithms, the detection and counting of the grains can be matched with particle size analyses.

RFID tracers do not need compulsory sampling as the detection is done remotely. The searching operations can start few hours after the injection (e.g., 6, 24, or 48 h) in order to study the short-term displacement. Given the long operational life (up to 50 years) of passive RFID transponders, mid- and long-term monitoring periods are feasible. The tracer search on the beach it is a time-consuming operation because every square meter of the beach surface has to be covered by the reader-antenna. The instrument, given its maximum detection range of 40–50 cm, needs to stay as close as possible to the beach surface (Fig. 4a). Several operative techniques had been conceived by scientists to make the tracer research quicker (Fig. 5). When a tracer is found, an informational screenshot on the connected device (laptop, smartphone, or tablet) along with an acoustical beep makes the researcher aware of the tracer detection (Fig. 4b).

Tracers and Coarse Sediment, Fig. 4 Research methodology of RFID tracers on (a) the emerged beach and (b) the submerged beach



Tracers and Coarse Sediment, Fig. 5 (a) The reader-antenna and its batteries: two PVC sheets make the reader waterproof; (b) Reader set on a plastic sledge to easily drag it on the emerged beach; (c) searching

operation made by a scuba diver; (d) searching operation on a steep beachface; (e) trailing of the sledge-equipped reader on the beach surface; (f) measuring location of a detected tracer in the submerged beach

Conclusions

Recently, many investigators throughout the world have been revisiting the tracer technique, especially for field measurements of longshore transport. Tracer advection can be measured on a 3-D basis; it is considered essential that investigators adopt such an approach if they want to obtain meaningful observations. Most field experiments have been confined to small- to medium-scale beach studies, but tracers can be employed successfully to study transport along sandbanks, continental shelves, dune fields, mixed beaches, or gravel barrier (see entry on Gravel Beaches). In a similar fashion, the timescale of these experiments tends to be short (hours to days), because of the small quantities of tracer injected. In the case of fluorescent tracers, only the usage of thousands of kilos of sand will ensure a significant recovery rate at the end of a long-term experiment. Large quantities mixed with sand during a beach recharge scheme could, for example, provide coastal managers (see entry on Beach Management) with an independent method of assessing where the sand is going during monitoring activities, in addition to data for the calibration of longshore drift predictors, to decide whether further intervention is needed. The main limitations of the method are the time and the effort required for data collection and analysis. Recent applications of automatic analyzers, particularly using image analysis methods, suffered from the cost and availability of high-resolution systems. With the recent boom of digital cameras, the power of image analysis has increased in parallel with a decrease in the costs involved.

RFID tracers, like the fluorescent ones, need to be produced in large quantities for significant recovery rates, but given the larger sediment size (usually pebbles or cobbles) hundreds of tracers are normally enough. The greatest improvement brought by the RFID technique is the identification of the tag inside the sediment particle: in this way, apart from a quicker tracing of the sediment in the field, painting and labeling are no longer necessary for the particle identification. Other advantages are the low costs of the transponder and the long operational life (potentially up to 50 years). The main limitation of RFID technique is that it can be applied only to coarse sediments (i.e., gravel field from pebbles to boulders); therefore, other tracing techniques, like fluorescent tracers, are still required on study sites with a wide range of sediment sizes (i.e., mixed beaches). Many studies in the recent years have focused on the displacement of RFID tracers and thanks to the univocal identification of the tag, monitored the physical variations of the particle (e.g., abrasion rate, Bertoni et al. 2016). These are crucial topics for coastal managers (see entry on Beach Management) when they have to quantify the life time of gravel nourishment or have to predict the physical behavior of fill material through the time.

Cross-References

- [Beach and Nearshore Instrumentation](#)
- [Beach Management Tools](#)
- [Boulder Beaches](#)
- [Cross-Shore Variation of Grain Size on Beaches](#)
- [Depth of Disturbance](#)
- [Gravel Beaches](#)
- [Reflective Beaches](#)
- [Shoreline Response to Littoral Drift Barriers](#)
- [Swash Zone Dynamics](#)

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Trottoirs

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Definition

“Trottoir” is French for “pavement” or “sidewalk.” The term has been imprecisely defined and applied in coastal geomorphology, where it has been used to describe a range of intertidal features formed through a variety of processes. Most commonly the term trottoir is used to describe narrow (one to several meters wide), sub-horizontal ledges elevated close to the mean tide level on relatively sheltered rocky coasts, although they may be much broader in more exposed settings. Characteristically, they consist of a biogenic crust several centimeters thick deposited over a shallow platform eroded in an underlying rocky substrate (Trenhaile 1987). Vermetid gastropods and various calcareous algae are major components of many trottoir, although a range of other calcifying organisms contribute. Much broader and extensive landforms composed of a thicker coralline algal veneer over an eroded rock platform have also been referred to as trottoir; for example, a *Lithophyllum* crust 0.5–1.0 m thick developed over eroded horizontally bedded Miocene sandstones covers more than 50,000 m² in the southern Arabian Gulf and has been described as a “coralline algal reef flat trottoir” (Peebles 1993). Intertidal bioconstructed overhangs up to a meter in width have also been described as trottoir, but such features are more correctly called cornices (syn: corniches). These

projections are best developed on vertical or steeply sloping intertidal rocks on sections of coast protected from strong wave action (Bird 2008). The terms “surf bench” or “surf ledge” have been used synonymously with trottoir but are best applied to describe a particular type where encrustations form on the floor of tidal notches, usually on limestone coasts exposed to persistent wave action (Pirazzoli 2007). In these locations, biogenic veneers protect the underlying substrate and inhibit erosion, but erosion may continue above to develop an intertidal bench that can be tens of meters wide. Particularly at sites exposed to vigorous wave action, these features may develop a distinctive rim at the outer edge where wave action encourages more rapid accretion to higher elevations. Surf benches elevated two meters above the high tide level have been reported in the literature (Focke 1978).

Processes

A variety of different processes are likely involved in the formation of the range of different features described as trottoir. Those trottoir more appropriately referred to as cornices may be entirely biogenic, although encrustation at the anchor point may protect the underlying rock substrate from physical, biological, and chemical erosion active on the zones below and above, amplifying their protrusion and prominence. The more characteristic bench-like trottoir form develops when a biogenic veneer is constructed over an erosional platform, the involvement of both constructional and erosional processes leading to contradictory hypotheses to explain their origin (Emery 1962). Surf bench formation, for example, has been attributed to the combined influence of corrosion within the spray zone and bioarmouring by encrustations at the trottoir’s seaward edge which is battered by waves. The relative timing, intensities, and positions of constructional and erosional processes across a trottoir vary depending on setting; factors such as antecedent substrate type and morphology, wave climate, tidal range, relative sea-level history, and ecological setting all likely to influence the rates of both erosion and bioconstruction. For example, notch and/or cliff retreat exposes new substrate for encrustation in high-energy settings but may be replaced by shallower notches with less accommodation space for encrustations to grow on at sites exposed to lower waves. It has been hypothesized that trottoir may reach a dynamic equilibrium with the processes responsible for their formation as erosion rates diminish as trottoirs grow wider. In many trottoir it is probable that both constructional and erosional processes proceed simultaneously; many bioeroding organisms such as blue-green algae and sipunculid worms routinely coexist with those depositing biogenic crust.

Organisms Involved

Biogenic calcium carbonates produced by many organisms are involved in trottoir formation, with the species at a site dependent on climate, exposure, hydrodynamic setting, and a range of other environmental factors. Vermetid worms (e.g., *Dendropoma petraeum*, *Vermetus triquetrus*, *Vermetus nigricans*, *Pomatocerus caeruleus*) and calcareous algae (e.g., *Lithophyllum tortuosum*, *Lithothamnium* sp., *Neogoniolithon notarisi*) are the most common contributors to the biogenic crusts described in the literature, although other calcifying organisms including oysters (e.g., *Crassostrea amasa*, *Crassostrea glomerata*, *Saxostrea* sp.) and sabellariid worms (e.g., *Sabellaria kaiparaensis*, *Sabellaria vulgaris*, *Galeorlaria caespitosa*) are also well represented (Kelletat 1995).

Location

Trottoir are found from northern Scotland to the tropics (Kelletat 1989). A wide range of forms are described from the Mediterranean, especially the warmer southern and eastern areas. Surf ledge or bench type trottoir are more common in tropical waters where they may cover platforms cut into raised reef limestones, and possibly merge with algal pavements constructed seaward of the erosion platform on high-energy reef coasts (Emery 1962). These forms have been described from the Caribbean (e.g., Curacao), Micronesia (e.g., Mariana Islands), Aldabra, Mexico (e.g., La Paz), and Australia (Kelletat 1995; Pirazzoli 2007; Emery 1962; Trenhaile et al. 2015). As noted above, most trottoir form near mid-tide level, although at some locations they occur slightly below this level (Trenhaile et al. 2015) and at others, especially where the surf ledge form develops, they may extend much higher in the tidal range, and even rise into the supratidal (Pirazzoli 2007).

Rate of Growth

Biogenic encrustation and trottoir formation can be rapid in the tropics, with the accretion rates of many biogenic encrusters being comparable with the 1–5 mm/year achieved by encrusting corals (Kelletat 1989). Measurements from algal reefs at St. Croix indicate rates of marginal extension and crustal thickening may exceed 13 mm/year and around 0.8 mm/year, respectively (Adey and Vassar 1975). At Abu Dhabi in the southern Arabian Gulf an algal reef flat trottoir between 0.5 and 1 m thick covers 50,000 m² at close to mean sea level (Peebles 1993). This extensive feature has formed since sea levels stabilized near their current elevation between around 1500 and 1100 years ago (Lokier et al. 2015).

Although most trottoir are of relatively modest size, they can clearly become significant landforms where conditions allow.

Significance

Trottoir usually have a subhorizontal upper surface formed in response to tidal exposure and are composed of resilient calcium carbonate with good preservation material that can be radiocarbon dated. These traits mean that they can be excellent relative sea-level indicators, especially in microtidal settings and where they can be compared with nearby trottoirs of different age and elevation. These attributes and their potential applications to sea-level studies have been recognized. Trottoir have been widely used to reconstruct relative sea-level changes during the Holocene, especially those associated with coastal uplift in seismically active areas (Palyvos et al. 2008; Faivre et al. 2013; Anzidei et al. 2014).

Cross-References

- ▶ Algal Rims
- ▶ Bioconstruction
- ▶ Bioherms and Biostromes
- ▶ Notches
- ▶ Sea-Level Indicators, Biologic
- ▶ Sea-Level Indicators, Geomorphic
- ▶ Sea-Level Indicators: Biological in Depositional Sequences

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Tsunami Deposits

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Synonyms

Tsunamiites; Tsunami beds; Tsunami layers

Definition

Tsunami deposits. The materials that accumulated from a tsunami event constitute a tsunami deposit or tsunami bed or tsunami layer. It may be an accumulation on land, in lakes, along the shore, or in the water offshore.

Tsunamiite. A sedimentary layer formed by a tsunami event. Some tsunamiites may as well be termed *seismites* emphasizing their ultimate origin from earthquakes.

Introduction

The word “tsunami” is a Japanese term (written as *tsu-nami*) which refers to “a big harbor wave.” The term “iminami” is sometimes used to denote “mega-tsunamis.” Monserrate et al. (2006) introduced the word “meteotsunami” to denote tsunami-like coastal sea waves induced by atmospheric processes.

Tsunami waves are characterized by their large radius of rotation (up to a couple of kilometers) and their rapid travel time (in the order of several 100 km/h).

The tsunami wave hits the coasts as a rapidly moving wall of water, thus overflowing the land, which has disastrous effects. The December 26, 2004, Indian Ocean tsunami killed about 230,000 persons; the December 28, 1908, earthquake in the Strait of Messina in Italy killed at least 100,000 persons; and the famous Lisbon earthquake that occurred on November 1, 1755 killed at least 60,000 persons.

Origin of Tsunamis and Tsunami Deposits

Tsunami waves are generated when water masses, in the ocean or in lakes, are suddenly dislocated in the vertical dimension. This can happen through the following five main causes (Fig. 1):

- When the seabed (lake bed) jumps up or down due to an earthquake
- When sediments suddenly slide down the ocean bottom at submarine slides or from the coast into the sea
- When objects are falling down into the water at impacts, calving icebergs, and blocks are falling
- At sudden dislocations of rock masses at submarine volcanic eruptions or volcanic eruptions of islands or coastal areas
- At submarine explosions in association with methane gas venting and nuclear bomb tests (the Bikini Atoll)

The dynamics of the tsunami waves set sediments in motions, which at their later deposition constitute “tsunami deposits.” These deposits may range from big blocks to fine clay particles or organic matter. Usually, those deposits exhibit graded bedding in a fining upward sequence.

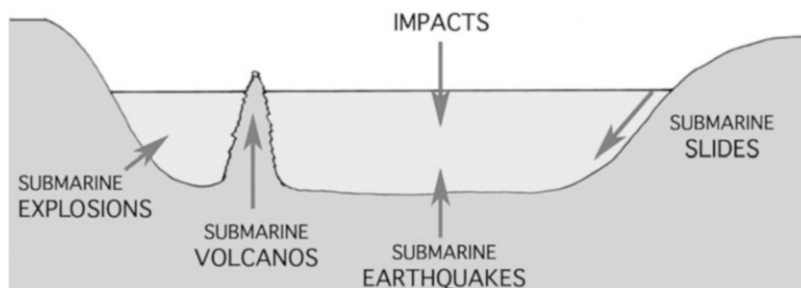
Tsunami deposits (beds, layers, units) are also termed “tsunamiites” (Shiki et al. 2008; Mörrer 2013).

Seismically Induced Tsunamis

The vast majority of big tsunamis are induced by submarine earthquakes. It is the vertical fault dislocation (normal or reversed) that leads to tsunami generation because they affect the vertical water column above. Horizontal fault dislocations (strike-slip faults), on the other hand, do usually not generate tsunamis because the vertical water component is not affected strongly enough. Many famous destructive tsunami events have an earthquake origin:

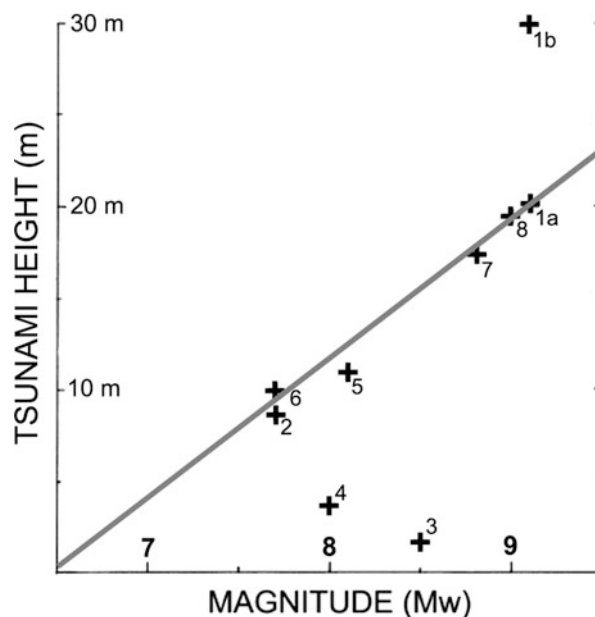
- The March 11, 2011, destructive tsunami in Japan was generated by a 9.0 magnitude earthquake. The wave reached 15–20 m height, at least 18,000 people were killed, and over 452,000 persons had to be relocated. The coastal damage was severe (Tanaka et al. 2012).

Tsunami Deposits, Fig. 1 The five main factors generating tsunami events as further discussed in the text



- The December 26, 2004, Indian Ocean tsunami was generated by a 9.1 magnitude earthquake off Sumatra giving rise to a reverse fault of 1,300 km length and several meters height (Lay et al. 2005). Around 230,000 persons were killed.
- The November 1, 1755, tsunami in Lisbon was generated by an 8.5 magnitude earthquake off Portugal. It is known as “the destruction of Lisbon” (well described by Fonseca 2005). The tsunami was recorded not only in Portugal, Spain, and Morocco but also in the far field in Brazil and the West Indies (Blanc 2011).
- The AD 365 (July 21) tsunami that destroyed the Library of Alexandria was induced by an 8.5 magnitude earthquake near Crete (Stiros 2010; Polonia et al. 2013).
- Other destructive tsunami events that occurred were on September 20, 1498, in Japan which was generated by an 8.3 magnitude earthquake and with 31,000 people killed; on October 28, 1707, in Japan generated by an 8.4 magnitude earthquake and with 30,000 killed; on June 15, 1896, in Japan generated by a 7.6 magnitude earthquake and with 22,000 killed; on August 13, 1868, in Chile generated by an 8.5 magnitude earthquake and with 25,000 killed; on April 24, 1771, in Japan generated by a 7.4 magnitude earthquake and with 12,000 killed; and on January 18, 1586, in Japan generated by an 8.2 magnitude earthquake and with 8,000 killed.
- In Sweden (today, a seismically rather calm area), 17 tsunami events have been recorded in association with paleoseismic events (Mörner 2016a), during the phase of intensive glacial isostatic uplift and intense seismic activity (Mörner and Dawson 2010; Mörner 2011).

There is a relation between maximum tsunami height and seismic magnitude (Lario et al. 2016). There are eight major tsunami events in the last 12 years that generated high-amplitude tsunamis. They are (1) Indian Ocean, 2004, of M 9.1; (2) Java, 2006, of M 7.7; (3) Benkula, 2007, of M 8.5; (4) Peru, 2007, of M 8.0; (5) Samoa, 2009, of M 8.1; (6) Mentawai, 2010, of M 7.7; (7) Chile, 2010, of M 8.8; and (8) Tohoku-oki, 2011, of M 9.0. The relation between tsunami height and seismic magnitude gives a straight line (Fig. 2). Having established this relation, one can introduce an observed



Tsunami Deposits, Fig. 2 Relation between observed maximum tsunami height and magnitude of causal earthquakes (1–8). Having established this relation, observed tsunami heights of paleoseismic events can be converted to corresponding earthquake magnitudes (Modified from Mörner 2016b)

tsunami height value and estimate the corresponding magnitude (Mörner 2016b).

Earth Slide-Induced Tsunamis

The Storegga submarine slide that occurred about 8,000 cal. years BP set up a major tsunami recorded along the Norwegian coast and in Scotland (Dawson et al. 1988; Bondevik et al. 1997).

Submarine slides (or coastal earth slides into the sea or a lake) may be induced by earthquakes or gravity-driven slope stability failure (avalanches).

Impact-Induced Tsunamis

In theory, any celestial body falling into the sea would set up a tsunami. The same is true for icebergs breaking off an ice front and falling into the sea (MacAyeal et al. 2011; Ferreira 2011). Darwin had a breathtaking experience of this in Tierra

del Fuego in 1838 (Darwin 1839). Even stones and block falling down a cliff into water set up local tsunami waves.

The proposed impact at the Cretaceous/tertiary boundary has led to much speculation on the occurrence of megatsunamis and tsunami-derived effects and deposits (e.g., Bourgeois et al. 1988). Often the diameter of a tsunami wave, which might have been in the order of kilometers, was confused with the wave height, which is in the order of meters to a few tens of meters.

Volcanic Eruption-Induced Tsunamis

The Krakatau eruption in Indonesia on August 27, 1883, set up a mega-tsunami of multiple waves and a maximum height of 37 m. Around 40,000 persons were killed. The 1883 Krakatau tsunami is recorded at 35 tide gauge stations scattered all over the Indian, Pacific, and Atlantic oceans. The far-field records may, however, primarily be a function of related atmospheric waves, i.e., meteotsunamis (as proposed by Monserrate et al. 2006).

The Island of Thera (Santorini) in the Aegean Sea exploded at a violent volcanic eruption about 1650 B.C. This event set up an extensive tsunami wave (Bruins et al. 2009), the effects of which have been very much debated and sometimes associated with the saga of Atlantis and its sudden disappearance.

Explosion-Induced Tsunamis

The nuclear bomb test explosion in the Bikini Atoll (23 in the period 1946–1958) set up tsunami waves in the Pacific.

Similarly, it has recently been shown that explosive venting of methane gas as a function of a sudden phase transition in the seabed (or lake bed) from methane ice (clathrate) to methane gas can set up tsunami waves of large height and extensive spatial impact (Mörner 2011, 2016b; Mörner and Dawson 2010).

Tsunami Deposits

Tsunami waves imply severe erosion and damage of cultural and natural objects in coastal areas. The redeposition of sediments and objects set in motion by the tsunami is termed tsunami deposits or tsunami beds or layers.

A tsunami wave hitting a coast usually consists of a sequence of separate waves, termed “a tsunami wave train.” This is sometimes recorded in the tsunami deposits (e.g., Wagner and Srisutam 2011; Mörner 2013).

Offshore Deposits

The tsunami wave has a large diameter of rotation. Therefore, it may generate erosion of the seafloor surface (as illustrated in Fig. 6 of Mörner and Dawson 2010). This erosion leads to the deposition of extensive turbidites, usually in strict graded

bedding. In Sweden, where the former seabed is uplifted and today constitutes dry land, these deposits can be studied in full details (Mörner 2011, 2013).

Submarine liquefaction events can occur together with tsunami events and generate the redeposition of sandy-silty material in widespread layers of turbidite character. They are likely to be a combined effect of liquefaction and tsunami.

A tsunami wave may drag down coastal deposits to significant depths further off the coast. So, for example, “submarine sandstorms” from five tsunami events in historical time in Maldives have generated the redeposition of sand with shallow-water gastropods and corals in caves at depths of 27 and 38 m (Mörner and Dawson 2010).

In nearshore areas, the tsunami may generate turbidites of angular pebbles and grits within clay beds extending far deeper than normal shore processes do (Mörner and Dawson 2010).

Submarine slides may extend as extensive turbidites out over the abyssal plane (e.g., Bouma 1962). The AD 365 earthquake and tsunami (Stiros 2010) generated an extensive turbidite in the Mediterranean (Polonia et al. 2013).

Shore Effects

When the tsunami wave, or rather train of waves, hits a coast, it usually causes extensive erosion, destruction, and relocation of objects (e.g., Keating et al. 2011). Often, the shore becomes severely remodeled (well documented at the March 11, 2011, tsunami in Japan by Tanaka et al. 2012).

The material set in motion must, of course, be redeposited either in the form of remodeling of the shore morphology or by carrying it inlands (point 3, below) or seaward (point 1, above).

Onshore Deposits

The tsunami wave may have a height of up to some tens of meters. The breaking in over a coast generates a run-up, which may reach much higher elevations, however (e.g., Abe 1995). The subsequent drawdown back-swash carries much material to depositional rest. This implies that the onshore to nearshore tsunami deposits are the product of two different phases of the coastal dynamics: the on-swash beds from the run-up phase and the back-swash beds from the drawdown phase.

On-Swash Beds from the Run-Up Phase

The tsunami wave hits the coast with tremendous energy, which is capable of severe destruction and relocation even of heavy object like boulders (e.g., Keating et al. 2011).

The run-up water masses become loaded with sediment particles (of all grain sizes, ranging from clay to boulders) and other objects picked up from the land surface swashed over (boats, trees, house debris, animals, humans, etc.).

Deposition usually occurs closer to the limit of the run-up where the energy decreases enough to allow for deposition. The most likely places of depositions are in depressions and lakes. The tsunami beds deposited often exhibit graded bedding.

One effective way of recording the run-up limit of past tsunamis is to core lakes at higher and higher elevation above the sea level (e.g., Möner and Dawson 2010). Tsunami beds recorded within normal lake deposits usually constitute of a sand layer of graded bedding containing deepwater planktonic microfossils (Mörner and Dawson 2010).

Back-Swash Beds from the Drawdown Phase

The back-swash phase may be responsible for the deposition of more extensive tsunami beds. Deposition primarily occurs in depressions and at the sea level. Occasionally, material may be carried offshore.

Sometimes these deposits are full of archaeological remains like in the coastal tsunami deposits of the Santorini event (Bruins et al. 2009).

Because of successively diminishing forcing at settling, even these deposits may be graded in bedding.

Summary

Tsunami deposits (or “tsunamites”) denote the accumulation of sediment and debris set in motion by tsunami events. These deposits represent accumulation in the onshore environment due to the run-up phase as well as the drawdown phase, in the shore environment due to the remodeling of the coastal morphology and the drawdown of sediments, and in the offshore environment due to the down-wash of littoral sediments to great depths and the erosion of the surface of the seabed by the tsunami wave.

Cross-References

► Seismic Displacement

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Uplift Coasts

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From the late 1800s the early geologists and geomorphologists such as W.M. Davis identified coasts of “emergence” (uplift coasts) and “submergence” (drowned coasts) relative to modern sea level. The idea was promulgated further by Johnson (1919) and Cotton (1974). In a more modern sense we now recognize continental margins as “passive” or “active” (subject to ongoing tectonic processes). Uplift coasts are typically associated with the latter.

Three primary mechanisms of coastal uplift may be identified:

1. Contemporaneous subduction processes at active plate margins are particularly evident around the circumference of the Pacific Ocean. Above the subduction zone the coastline is subjected to numerous tectonic and seismic processes, but particularly uplift and lateral (wrench) faulting, as for example along the northern California coast. Active plate marginal processes creating uplift coasts also occur at a medium regional scale along the Pacific South American coast, New Zealand, New Guinea, Indonesia, and Japan. A type site for an active subduction coast undergoing uplift is the East Coast province of New Zealand.
2. Isostatic rebound from the melting of continental ice sheets, whose load depressed the earth’s crust, provides a second mechanism for uplift of the coast. Resulting uplift coasts are particularly evident over large coastal regions in the northern hemisphere, in particular Alaska, Canada, and northeastern USA; Russia, Scandinavia, and northern

Europe. The latter include uplift coasts of the ancient rock terrains of Scandinavia and the British Isles, as well as the modern Pleistocene glacial tills of the north German plain.

3. Diastrophism – tectonic movements of the earth’s crust, including both epeirogenic (regional uplift without significant deformation) and orogenic (mountain building with deformation) – can produce sectors of uplift coasts. For example, block faulting producing horst and graben structures occur along the South Australian coast (Twidale, 1968), while block faulting influences much of the central coast of China, as well as the coast of Chile and Japan.

Seibold and Berger (1993) illustrate how throughout geological time, a major sedimentary wedge formed around the perimeter of the continents. Upon uplift, it is these sedimentary wedge deposits which undergo erosion from marine action at the coast. Subsequently when they are uplifted beyond the marine influences, subaerial denudation over time erodes out and masks the original specific coastal geomorphic features. However, epeirogenic uplift may also occur on continental margins where there is no sedimentary wedge, so that granitic rocks of continental origin characterize the coastal geomorphology subject to the uplift. On a larger spatial scale, some uplift coasts also present emerged “continental shelves” and continental slopes, and possibly even the “continental rise.”

When considering uplift coasts, one needs also to bear in mind the timescale of its formation: (1) Recent emergence of the coastline leaves a distinctive imprint on the terrain from the previously active marine processes. Landforms that result include raised terraces, often in flights as in New Guinea, abandoned strand lines as in Scandinavia, coastal plains and abandoned cliffs, shore platforms, and stacks. Examples of the latter include the uplifted shore platforms from historical earthquakes such as the 1855 Wellington event which raised the shoreline by ~2 m (Fig. 1), and the coastal plain at Napier which, prior to the 1931 earthquake which destroyed the City of Napier, was a large shallow lagoon, but is now a coastal

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Uplift Coasts, Fig. 1 The uplift shoreline at Wellington, New Zealand, which emerged ~ 2 m after the 1855 earthquake. Note the degraded cliff and uplifted shore platform, now grassed (Photo: T. Healy)



Uplift Coasts, Fig. 2 The tectonically upthrust emerged coastal plain, which was previously the Ahururi lagoon (remnant in the foreground), as viewed in 1981, some 50 years after the uplift of ~ 2 m associated with the 1931 earthquake at Napier, New Zealand. Note the abandoned stacks, and degraded cliffs. This is in an active subduction zone, with the hills in the background subject to continuing tectonic uplift (Photo: T. Healy)



plain (Fig. 2). (2) As time of emergence progresses, normal subaerial weathering and denudation processes take over, acting on the emerged features. Thus the cliffs become degraded, and stream incision occurs on the raised terraces. Eventually all that is identifiable in the landscape is the break of slope indicating location of a previous shoreline.

Cross-References

- [Coastal Changes, Gradual](#)
- [Coastal Changes, Rapid](#)
- [Coral Reefs, Emerged](#)
- [Faulted Coasts](#)

- ▶ [Isostasy](#)
- ▶ [Marine Terraces](#)
- ▶ [Shore Platforms](#)
- ▶ [Submerging Coasts](#)
- ▶ [Tectonics and Neotectonics](#)

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Vegetated Coasts

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Vegetated coasts are those where rooted vascular plants are a persistent feature of the coastal landscape. This can include dunes, salt marshes, sea-grass beds, gravel barriers, and rocky coastlines. The type of vegetation, as well as its importance in shaping the coastal environment, varies according to climate, sedimentary deposits, and wave and tidal energy regimes. These variables control both the type of vegetation that can survive, and the physical context within which the plants interact with sedimentary and geomorphic processes to shape coastal landscapes. Vegetation not only exists in many coastal settings, it is an important biogeomorphic agent and as such is an integral link between the ecological and landform dynamics of coastal systems. While the presence of vegetation in dunes or on salt marshes has long been recognized as an essential characteristic of those environments, the importance of plant growth forms and their adaptations to stressful physicochemical conditions in shaping coastal landforms has only more recently received attention. For example, Bauer and Sherman (1999, p. 73) describe foredunes as being “geomorphologically conditioned by the germination, colonization, and succession of vegetation assemblages characteristic of coastal environments.” Clearly, understanding the dynamics of vegetated coasts requires not only knowledge of the prevailing physical and sedimento-logical conditions but insights into vegetative form and function.

Vegetation at the Coast

Ecological Role

Coastal ecosystems are widely recognized as some of the most productive on earth. Carter (1988) summarizes the net

productivity of coastal ecosystems and notes that, with the exception of coral reefs, highest levels are associated with vegetated coastal systems including subtidal seaweeds and sea grasses, and intertidal marshes and mangroves. The ecological importance of vegetation to coastal systems was recognized by Teal (1962) who suggested that plant detritus, after undergoing microbial decomposition and enrichment, was the basis of the salt marsh food chain. While more recent work has shown that much of the detritus in mangroves and salt marshes is directly consumed by benthic scavengers, and that benthic microalgae and phytoplankton provide important high-quality food sources for higher trophic levels, the importance of vegetation as structure and refuge for secondary producers and consumers, especially nekton, is widely accepted (see Weinstein and Kreeger (2000) for review of current understanding of tidal marsh ecology).

The high productivity of some vegetated coastal systems can result in a large amount of plant debris in the coastal zone. Where high tides and wave action raft this debris onto beaches, the ecological processes can be similar to those described above for marshes. Alongi (1998) described how detached algae and sea grasses can form stacks up to 2 m high on beaches in western Australia, and that the main utilization pathway for the detritus is via colonizing microbes and surf-zone amphipods.

Adapting to Stress

Rapid and regular fluctuations in water level, along with salinities varying from fresh to hypersaline, are among a number of stresses facing coastal vegetation. In some environments these combine with direct attack from waves, human disturbances, and grazing pressure to produce conditions where only specialized plant growth forms can survive. Table 1 illustrates the variety and combination of stresses facing coastal plants. The presence of vegetation in these various coastal environments is testament to their ability to adapt and cope with such harsh conditions. Duke (1992, p. 65) notes of mangroves that the “combination of morphological and physiological adaptations seen in this diverse and

Vegetated Coasts, Table 1 Examples of the physical stresses on vegetation in various coastal environments

Stress/environment	Rocky shore	Gravel barriers	Sand dunes	Sea grasses	Salt marshes	Mangroves
Salinity	■	■	■	■	■	■
Waterlogging			■		■	■
Water deficits	■	■	■		■	■
Lack of anchorage		■	■			
Lack of nutrients		■	■			
High wave activity	■	■				
High temperature fluctuations	■	■	■			■
Grazing	■		■	■	■	
Human disturbances		■	■	■	■	■

unique group of plants have no equal.” The prop roots of *Rhizophora* spp. and the pneumatophores of *Avicennia* spp. are two of a number of ways in which different mangrove species deal with one of these stresses – waterlogging and soil anoxia. Another of the more commonly known adaptations of coastal plants is that way in which the common dune grass *Ammophila arenaria* rolls its leaves under dry conditions to limit the leaf area exposed and reduce water losses. Even lower plants, such as seaweeds, show adaptations such as the way they cope with extreme water losses during low tide and can rapidly rehydrate and resume photosynthesis soon after resubmergence.

The adaptations made by coastal plants to survive in these conditions are frequently structural. Plants that are exposed to excessive flooding and anoxic soils conditions respond by developing air spaces, or aerenchyma, in their root and stem tissue allowing oxygen to diffuse from the aerial parts of the plant to the roots. Common responses to high salt concentrations include barriers to salt uptake within the root system, such as in the mangrove *Avicennia*, and the excretion of unwanted salts through specialized salt glands in the leaves.

Some plants which can tolerate stressful conditions are so specialized that they are readily outcompeted as conditions ameliorate. For example, in laboratory experimental studies, Adams (1963) found that *Spartina alterniflora* tolerates a wide range of salinities. However, in the natural environment it is restricted to the lowest parts of salt marshes where salinity and waterlogging stress are highest. Bertness (1991) showed that at higher elevations in New England salt marshes, *S. alterniflora* is outcompeted by other species such as *Spartina patens*. Importantly, Bertness (1991) concludes that while the lower intertidal limit of growth for individual marsh species is set by their tolerance for physical stresses, the upper intertidal limits are set by plant competition.

A similar type of situation seems to occur with the beachgrass *Ammophila*. According to Packham and Willis (1997) it is capable of withstanding sand accretion rates of approximately 1 m/year, but measures of vegetative vigor usually decline with lower burial rates and in more stable parts of coastal dune systems it is commonly replaced by

other species. This has been attributed to competitive interactions among plants, but Disraeli (1984) showed that the growth of *Ammophila* was lower in areas of shallower burial in the absence of competitors. This phenomena has variously been attributed to the lesser ability of older roots to uptake nutrients and the effects of soil-borne pathogens and parasites which are ameliorated by the delivery of new sand.

The net result of these adaptation and interactions is that many vegetated coasts exhibit clear zonation of plant species in response to stress gradients (e.g., in tidal inundation, salinity, or wave energy).

Sensitivity to Disturbances

Most coastal environments are subjected to periodic physical disturbance by storm waves, storm surges, or other events that alter, at least in the short term, the stability of the substrate. In addition to such episodic events, vegetated coasts can be sensitive to more regular erosional or depositional events that result in progressive environmental change. For gravel barrier systems, Scott (1963) identified five stability classes each characterized by different plants. Very unstable areas were devoid of vegetation, while gravel beaches stable between spring and fall were characterized by annuals. Increasing stability led to the presence of short- and then long-lived perennial plants and areas which had been stable for longer periods became covered with heath-like vegetation. The response of the vegetative community in this case appears to be to the frequency of physical disturbance. However, the importance of organic matter within gravels both for holding moisture and for preventing seeds from sinking too deeply for establishment and germination (Packham and Willis 1997) implies that this is not simply a response to surficial disturbance of the plants. With increased gravel stability, the accumulation of organic matter can proceed providing an environment more hospitable to a wider array of plants. Many smaller annual plants in gravel systems rely on rain and dew as a source of moisture, while larger perennials can penetrate deeper and although unlikely to be able to access the water table may be able to reach freshwater lenses held in smaller gravel barriers.

While physical disturbances associated with storm events or floods usually impact coastal systems at the scale of kilometers, coastal plants can be sensitive to natural disturbances at smaller spatial scales. The most common of these in coastal salt marshes are ice and wrack. In northern latitudes, seasonal ice cover and ice rafting of vegetation, debris, and substrate, as well as scouring by ice blocks is an almost annual occurrence. Bertness (1999) characterizes salt marshes in these areas as in a continual state of recovery – both vegetatively and morphologically. The deposition of wrack – mats of dead plant material rafted to high marsh areas by high tides or storms – can lead to the burial of underlying marsh vegetation. In north Norfolk marshes, Pethick (1974) attributed the increased density of shallow marsh pans on high marshes to wrack. Once vegetation dies, soil salinity increases in the bare areas on the high, infrequently flooded marsh. This salinity can prevent the recolonization by marsh plants and the pans become permanent features of the marsh surface. Bertness (1999) describes how when such bare areas can be colonized by more salt tolerant species, these “fugitives” can ameliorate soil conditions facilitating the invasion of the site by the dominant plant species. Thus the persistence of the disturbance “feature” in the coastal system depends upon the harshness of the physical conditions and the availability of plants that can tolerate such conditions.

Vegetation as Biogeomorphic Agent

The role of plants in coastal biogeomorphology is largely constructive – acting to enhance deposition or protect surfaces from erosive forces. The vascular plants inhabiting soft sediment substrates on the Atlantic Coast of the United States have been termed “bioengineers” by Bertness (1999) because of their ability to stabilize substrates and enhance sedimentation. The role of salt marsh plants in contributing to the vertical growth of coastal marshes, both directly through the accumulation of organic substrate and indirectly by baffling tidal flows and enhancing sediment deposition has been well documented (Reed 1995). Similar direct and indirect effects of vegetation appear to operate on cobble beaches in New England, where Bertness (1999) describes how, where *S. alterniflora* has colonized areas of the high intertidal, it ameliorates summer heating effects by shading, and stabilizes the cobbles which would otherwise be subject to dislodgement and rolling during winter storms. The *Spartina* also “buffers” the effect of waves and provides a calmer environment where other salt-tolerant plants can survive. The recruitment of these plants is also facilitated by the *Spartina* as it traps waterborne seeds.

Perhaps one of the clearest examples of a vegetated coast where the vegetation plays a crucial role in the formation and development of geomorphology is the beach–foredune

environment. Hesp (1984) asserts that foredunes develop on the seaward most vegetated sand surface behind active beaches and identifies four types of incipient fore-dune formation, each distinguished by their relationship to specific vegetative forms or structures. As Hesp also claims that beach ridges are actually relict foredunes, he implies an overriding role for vegetation, in various forms, for the initiation of coastal dune systems, clearly stating (p. 88) that he “believes that biologic processes and pan-aerodynamic interactions assume foremost importance in influencing initial morphologic variation on foredunes.” Furthermore, Carter and Wilson (1990) attribute the role of vegetation in the initial stages of foredune development to the development of complex three-dimensional fluid flow over dunes at Magilligan Point, northern Ireland. They note that gaps in vegetation during foredune initiation can result in gullies or “low saddles” that cross the foredune. These features were observed to funnel airflow through the dune, sometimes creating aeolian mounds on the leeward side, and occasionally resulting in blowouts.

Intertidal vegetation interacts with waterflows slightly differently than dune grass with airflows in that the vegetation can be either completely or partially submerged by the waterflow depending upon the tide. For instance, sea grasses have been shown to strongly influence flows over the bed by reducing turbulence within the canopy and the development of a “stratification” above the canopy. Living in seawater means sea grasses do not require the structure support as terrestrial vegetation and their growth forms are frequently extremely flexible. This leads to deflection and compression of the plant canopy under strong flows, reducing friction. Differences in sea-grass morphology, from small shallow rooted to larger strap-bladed forms, alter the specific interaction between the plants and the flow but in most cases the presence of the vegetation significantly increases the mean threshold velocity for sediment motion when compared to adjacent bare sands.

The role of vegetative bioengineering in coastal dynamics is sometimes more readily seen when disturbance to the vegetation results in dramatic system change. Bertness (1999) reports a dramatic change from sandy and muddy subtidal substrates to coarse cobble bottoms after a wasting disease destroyed much of the eelgrass beds on the northern Atlantic shores of the United States. In addition to this physical change, the ecological consequences were substantial with a population crash in the bay scallop which relies on eelgrass not only as a food source but to provide structures for young recruits to attach to.

In dune environments, the loss of vegetative cover due to human disturbances such as trampling and all-terrain vehicle (ATV) use are frequently viewed as leading to the degradation of the dune system and prompting a need for management action. However, Thomas (1999) cautions that patchy development of blowouts is a natural feature of most dune systems

and that although disturbance of the vegetation can undoubtedly lead to increased turbulence this should not simply trigger efforts at revegetation. Rather a detailed understanding of the vegetation surface conditions must be used to project the implications of the apparent disturbance and plan any remedial action. This call for a more holistic recognition of vegetated coastal environments as biogeomorphic systems recognizes that management actions must appreciate the natural dynamics of the system as well as its form and surficial characteristics.

Future Issues for Vegetated Coasts

The importance of plant growth and survival to the sustainability of both landforms and ecosystems of vegetated coasts means that environmental changes that alter vegetative structure can have important implications. Two factors that threaten vegetated coasts are the massive increases in coastal populations, and the associated development pressures, and future climate changes.

The coastal population of the United States is currently growing faster than the nation's population as a whole, a trend that is projected to continue. By 2010, 75% of the US population is expected to live within 50 mi. of the coast. Nicholls and Small (2002) identify 23% of the global population as living in coastal areas and note that most of the near-coastal population does not live in large cities but in smaller towns and in rural areas. The growing coastal population, increasing economic development, and expanding coastal communities exert pressure on vegetated coasts through recreational pressures, direct damage due to filling or development, and lowered water tables. Pinder and Witherick (1990) point to the loss of coastal wetlands associated with growth of ports and urbanization in the second half of the twentieth century, and dunes and other intertidal vegetated habitats have been similarly consumed. The societal need for development space at the coast, especially for industries and port facilities that are directly functionally related to the coastal ocean, combined with the diminishing area of land available and the recognition of the ecological, recreational, and esthetic value of vegetated coasts, has led to increasing examination of the potential for creating new land specifically to support development. However, while this approach may relieve development pressure on existing systems, it has consequences for coastal sedimentary and hydrodynamics. Arens et al. (2001) examined the potential environmental impacts of offshore island development in the Netherlands and noted effects associated with construction and to the marine ecosystem based on the footprint of the island and the direct loss of habitat. Coastal dunes and higher plants were largely unaffected and Arens et al. projected only local alterations in the sediment budgets and wave climates that support the adjacent vegetated coasts.

While vegetated coasts are often characterized by plants adapted to living in stressful conditions, changes in climate may increase stresses beyond their tolerance range resulting in shifts to nonvegetated coastal systems and potentially dramatic geomorphic transitions. Scavia et al. (2002) note that projected increases in sea level are unlikely to have near-term catastrophic impacts on coastal wetlands and mangroves, but when combined with other stresses such as alterations to local hydrology or sediment movement patterns associated with development pressures, the long-term consequences may be severe. For vegetated coastal forms to survive they must increase elevation in place (e.g., accrete soil), or migrate inland to keep pace with the rising sea. However, coastal structures and river management alter sediment supplies, and thus limit vertical building, and developed coastlines limit migration. Management approaches for vegetated coasts during climate change must be focused on alleviating the stresses imposed by human alteration of coastal systems, and allowing transitions in form and ecology to cope with the changing sea level, temperature, and precipitation regimes. Vegetated coasts are among our most valued, and highly pressured systems. By their very nature they can survive change and harsh conditions. Our management actions must allow the natural cycles that have sustained them in the past to continue long into the future.

Cross-References

- [Bioengineered Shore Protection](#)
- [Dune Ridges](#)
- [Mangroves, Ecology](#)
- [Salt Marsh](#)
- [Tidal Creeks](#)
- [Tidal Environments](#)
- [Wetlands](#)

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Vibracore

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Vibracoring is a state-of-the-art sediment sampling methodology for retrieving continuous, undisturbed cores. Also referred to as *vibrocoring*, this mechanical drilling technique is used to collect core samples (referred to as either *vibracores* or *vibrocores*) from unconsolidated, loosely compacted, or semi-lithified materials by driving a tube with a vibrating device. Large, heavy-duty vibracores can work in water up to 5000 m deep and can retrieve core samples up to 13 m in length. In coastal shallow-water environments where use of heavy equipment is limited by trafficability and ground

support on land and high-energy conditions alongshore, cores are shorter, usually about 6 m in length.

The coastal/marine environment presents specialized conditions that are not encountered on *terra firma* or in deeper oceanic waters. Because many onshore areas are characterized by coastal wetlands with extensive areas of organic soils in marshy or swampy conditions, tidal sand and muds, or lacustrine and lagoonal facies, beach sands, or chenier-type materials, access for conventional drilling equipment and personnel is often limited, if not by biophysical conditions, then at least economically. Financial considerations are often limiting when costs of sediment retrieval are very high on a per-sample basis. Large multinational corporations involved in petroleum exploration can go almost anywhere to penetrate the most inhospitable environments while sparing no expense to obtain conventional long-drill cores. Most coastal/marine research, however, operates on a reduced cost basis that must be efficient and cost-effective. There are also numerous subaqueous environments that are normally hostile to the positioning of conventional drilling equipment and so alternative sediment-sampling methods are sought. Retrieval of shallow-water sediments using conventional cylindrical coring methods (e.g., those that employ gravity or piston cores) is limited by shallow water depth as well as the nature and habit of the materials requiring undisturbed collection. Resistance of the sediment restricts penetration of gravity cores and core-lengths typically obtained are less than 3–5 m, depending upon the firmness of the sediment.

Sediment cores obtained from a vibracore system are invaluable because they permit direct, detailed examination of composition and layering in sequences of subsurface sediment (Lanesky et al. 1979; Watson and Krupa 1984). Examination of material sequences in vibracores provides information regarding the history of depositional environments (e.g., Brooks et al. 1995) and the physical processes that were operative during sedimentation. Vibracores and subsamples derived from them find almost limitless applications in scientific geological research, engineering and geotechnical pre- and re-design, and environmental investigations. The versatility of the vibracoring technique is illustrated by its utilization in sampling riverine sediments, lacustrine deposits, organic accumulations in marshes and bogs, as well as nearshore shallow water coring. Although examples of applications are legion, of primary interest to most coastal specialists are results that are relevant to the scientific study of coastal zone morphodynamics, evolutionary sequences, and exploratory searches for beach-quality sand that is suitable for beach renourishment.

The utility of vibracores, as seen in these few examples from a huge literature, has been demonstrated in: (1) subsurface exploratory sampling for geotechnical purposes (e.g., Meisburger 1990; Larson et al. 1997), (2) determination of depositional stratigraphy and geomorphological history in

marine, estuarine, fresh water, or wetland environments including marshes, swamps, and peat bogs (e.g., Snowden et al. 1977; Amos 1978; Kraft et al. 1979; Stevenson et al. 1986; Morang et al. 1993; Kirby et al. 1994), (3) glacioeustatic sea-level changes (e.g., Gehrels 1994; Harvey et al. 1999), (4) integrating studies of barrier island evolution, bars, lithofacies, and sedimento-stratigraphy (e.g., Davis and Kuhn 1985; Davis et al. 1993; Brooks et al. 1995), (5) offshore sand searches to locate beach-quality sands for beach restoration (e.g., Finkl et al. 1997; Freedenberg et al. 2000), (6) environmental studies of pollution or contamination by hydrocarbons or heavy metals (e.g., Varekamp 1991), (7) investigations to determine seafloor environments or bottom types (e.g., Barnhardt et al. 1998; Knebel and Poppe 2000), (8) deltaic and sedimentary shelf processes (e.g., Delaune et al. 1983; Brooks et al. 1995; Levitan et al. 2000; Toldo et al. 2000), (8) collection of samples for chemical, biological, and physical analyses (e.g., Kadlec and Robbins 1984; Gehrels 1994), (9) verification of seismic stratigraphy or seismic stratigraphic sequences (e.g., Shipp et al. 1991; Knebel and Poppe 2000), (10) preliminary investigations for the purpose of determining the presence and positions of paleoshorelines (e.g., Gayes and Bokuniewicz 1991; Shipp et al. 1991), and (7) stratigraphic and mineral surveys (e.g., Barusseau et al. 1988; Brooks et al. 1995). Use of vibracores to substantiate seismic interpretations (e.g., sidescan sonar, Uniboom seismic reflection-sub-bottom boomer profiles) is a major application that integrates geological and geophysical methodologies to advantage in many studies. Marine or sub-aerial unconformities, seismic reflectors, stratigraphic units, internal structures of morphodynamic units, sediment textural properties, and even geomorphic features can be identified in vibracores. Although vibracores find use in many diverse applications, the overall setup of coring systems is an important consideration.

Early Attempts to Obtain (Undisturbed) Samples in Coastal/Marine Environments

Retrieval of marine sediments is nearly as old as the science of oceanography itself. Initial interest focused on getting an idea of what kinds of sediments were deposited on the seafloor but, then, as technological advancements assisted the desire for more complete information, new methods of bottom sampling were developed. The purposes of sampling marine sediments varied among professionals as the pioneering biologist, petrographer, sedimentologist, and civil engineer had different demands. Reviews of some historical efforts in bottom sampling are summarized, for example, by Trask (1939), Sanders (1960), Rossfelder and Marshall (1967), and Watson and Krupa (1984) who note the pros and cons of numerous platforms, rigs, and devices such as the Strøm coring tube, Twenhofel coring tube, Varney pile-driver sampler, Trask suction sampler, Renn

sampler, Piggot coring apparatus, Kudinov vibro-piston core sampler, etc. These early devices suffered from a variety of shortcomings that were eventually overcome by improved sampling methods. Modern methods of coring are not without problems, but they have been reduced to the greatest possible extent to satisfy the purpose of a particular analysis.

Terrestrial deposits in coastal marshes and on coastal plains were initially sampled using hand augers, which are still in use today for special applications (see discussion in Soil Survey Division Staff 1993). Screw or worm augers do not provide undisturbed samples, but they can bring up materials from several meters depth by adding extra lengths to the shaft. Barrel augers, core augers, bucket augers, on the other hand, have a cylinder or barrel to hold the soil, which is forced into the barrel by cutting lips at the lower end. The upper end of the cylinder is attached to a length of pipe with a crosspiece for turning by hand at the top. Although both ends of the cylinder are open, the soil generally packs so that it stays in place while the auger is removed from the hole. Barrel augers disturb the soil less than screw augers. Soil structure, porosity, consistence, and color can be better observed. Barrel augers work well in loose or sandy soils and in compact soils. They are not well suited for use in wet or clayey soils, though an open-sided barrel is available that works well. They also work poorly in stony and gravelly soils. Barrel augers bore more slowly than screw augers or probes, but they are easy to pull from the hole.

The Dutch auger is a modified barrel auger having two connected straps with lips. The cylinder is about 5–10 cm in diameter. The Dutch mud auger works well in moist or wet soils of moderately fine or fine texture, but poorly in other moist or wet soils and in all dry soil.

Although soil augers are simple in design and somewhat crude in appearance, considerable skill is required to use them effectively and safely. They must be pulled from the soil by using a technique that puts stress on the leg muscles, rather than the back muscles, to avoid serious back injury. Twisting the auger firmly while pulling takes advantage of the inclined plane of the screw to break the soil loose. A pair of pipe wrenches is needed to add and remove lengths of shafts and bits.

Examinations of deep deposits of peat are made with special tubelike samplers. A peat sampler designed by the Macaulay Institute for Soil Research (Aberdeen, Scotland), for example, takes a relatively undisturbed volume that can be used for measurement of bulk density. The Davis peat sampler, consists of 10 or more sections of steel rods, each 60–120 cm in length, and a cylinder of brass or Duralumin, approximately 35 cm long with an inside diameter of about 1.9 cm. The cylinder has a plunger, cone-shaped, at the lower end and a spring catch near the upper end. The sampler is pressed into the peat until the desired depth for taking the sample is needed. Then the spring catch is released, allowing the plunger to be withdrawn from the cylinder. With the plunger withdrawn, forcing it further downward fills the

cylinder. The cylinder protects the sample from contamination and preserves its structure when the sample is removed. With this instrument, one can avoid the error of thinking that firm bottom has been hit when actually a buried log is encountered.

Probes consist of a small-bore tube that has a tempered sharp cutting edge slightly smaller in bore but larger in outside diameter than the barrel. Approximately one-third of the tube is cut away above the cutting edges so that the soil can be observed and removed. Probes are about 2.5 cm in diameter and about 20–40 cm in length. The tube is attached to a shaft with a “T” handle at the opposite end. Adding or removing sections can vary shaft length. Probes can be used to examine the soil to a depth of 2 m. Rubber or plastic mallets can be used to drive the tube into the soil; a pair of pipe wrenches is needed to add and remove lengths of shaft.

Probes work well in moist, medium textured soils that are free of gravel, stones, and dense layers. Under these conditions, the soil can be examined faster than with an auger. Probes are very difficult to use in dry, dense, or poorly graded soil, and in soil containing gravel or stones. Probes disturb the soil less than augers, but they retrieve less soil for examination. Probes are light and easily carried, and they pull from the hole more easily than screw augers. Use of a soil probe is the fastest method to collect samples of surface layers for analysis. Probes used with power equipment have wide applications in surveys of surficial sediments. Coastal specialists are increasing turning to mechanical sampling devices as the need increases for longer cores of undisturbed samples. Vibracores meet these needs in many different types of environmental situations, but there are still limitations to collection of coastal/marine sediments. The essentials of vibracoring systems, which supplement hand sampling in terrestrial coastal environments and partially replace other types of marine bottom samplers, are briefly summarized in what follows.

Vibratory Coring Systems

These relatively simple devices, referred to as vibracorers, consist of three essential components: frame, core barrel, and vibrator (Hoyt and Demarest III 1981). The frame, which allows the corer to stand free on the seafloor, consists of a quadrapod or tripod arrangement with legs attached to a vertical beam that in turn supports and guides the core barrel and vibrator (Fig. 1). The core barrel and vibrator slide on the beam for coring and retraction. The core barrel assembly consists of 7.5 cm or 10 cm diameter (thin walled) aluminum pipe fitted with a cutter head and core catcher (which holds the sediment inside the barrel when it is withdrawn from the sediments), and a plastic tube inner liner that contains the core material (cf. Fig. 2). There are many different kinds of



Vibracore, Fig. 1 A 10-m vibracore (Alpine rig) being raised from the Gulf of Mexico seafloor offshore from Louisiana. The quadrapod stand stabilizes the vibracore frame, which sits on the seabed and allows the pneumatic vibrahead to slide downwards with core penetration. The entire apparatus is lifted aboard the research vessel using a hydraulic crane that reaches over the stern (photo courtesy of Coastal Planning & Engineering, Boca Raton, Florida)

innovative assemblies that can include, for example, an electronic penetrometer to record time and depth of penetration of the core pipe into the sediments (Smith 1992). Deploying and retrieving the corer usually requires a hydraulic crane (Fig. 3), A-frame, or similar winch or hoisting equipment with a lifting capacity of at least 10–15 tons (Meisburger 1990). Shallow protected waters may allow use of small barges, often of simple construction from polystyrene blocks wrapped in heavy-duty industrial plastic and encased in a wooden frame, mounted with a superstructure from which the vibracoring device can be deployed (e.g., Wright et al. 1999). A marine Kiel Vibracorer can, for example, be adapted for deployment from a well in the barge.

Smaller, lightweight, hand-carried units are used to advantage in coastal environments on marshland or in very shallow water. Fig. 4, for example, shows a small backpack power unit



Vibracore, Fig. 2 Split core liners from vibracore that penetrated sandy bottom sediments in an inter-reefal area offshore from Miami, Florida. The vibracores are split longitudinally, one half is retained for archival purposes and the other is logged and analyzed in the laboratory for grain-size parameters, organic content, coarse fragments (e.g., loose shell hash, pieces of coral, chunks of coquina), mineral composition (e.g., silicates versus carbonates), etc. Note: Core barrels are generally cut into sections that are about 2 m in length. Depending on the length of core obtained, there may be a small section left over that would be difficult to split lengthwise. These small leftover sections also tend to be disturbed because they are at the end of the core. The two PVC ends shown here are used to store “bit samples” because the core-cutting bit does not have a liner. Typically, when a core is being recovered, the material in the bit falls out because it is below the core catcher. Sometimes the sample stays in the bit and it is saved as shown in the photo. The two piled samples sitting on paper in the center of the photograph are plugs of rock, cut from the subsurface, that were lodged in the bit (photo courtesy of Coastal Planning & Engineering, Boca Raton, Florida)

that is deployed in shallow water to obtain marsh sediments. Once driven into the marsh sediments, the core is retrieved by various methods for pullout but in this case a hand-operated come-along is used to extract the core (Fig. 5). After the core is extracted, it can be transported to the laboratory or split in the field as shown here (Fig. 6). Because there was easy access to a grassy work area near the sample site, the core barrel was carried to dry land and split using a skill saw. Splitting in the field provides opportunity for immediate inspection of the core and assessment of environmental conditions, as shown in Fig. 7. Other advantages accrue from splitting a core in the field because it can be immediately determined whether the



Vibracore, Fig. 3 Same vibracore assembly unit shown in Fig. 1 being laid down on the deck. The long core inside the 10-cm-diameter core barrel will provide valuable undisturbed sediment data from the seabed on the continental shelf off Louisiana. Operations such as this one require experienced crew to handle these large vibracoring units (photo courtesy of Coastal Planning & Engineering, Boca Raton, Florida)



Vibracore, Fig. 4 Small, portable vibracore. A backpack power unit operates this pneumatic vibracore; the core barrel is stabilized by hand as the unit penetrates the marsh sediments. The lightweight aluminum A-frame is carried to the site and waits for use in retrieval of the core (photo courtesy of Steve Krupa, South Florida Water Management District, West Palm Beach, Florida)



Vibracore, Fig. 5 Retrieval of vibracore using a portable aluminum A-frame. The lightweight A-frame has pads on the feet so the unit does not sink into the sediment when load is applied. One leg of the A-frame has steps so the driller can extract the core barrel with a hand-operated come along. The A-frame must be placed vertically over the core barrel to ensure easy extraction without flexure or bending of the core. The A-frame collapses into a compact unit for transporting (Photo courtesy of Steve Krupa, South Florida Water Management District, West Palm Beach, Florida)



Vibracore, Fig. 6 Splitting a vibracore. As shown here, the core barrel is laid in a wooden frame and an electric circular saw uses the frame as a guide to cut the flight lengthwise. The upper half is then rotated away from the frame and placed on the ground for observation and sampling (Photo courtesy of Steve Krupa, South Florida Water Management District, West Palm Beach, Florida)

core is short due to loss of sediment or compaction, or whether there are other abnormalities such as coarse materials that plug the core causing gaps in sediment retrieval, etc. The sampling program can be modified on the basis of what is observed in the recovered materials. This flexibility in the field is important because deployment and setup are often significant costs in sample retrieval.

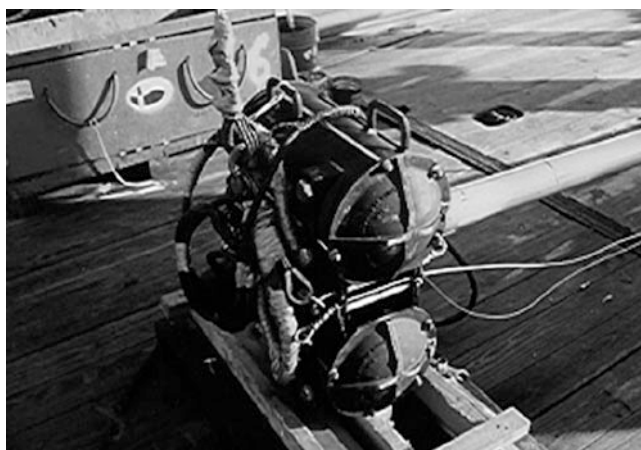


Vibracore, Fig. 7 Split core. After cutting the core barrel lengthwise, the contents are visually inspected for a range of physical parameters such as sediment grain size, color, stratigraphic discontinuities, coarse fragments, etc. Shown here is a fine-grained marsh sediment from the south coast of Long Island, New York (photo courtesy of Steve Krupa, South Florida Water Management District, West Palm Beach, Florida)

Equipment, Design, and Function

The main consideration in developing vibracoring systems, often referred to as the “Rule of Deployment,” that most researchers learn by experience is that “the cost of an operation is related to the size of the vessel which is related to the size of the draw works which is related to whatever hangs at the end of the cable.” These concerns are especially important to boggy onshore sites where access is limited and in shallowwater work where vessel size and equipment load is restricted. Realizing the need to minimize the weight of individual parts while maximizing the overall force/weight ratio, it becomes obvious to select equipment that can be handled with limited manpower from all-terrain vehicles, small vessels, and even inflatable barges (e.g., Hoyt and Demarest III 1981). A variety of vibracore units are commercially available. Some are small, lightweight, and portable, whereas others are large heavy units that can only be deployed from large vessels. There are various types of bottom-standing rigid frames that are ordinarily used for stabilizing and guiding vibracores (cf. Figs. 1 and 3 for heavy-duty marine examples and Figs. 4 and 5 for light-duty terrestrial setups).

The principle behind a vibracore is the development of high-frequency, low-amplitude vibration that is transferred from the vibracore head (vibrahead) down through the attached barrel or core tube (Fig. 8). The vibrator generates sufficient vibrations by means of pneumatic/ hydraulic/electric motors that are sealed in a submersible housing. Vibrations combined with instrument weight drive the core barrel into the sediment/substrate. The vibrational energy delivered



Vibrocoring, Fig. 8 Detail of the parallel mount of electric vibrocoring heads, showing extension of the core barrel below the vibrahead (photo courtesy of Coastal Planning & Engineering, Boca Raton, Florida)

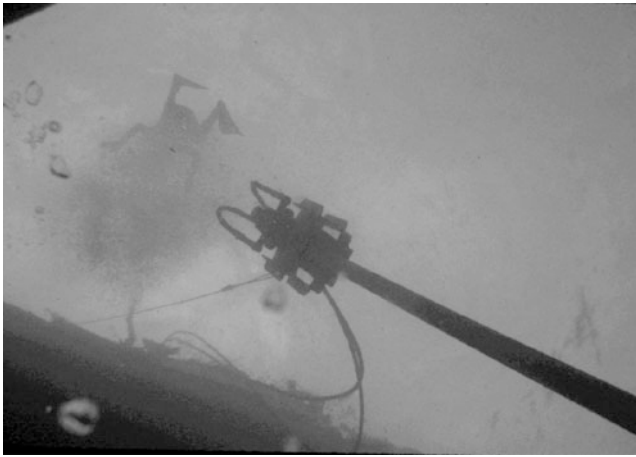
to the core tube (Fig. 9; cf. Figs. 1 and 3) induces vertical penetration by temporarily displacing sediment particles to overcome frontal resistance and wall friction. The technique is thus most efficient in water-saturated sediments because vibration increases the pore pressure along the wall of the core tube by generating a thin layer of liquefaction. As the vibrating tube penetrates the sediment, it displaces the bedded particles on both sides of the wall. This results in the collection of a largely undisturbed core of sediment within the vibrocoring tube. Although useful in water-saturated subaerial environments, such as in marshes, underwater sediments present the optimum medium of application. Starting in the 1950s, vibrocoring became increasingly accepted as an efficient and useful method for collecting underwater core samples. The technique was initially slow to gain acceptance because vibrators could not be easily outfitted for underwater use; the problem was soon overcome and a variety of vibrators are now available. Figure 10 shows a typical medium-sized vibrocoring assembly being lowered over the side of a research vessel.

Among the three basic types of vibrators (i.e., pneumatic, hydraulic, and electric), pneumatic piston vibrators were preferred in early research because they worked underwater and did not involve the undersea use of electrical current. Although still deployed in many surveys, this type of vibrocoring setup requires an air compressor and the hoses that sometimes become an impediment in swift or choppy waters. Hydraulic vibrators use fluid flows in a closed circuit in balance with the ambient environment. A hydraulic power plant and an umbilical hose are required, presenting similar drawbacks to pneumatic vibrators. All things considered, including the force/weight ratio from the power source to the vibrahead, electric vibrators become an attractive choice, particularly when the power source is already part of the



Vibrocoring, Fig. 9 The 7.5-cm-diameter core barrel is attached to the electric vibrahead assembly prior to being lowered overboard into the Atlantic Ocean as part of an offshore sand search near Miami, Florida. Although the unit is of medium size, the weight of the vibrocoring assembly requires crane winching from the deck (photo courtesy of Coastal Planning & Engineering, Boca Raton, Florida)

vessel system. Many researchers conclude that electric vibrators are good choices for work in shallow waters viz. those characterized by surf or spray zones. There are two main types of electric vibrators, electromagnetic vibrators and rotating-eccentric vibrators. The second type is often preferred because the dynamics of the overall spring-like system soil-core-tube-vibrahead, the contra-rotating eccentric vibrators mounted in parallel (cf. Fig. 8), does not require mechanical linkage or special gear for establishing synchronized motion for delivering oscillatory force along a vertical axis. Many combinations of vibrocoring assembly are available for a variety of environmental conditions. Lightweight parallel-mount twin vibrators that, for example, run at 8,000 V/m on standard 110 V AC current, can be operated with a 5 kW camping-type generator. Heavier models often run at 2,800–3,400 V/m on 3-phase, 50 or 60 Hz, 220–440 V power sources. There are also vibrators, operating in the



Vibracore, Fig. 10 Diver's view of vibracore being lowered over the side of the research vessel. The dark area in the upper left of the photo is the bottom of the vessel's hull. Cables to power the vibrahead and a lifting cable are visible to the left of the diver near the water surface (photo courtesy of Coastal Planning & Engineering, Boca Raton, Florida)

medium-frequency range, based on a single vibrator that delivers an oscillatory force to the vertical axis, also often operating on 3-phase 220–440 V. Other variations of basic setups include a “vibrotorsional” operating mode where the two contra-rotating vibrators are mounted along the same axis instead of being in parallel. This coaxial configuration or arrangement synchronizes spontaneously, as in a parallel mount, while adding a horizontal oscillatory torque of small amplitude to the main vertical oscillation. Above the medium range, there is a domain of vibracoring known as “resonant drive” with frequencies on the order of 200–360 Hz (12,000–22,000 V/m) and amplitudes down to a fraction of a millimeter. In this resonant drive mode, the core tube vibrates like a musical string with stationary nodes and antinodes that help to efficiently overcome wall friction, but these units lack force and amplitude for frontal penetration. There is thus a range of vibracore types that can be deployed to specific use.

Advantages and Disadvantages of Vibracoring

Classical problems associated with coring in general include, for example, flow of external sediment into the core barrel due to increased or reduced stress below the core nose, wall friction, sediment deflection below the cutter, shear failures, thinning or compression of softer strata, thixotropic liquefaction, and textural rearrangement (e.g., Rossfelder and Marshall 1967; Smith 1984; Crusius and Anderson 1991; Morton and White 1997). Withdrawal disturbances may take place in response to decrease in hydrostatic pressure below the sample, increase in pressure over the sample, lack

of wall friction, and adhesion between the sample and the core wall. The sampler may inadvertently be tipped over and dragged along the bottom because of improper retrieval. Even though modern sampling methods have been overcome, there are still sampling problems associated with state-of-the-art core retrieval using vibracore samplers. Some of these difficulties are briefly outlined as precautionary observations for a system of bottom sediment sampling that is generally reliable (e.g., Blomqvist 1985).

Vibracores have the advantages of simple construction and easy mobilization, but they are sometimes unwieldy in congested commercial areas such as harbors, and their cost may be beyond the budget of small consultants or universities (Larson et al. 1997). Vibracores are not very effective in either very compact (dense) or very loose sediment; they are not suitable for sediments containing large clasts that can block penetration of the core barrel.

While common vibratory corers are capable of penetrating 6 m or more in unconsolidated sediment, actual performance depends on the nature of the sub-bottom sediment. Under unfavorable circumstances (viz. rough seas, rocky bottom) very little sediment may be recovered. Limited recovery may be primarily due to lack of penetration of the core barrel, blockage of the bit by rock, or loss of sediment during recovery. Core penetration is measured both visually and with an electronic penetrometer. Core recovery is determined by measuring the total length of sediment in the core barrel immediately following retrieval, and the percent recovery calculated as follows (Brooks et al. 1999):

$$\% \text{ Recovery} = (\text{length recovered} / \text{length penetrated}) \times 100$$

In certain instances, penetration refusal is met before full penetration of sediment is achieved (Anders and Hanson 1990; Morton and White 1997). In such cases the vibracore is removed, and the short core is extracted and stored. A new core liner is installed and the vibracore is again deployed. The core barrel is hydraulically jetted down to the depth of the refusal and then the regular vibracoring is resumed to the targeted depth. In other cases, refusal means that the vibracore cannot be driven any further into the sediment and the borehole is terminated. Coarse fragments such as diamictite, scree, lag, or carbonate/coral rubble may, for example, be larger than the core barrel or the deposit may be so densely packed, if containing smaller diameter clasts, that vibracoring into these materials becomes problematic. Core barrels sometimes get stuck in the sediment and removal requires patience, strength, and some ingenuity.

Vibratory corers are capable of penetrating up to 12 m of unconsolidated sediment, but actual performance depends on the nature of the sub-bottom material. Under unfavorable conditions, however, less than 1 m may be recovered; no coring device has been developed that eliminates the potential

for core shortening. There are, however, practical solutions to problems associated with vibracore sample loss or compaction (Smith 1992). Core shortening may result from several factors that include, for example, physical compaction, sediment thinning, or sediment bypassing (Morton and White 1997). Limited recovery occurs in response to lack of penetration where stiff clays, gravel, and hardpacked fine to very fine sand are usually most difficult to penetrate. “Freezing” of material in the core liner, which is an age-old coring problem due to skin friction before full penetration is reached (e.g., Rossfelder and Marshall 1967), stops new material from entering the sampler while the core barrel continues to penetrate; this process may result in exclusion of underlying sediments so that some strata are bypassed and not recovered in the core barrel. Lubrication of the inner wall of the core barrel can reduce friction and prevent plugging as additional sediments enter the core barrel. Choice of lubricant to reduce friction is recommended as long as the chemical additives do not interfere with chemical analyses planned for the cores (Morton and White 1997). Compaction and loss of material during recovery can also cause discrepancy between penetration and recovery, but occurs less frequently. It is sometimes not possible to correct for shortening when it is not certain whether the shrinkage was due to dewatering or loss of core out of the bottom (Wright et al. 1999). Harvey et al. (1999) found, for example, that compaction may range from negligible in clay-rich cores up to 40% for organic-rich mangrove-dominated cores. The coring technique, including setup and deploying the corer, coring and recovery is quite rapid compared with standard soil boring operations. Usually, a 6-m-long core can be obtained in a matter of minutes under ideal conditions.

In some areas where vibracoring is difficult, rough ideas of sediment composition can be obtained by alternative sampling procedures. Wash borings, for example, can sometimes provide estimates of grain size and composition by flushing out sediments using compressed air (Fig. 11). Jet probes (*q.v.*) also provide rough estimates of bottom sediment types but, like wash borings, they do not provide undisturbed samples that are recoverable for further offsite analysis.

Core Logging and Sample Analysis

During the field data collection phase in large surveys, a preliminary analysis of the cores and samples from the cores is made on a daily basis to obtain information for making advantageous modifications to the survey plan. In addition, when the specific site surveys of high-potential borrow sources are undertaken immediately following the general survey; the preliminary analysis must suffice for selection of these sites, and needs to be as complete as possible. The scope



Vibracore, Fig. 11 A “wash boring” that is operated by jetting down a pipe within a pipe at intervals of 1–2 m and “washing” up a sample with a blast of compressed air between the pipes. This exploratory technique was previously used to provide preliminary information on the quality of sand in a potential borrow area before engaging more expensive vibracoring operations. Although surface sand samples can be easily obtained, it is necessary to know what type of material lies at depth. This method of sampling provided an inexpensive way to sample sediment at depth, which would be “washed up” the pipe and trapped in cloth bags that were used to catch the samples. The heavier materials (gravel, etc.) could not be lifted in the pipe, but divers could “hear” gravel and rocks banging against the casing to establish the presence of unusable coarse materials. Fines could not be trapped because the material would pass through the cloth bags used to trap sediment samples; the “cloudiness” of the water, however, provided experienced divers with a rough indication of how much silt plus clay was present in the sediment. The penetration of the pipe was similar to jet probes that are now used in order to test the depth of unconsolidated sediment. Although crude, this method provided useful information for refining sand searches. The limitations of disintegrated wash borings compared to undisturbed vibracores emphasize the value of cores that can be subsampled, analyzed, and archived. Wash borings today find specialized applications by most researchers (photo courtesy of Coastal Planning & Engineering, Boca Raton, Florida).

of preliminary core and sample analyses in the field is limited by: (1) only partial visual and physical access to the cores for preliminary logging and sampling; (2) the type and extent of sample analyses that are possible in the field, which in many

respects are not comparable to laboratory analysis; and (3) the field analysis for each core that must be completed in a limited time frame in order to keep pace with the progress of the survey.

After each core is taken, the liner containing the cored materials is removed and replaced by a fresh liner. Liners may be clear acrylic plastic tubes that allow observation through the wall. Heavy scratching by granular material or silt and clay particles may smear the inner wall obscuring the contents. Where the cored material is visible, logging and selection of samples can be made. Some corers use aluminum or opaque plastic tubing as core barrels without liners and use the core barrel themselves as the containers using a fresh core barrel for each core run.

Access to cores is best obtained by splitting the cores lengthwise to expose the cored section (Figs. 3 and 7). Prior to sampling or disturbance, the core materials are usually photographed next to a scale. Prior to sampling for laboratory analysis, cores are examined visually to determine pertinent characteristics such as size distribution and composition in terms of relatively broad categories. For this purpose, a hand lens and size comparison charts are useful aids. In addition, samples that appear on visual examination to be possibly suitable as fill material should be further analyzed to obtain data on their size distribution characteristics by more accurate means than visual inspection. This can be done using small-diameter sieves to separate small samples into appropriate size fractions, which are then weighed to determine the percent weight of each size fraction. Minimal equipment needed for this procedure is a drying oven, small-diameter sieves covering the Wentworth W sand size ranges at 1–2 phi intervals, and a small top-loading electronic balance with a precision of at least 0.01 g (Meisburger 1990).

Usually, cores are opened in the laboratory, logged, photographed, and portions are then sampled for various purposes. The presence of carbonate shelly debris or shell layers, bioturbation clasts and infills, gravels, (de)oxygenated surfaces, color changes, and other notable features are recorded prior to sample removal (Figs. 12 and 13). Notations are also made for the presence of epifauna, abraded foraminifera tests, coproliths, or any other unusual biological feature such as insect carapaces and appendages. Molluscan assemblages (including microfossils) are, for example, often removed from the core prior to subsampling for other types of analyses (e.g., Barusseau et al. 1988; Wright et al. 1999). Standard laboratory sediment analysis may be conducted following the sieving methods proposed by Wentworth (1929), by using a rapid sediment analyzer (sand fraction) (e.g., Schlee 1966), or a Coulter Counter (silt and clay fractions) (e.g., Shideler 1976). In addition to standard particle-size analyses, there are numerous other kinds of analysis that can be performed on core materials,

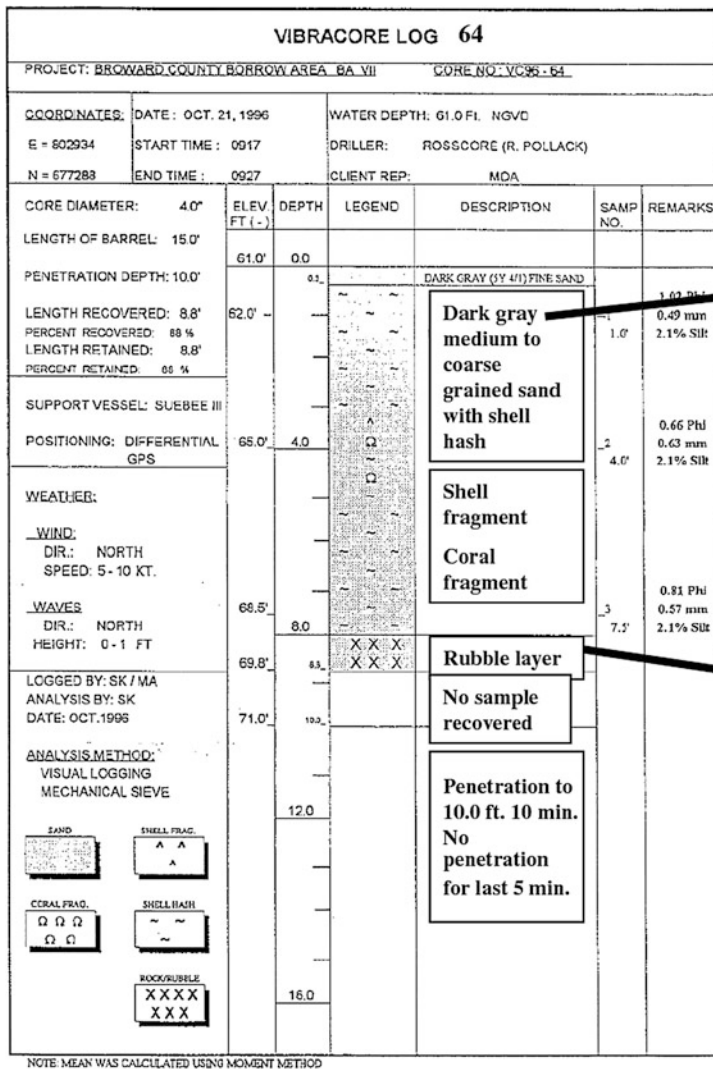
depending on their nature and composition. Organic-rich deposits such as peat, for example, can be vibrocoring taking care to make corrections for compaction (usually –10%) by making stratigraphic comparisons with an uncompacted Eijkelkamp core from the same site (Gehrels 1994). Detrital plant fragments are often selected for dating by the accelerator mass spectrometer (AMS) ^{14}C method (e.g., Gehrels and Belknap 1993), in preference to bulk samples that can be contaminated by humic acids and younger roots (Belknap et al. 1989). Methods for foraminifera sampling and sample preparation are similar to those described by Scott and Medioli (1980).

Bottom Sampling for Coastal Sand Searchers

Samples of seafloor sediments are required for numerous purposes, the least ambitious of which is to gain a rough idea of sediment types. Different kinds of sampling devices are usually geared for the collection of fine or coarse-grained surficial sediments or continuous cores to specified depths. There are a variety of grab-type samplers of different sizes and design that are used for obtaining surficial samples. Most consist of a set of opposing, articulated, scoop-shaped jaws that are lowered to the bottom in an open position and closed by some mechanism. In this process, a sample is retrieved between the closed jaws. Many grab samplers can be deployed by hand while others require lifting gear. Dredge samplers, which can be dragged a short distance along the bottom to dredge up a sample, are sometimes used in place of grab samplers that are subject to losing finer material during recovery when shells or gravel prevent complete closure of grab samplers. Jet probes (*q.v.*) and wash bores are sometimes used on a reconnaissance basis to gain general information on sediment types. While obtaining surficial samples is helpful, it is of limited value because vertical projection of surface data is highly unreliable. Also, the expense of running tracklines for the sole purpose of sampling surficial sediments is not economically justified by the value of the data obtained (Meisburger 1990).

Direct sampling of sub-bottom materials is essential for borrow source identification and evaluation. This is usually accomplished by means of a continuous coring apparatus that can obtain cores 7–13 m in length (cf. Fig. 1). In the types of sediment usually encountered in borrow site exploration, gravity corers are not suitable for obtaining cores of the requisite length, and some form of powered corers must be used. In most cases, vibrator-driven coring devices have been used for this purpose (Meisburger and Williams 1981).

Collection of surface samples is not useful for most geological and geotechnical applications because depth parameters are required for various purposes. Although surficial

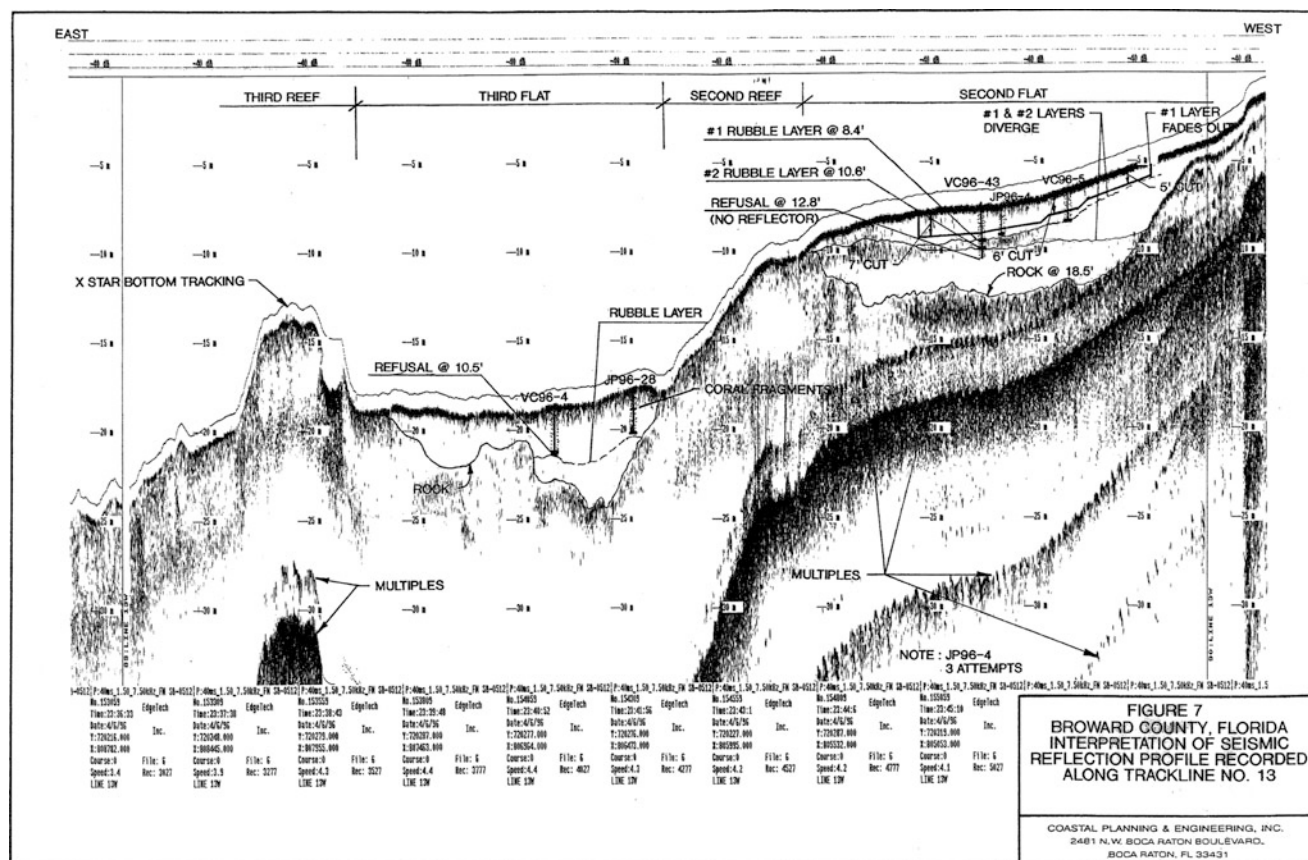


Vibracore, Fig. 13 Example of a vibracore log showing a marked discontinuity in a basal rubble layer. Photographs of the split core (right side of diagram) show medium to coarse-grained sand interspersed with shell hash inclusions that terminate in a very coarse carbonate rubble layer. Sediment loss and refusal of the core barrel occur in the

rubble layer that cannot be penetrated. The value of composite vibracore logs, as displayed here, is immediately apparent because they red flag potential problems when dredging for beach-quality sands that will be mined offshore and pumped onshore for beach replenishment (courtesy of Coastal Planning & Engineering, Boca Raton, Florida).

confidence in selection of offshore borrows. Figure 14 shows an east-west, cross-shore seismic reflection profile (based on a CHIRP X-Star sonar) on the inner continental shelf of south-east Florida in Broward County. The interpreted and annotated sub-bottom profile shows the sedimentary cover that overlies and partially infills inter-reefal troughs, identified as the second and third flats. Even though the seismic survey results can identify rock surfaces of the coral reefs (and buried reef), some carbonate rubble layers are not shown on the trace if they do not present a reflector. Examination of vibracores precisely locates the depth and thickness of coral fragments present in the sandy inter-reefal infills. Summary cross-sections, as shown in the example of Fig. 15, although

based on seismic and geotechnical data, are confidently constructed using closely spaced vibracore logs. As shown in Fig. 15, materials that are unsuitable for beach renourishment (i.e., should not be dredged) are indicated (by a warning red color in the original diagram) as sand with silt, silty sand, gravel, coral fragments, and limestone. Vibracore locations are noted in this fence diagram by a simple coded annotation such as VC-96-61, which indicates a vibracore borehole (VC) that was collected in 1996 (96) as borehole number 61 in a sequence. The strategic role of vibracores as verification or searuthing of geophysical survey data and as key indicators of actual sedimentary conditions in their own right is patently obvious.



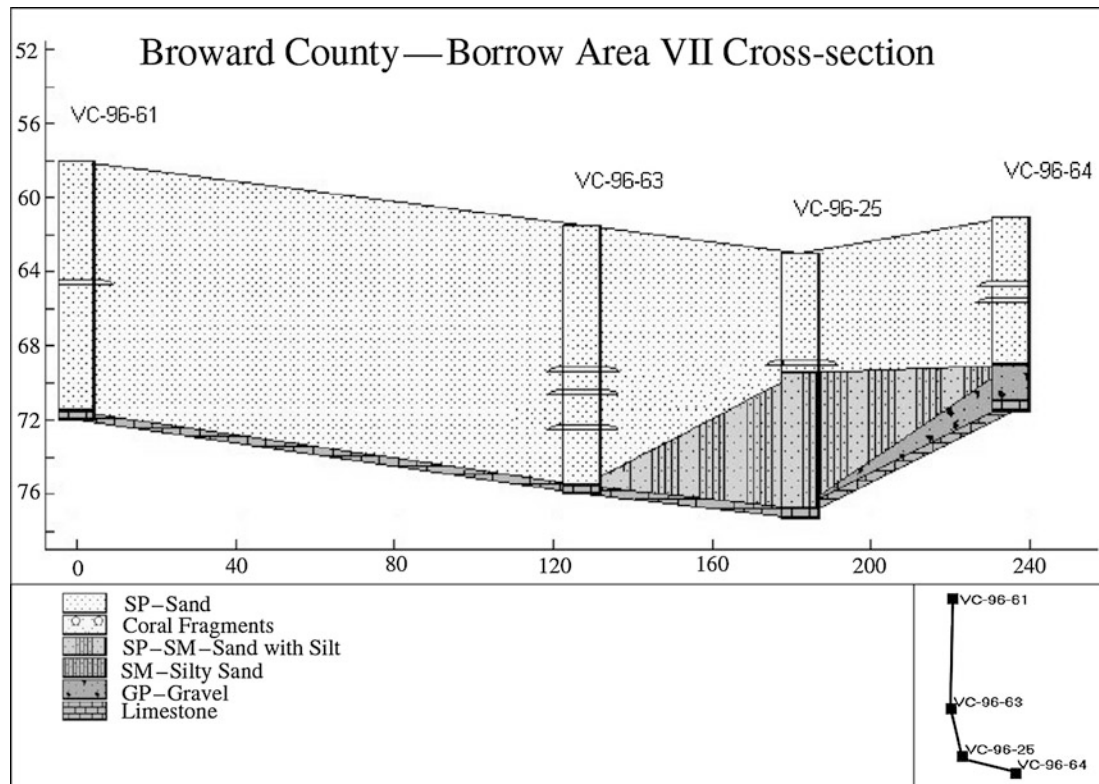
Vibracore, Fig. 14 Cross-shore seismic reflection profile and vibracore locations, southeast coast of Florida in Broward County. This north to south view transects the shore-parallel Florida Reef Tract (so that the east or left side of the diagram leads to deeper offshore water). Depressions between parabathic coral reefs fronting the southeast Florida Peninsula contain sedimentary deposits that range from areas of silt plus clay to coarse rubble accumulations, but most areas supply sandy deposits. The second sand flat lies at about 5 m depth whereas the deeper third flat lies under approximately 17 m of water. The vibracores groundtruth the seismic reflection record by verifying the

thickness of inter-reefal sandy sediments. Note the presence of carbonate rubble layers at depth in the second and third sand flats (shore is to the west, right side of diagram). Information obtained from the vibracores is essential to complete interpretation of seismic data. The presence of carbonate rubble and coral fragments delimits dredging operations to areas within inter-reefal sand flats that are devoid of materials that are unsuitable for beach renourishment. Offshore sand searches often rely on vibracore data for detailed analysis of sedimentary deposits and interpretation of geophysical surveys (courtesy of Coastal Planning & Engineering, Boca Raton, Florida)

Conclusion

Vibracores are specialized sampling procedures for obtaining continuous, undisturbed cores. Although limited by the maximum length of retrievable core (about 7–10 m for most purposes), vibracores find application in many different kinds of coastal studies where undisturbed samples need to be collected from surficial sediments on land or under water. Vibracores can provide invaluable physical, chemical, and biological information that is otherwise unobtainable. Vibracoring systems range from portable, inexpensive setups to more complicated assemblies that require elaborate service platforms on ships or barges fitted with powerful hoisting equipment. Vibracores are an essential component of multifaceted surveys, such as offshore sand searches that attempt to locate potential borrow areas containing large volumes of

beach-quality sands. Although vibracores have a number of limitations that preclude their use under a range of conditions, they are deployed to advantage in many different kinds of scientific and engineering applications. In many respects, vibracores are the unsung heroes of coastal research that depends on acquisition of information posited in the sedimentary record. Collection of unbiased reference information concerning vibracores is somewhat difficult because few papers focus on the vibrocore methodology *per se*. Vibracoring is a sample collection technique that usually only finds mention in the “methods” section of research reports, often as a backup to broader topics such as geophysical or engineering survey. Nevertheless, information obtained from vibracores should not be minimized, as ancillary when in fact in many cases it is primary, even though it is often collected last.



Vibracore, Fig. 15 Example of a typical cross-section constructed from vibracore logs for a borrow area offshore southeast Florida in Broward County. The section shows the presence of inter-reefal sands overlying a limestone base (left side of section) and highlights unsuitable fine-grained (silty) materials and coarse gravels (right side of section) as in core VC-96-25. Fine-grained and coarse-grained materials are

unsuitable for beach renourishment projects and must be identified prior to dredging so they are not placed on the beach. Because such materials are not compatible with native beach sands, their presence reduces the potential of borrow areas and becomes an important factor in estimating offshore reserves of extractable beach-quality sands (courtesy of Coastal Planning & Engineering, Boca Raton, Florida).

Cross-References

- [Beach and Nearshore Instrumentation](#)
- [Beach Stratigraphy](#)
- [Coastal Sedimentary Facies](#)
- [Coastal Soils](#)
- [Jet Probes](#)
- [Mining of Coastal Materials](#)
- [Monitoring Coastal Geomorphology](#)
- [Nearshore Geomorphological Mapping](#)
- [Offshore Sand Sheets](#)
- [Sequence Stratigraphy](#)
- [Shoreface](#)

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Volcanic Coasts

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Volcanism, the ejection of molten magma from the earth's interior onto the surface, is petrologically classified into three types, namely, basic, intermediate, or acidic, depending upon the proportion of silica in the volcanic ejecta. Basic volcanic rocks contain a high proportion of ferromagnesian minerals (Fe Mg) relative to silica (SiO₂), and are thus relatively dense, heavy, and dark colored, and tend to be ejected as relatively nonexplosive and fluid lava streams. Acid volcanic ejecta contains a much higher proportion of silica, is accordingly relatively light and light colored, more viscous,

Terry R. Healy: deceased.

and is ejected explosively. Intermediate volcanic ejecta products are also intermediate in composition and explosive properties. These properties determine the rock hardness and influence the susceptibility of the rocks to physical and chemical weathering.

Basalts are the major basic volcanic deposits. These are typically extruded as fluid lava streams, such as at Kiluea, on the main island of Hawaii (Macdonald et al. 1983). The fluidity of basalt lava normally allows it to exude interstitial gases easily so that the eruptions are relatively explosive-free, and the flowing lava streams follow pre-existing valleys, forming lava pools or sheets in low-lying topography. Often the fluid lava solidifies as massive thick sheets on land, and as a result of the slow cooling, the basalt fractures into hexagonally jointed vertical columns. When fluid basalt is extruded into water the surface cooling is much faster, typically resulting in closely jointed “pillow lavas.” Basalt extruded from a single vent often forms a concave upward volcanic cone due to minor “fire fountaining” of pyroclastic deposits occurring around the vent, and resulting in scoria cone formation. However, little volcanic ash (tephra) is ejected.

Andesite lavas tend to possess greater viscosity, with which is associated greater explosiveness as the lava is ejected. Andesitic eruptions can form thick sheets but when ejected from a single vent tend to be explosive and accompanied by ejection of large volcanic “bombs” into the atmosphere, along with scoriaceous material, and finer tephra material. When ejected from a single vent, the volcano tends to form a mound around the vent. Frequently, however, much of the volcano form is built up from the intermittent eruption of volcanic lava chunks, bombs, and tuff. If these materials become fluidized with waters, they flow as mud-flows or *lahars* down the volcano slopes, ending up as irregular hummocky topography around the lower slopes of the volcano (Cotton 1944). Frequent lahar activity may form a ring plain of laharic deposits – Mt Taranaki in New Zealand is a classic example.

Acid, or *rhyolitic* volcanism, produces gas-rich and viscous lava (Cotton 1944), and typically the volcanic ejecta is explosively extruded as viscous rhyolitic or dacitic lava. In extreme cases of Pelean type eruptions, massive burning clouds of molten incandescent pumiceous and fine ash ejecta, exuding gases, are explosively erupted and reach high into the atmosphere. The explosion cloud of incandescent particles eventually collapses, creating *nuées ardentes*, consisting of hot glowing clouds of self-lubricating gas-exuding pumice and ash, which may flow at great speed across the landscape. When this material pools in topographic lows and cools relatively slowly, the fragmented rock clasts liquefy, weld together, and solidify as *ignimbrite* sheets – but typically of less density and hardness than the andesitic and basaltic flows. Acid volcanic eruptions often result in considerable

ejection of pumice deposits, which may blanket the landscape and are subsequently easily eroded to provide considerable sediment input from the rivers to the coastal littoral zone.

Apart from these “fundamental” types of volcanic deposits, volcanic breccias, and conglomerates can be deposited into the marine environment. The resulting rocks possess a fine tuffaceous matrix of mixed marine and volcanic origin separating the larger breccia clasts of basalt or andesite.

The various volcanic lithologies subjected to coastal marine processes exhibit a wide variety of rock hardness, composition, jointing pattern, as well as existing in a variety of wave energy, tidal range, and climatic (temperature) regime environmental conditions. As expected, a concomitant variety of coastal volcanic landforms result.

Erosion Processes and Geomorphology of Volcanic Rocky Coasts

Erosion processes along rocky coasts are primarily physical, subaerial, and biological. Physical processes include the destructive power of breaking waves, and the consequent hydraulic and pneumatic pressures exerted when broken wave bores are forced along joint planes. Abrasion by the sediments carried by the waves can occur at the base of cliffs and on shore platforms surfaces, but for coastal outcropping volcanic rocks, is generally a minor process. Subaerial processes include weathering within the atmosphere. Particularly potent, especially for fine grained tuffaceous volcanic rocks, is the effect of constant wetting and drying in the intertidal and supratidal zones, a process controlled by the level of permanent pore space saturation of the rock, termed “water level weathering” by Wentworth (1938). Above the level of pore water saturation, subaerial weathering and chemical oxidation predominates. Volcanic breccias can illustrate the importance of subaerial weathering and wetting drying process in shore platform evolution. Wetting and drying is particularly effective on the fine-grained tuffaceous matrix, so that in areas of low wave energy a near-horizontal, high-tide level bench-type shore platform evolves (Fig. 1). On the other hand, where the breccia outcrops and is subjected to high-energy waves, a higher-level irregular notch forms as a result of a higher level of rock pore space saturation, but no horizontal bench occurs.

Biological weathering is typically not an important geomorphic process for volcanic deposits, although it is for certain other coasts such as limestone.

Because volcanic rocks are often massive, indurated, hard, and with coarse jointing patterns, they are typically resistant to the various erosion processes operating at the coast, especially the subaerial and biological processes. Paradoxically a major function of marine action (i.e., the physical power of the breaking waves) is to erode along joint planes and fissures,



Volcanic Coasts, Fig. 1 A high-tide bench-type shore platform with undercut notch cut into a cliff of volcanic breccia. The sub-horizontal platform planed at the high-tide level is due to “water level weathering”. (Photo: T. Healy)

leading to destruction of the platforms and cliffs, and removing the weathering debris from the foot of the cliffs, thereby maintaining a steep cliff profile. The weathered products from the marine erosion of the volcanic rocks are typically of boulder and cobble size, which are often transported by the waves to create coarse boulder and gravel beaches nearby (Fig. 2).

Where unweathered basalts or andesites outcrop at the coast, these hard, dense rocks, are resistant to weathering by marine processes, especially where they possess a low intensity of jointing, that is, the resulting geomorphology is of terminally steep cliffs, perhaps exhibiting incipiently formed narrow high-level benches and topographically irregular platforms. Caves are frequent where the breaking waves have forced along the vertical joints. A classic case is the well-known “Giants Causeway” in Ireland. For tropical coasts, chemical weathering is much more intense, and basalts may be weathered to clays; in such cases broader shore platforms can evolve.

For rhyolite flows, the more viscous lava tends to form “plastered layer mounds” or domes, and these are also



Volcanic Coasts, Fig. 2 Erosion of an andesite cone in a high-wave energy environment, resulting in steep cliffs, and a coarse clastic beach of boulders surmounting a subaqueous shore platform. (Photo: T. Healy)

relatively resistant to erosion by the sea. Over Holocene time since sea level has been at its approximate present level, the sea has often eroded jagged irregular cliffs, as at Mayor Island, New Zealand. Ignimbrites are likewise typically massive with few joints, and thus form high cliffs, as along New Zealand’s Coromandel coast. Compared to basalt and andesite, the rhyolites and ignimbrites are less dense and susceptible to greater weathering rates. Some ignimbrites with high calcium content may evolve coastal geomorphology similar to tropical limestones (Fig. 3) with undercut notches and sub-horizontal, subaerially weathered shore platform surfaces planed to the level of rock pore space saturation.

Erosion of coastal lahar deposits along New Zealand’s high-energy Taranaki coast, results in lahar cliffs, irregular shore platforms, often littered with boulders, small embayments with boulder and cobble beaches, and rugged offshore reefs surmounting submarine platforms. Erosion of the lahar deposits also produce large quantities of

Volcanic Coasts,

Fig. 3 Gently sloping shore platform (foreground) with a veneer of sediment, passing to a high-tide level, sub-horizontal bench and notch cut into ignimbrite in a low-wave-energy environment. The high-tide bench and notches are reminiscent of tropical limestone cut features. (Photo: T. Healy)

**Volcanic Coasts, Fig. 4**

Boulder beach surmounting a broad intertidal platform, formed from erosion of low lahar deposits, Taranaki, New Zealand. In the foreground are remnants of a deposit of pure black titanomagnetite “ironsand,” also an erosion product of the andesitic lahars. (Photo: T. Healy)



titano-magnetite “black” ironsand (Fig. 4) which further updrift produce high dunes mined as an iron ore deposit at Taharoa.

Other Volcanic Landforms at the Coastline

Often specialized volcanic landforms occur coincidentally in the presently active coastal environment. Thus,

large-scale collapsed volcanic vents, or *calderas*, such as Santorini in the Greek archipelago (Fig. 5), or Banks Peninsula of New Zealand, have become modified by stream erosion, and flooded by the sea, forming enclosed harbors (Cotton 1942). Scoriaceous tuff rings and *maars* formed from relatively small-scale phreatic explosions, are relatively easily eroded and the sea can erode through to form a horseshoe-shaped harbor (Searle 1964).

Volcanic Coasts, Fig. 5

View into the caldera at the town of Thera on Santorini, Greece.
(Photo: M. Schwartz)



Active volcanism in the coastal zone can lead to some spectacular visual effects. When lava flows into the sea “pillow lavas” are formed, along with steam and phreatic eruptions, from the rapid cooling of the lava skin in contact with the water, creating closely packed ellipsoidal masses with radial jointing.

Collapsing *nuée ardente* pumice clouds flowing into the sea are a spectacular site and are even believed capable of causing tsunamis from the rapid displacement of the seawater. Likewise a major eruption in the sea can also cause tsunamis (Dudley and Lee 1988; Seibold 1995). In most cases a flank of an active volcano is suddenly uplifted or depressed, thereby generating the tsunami. But for the case of the island of Krakatoa in 1883, the actual violent explosion itself produced waves as high as 40 m crashing ashore in Java and Sumatra, and killing some 36,000 people.

Sediments Originating from Coastal Volcanic Rocks

A range of sediments are derived from weathering and erosion of volcanic deposits in the coastal zone.

- Basalts and andesites rarely weather to sand, but rather to cobbles pebbles and boulders. An unusual example of a specific beach material originating from basalt flows is at Kalapana beach, Hawaii, which comprises black glass fragments from the nearby active Kilauea volcano lava tongue entering the sea (Macdonald et al. 1983).

- Iron-rich heavy minerals weathered from andesitic and rhyolitic deposits may introduce a specific concentration of beach sediments, occasionally in vast volumes. The black ironsand deposits of titanomagnetite on the west coast beaches and dunes of New Zealand’s North Island, result from weathering of the Taranaki andesitic laharic deposits which form the cliffs around Taranaki. But heavy minerals more typically occur as trace rather than bulk minerals; thus they can be used as tracers for the direction of littoral drift and the provenance of the beach sands (Komar 1998).
- Acid volcanism, typically producing softer deposits of pumiceous tephra and ignimbrites, may produce large suites of minerals for sandy beach deposits. Pumiceous tephra, for example, is easily broken down in the fluvial and marine environments to its constituent minerals such as quartz, feldspars, obsidian (volcanic glass), and heavy minerals (including augite, hypersthene, hornblende, cummingtonite, and titanomagnetite). New Zealand’s Bay of Plenty Holocene dune ridge barrier systems provide an excellent example of sand deposits related exclusively to acid volcanic provenance from the nearby Taupo Volcanic Zone.
- Clays weathered from acid volcanic products include allophane, while volcanic glass weathers in very short time to the montmorillonitic type clay, smectite, in shallow marine deposits.
- In areas of coastal dune deposits blanketed by a series of tephra, the sequence of tephra can be used for dating the coastal progradation and evolution. The classic paper illustrating this method is by Pullar and Selby (1971).

Cross-References

- [Cliffed Coasts](#)
- [Gravel Beaches](#)
- [Shore Platforms](#)
- [Weathering in the Coastal Zone](#)

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Vorticity

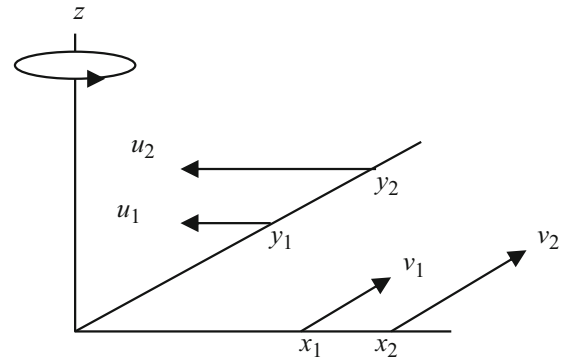
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Definition

Vorticity is the tendency for spin or rotation in a fluid (i.e., vortex flow). As such, it is a vector and can be separated into components of spin about the vertical axis and either or both of the horizontal axes. Eddies observed as water moves past obstacles such as bridge pilings is an example of relative vorticity flow about the vertical axis.

Introduction

Figure 1 shows a Cartesian coordinate system with the x-axis pointing eastward, the y-axis pointing northward, and the z-axis pointing upward in the opposite direction to the gravity vector $\vec{g}$. The x-y plane is parallel to a level surface, and relative vorticity “zeta” (ζ) about the z-axis is defined as $\zeta \equiv \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$. In this equation, $+v$ is flow in the northward direction, and $+u$ is flow in the eastward



Vorticity, Fig. 1. Illustration of relative vorticity ζ in anticlockwise (cyclonic in the northern hemisphere) rotation about the z-axis at a point $x = 0, y = 0$

direction, relative to the fluid in which the eddy is imbedded. Components of relative vorticity can be defined about the other two axes, but the most important one in mesoscale oceanography and meteorology is the vertical component, ζ .

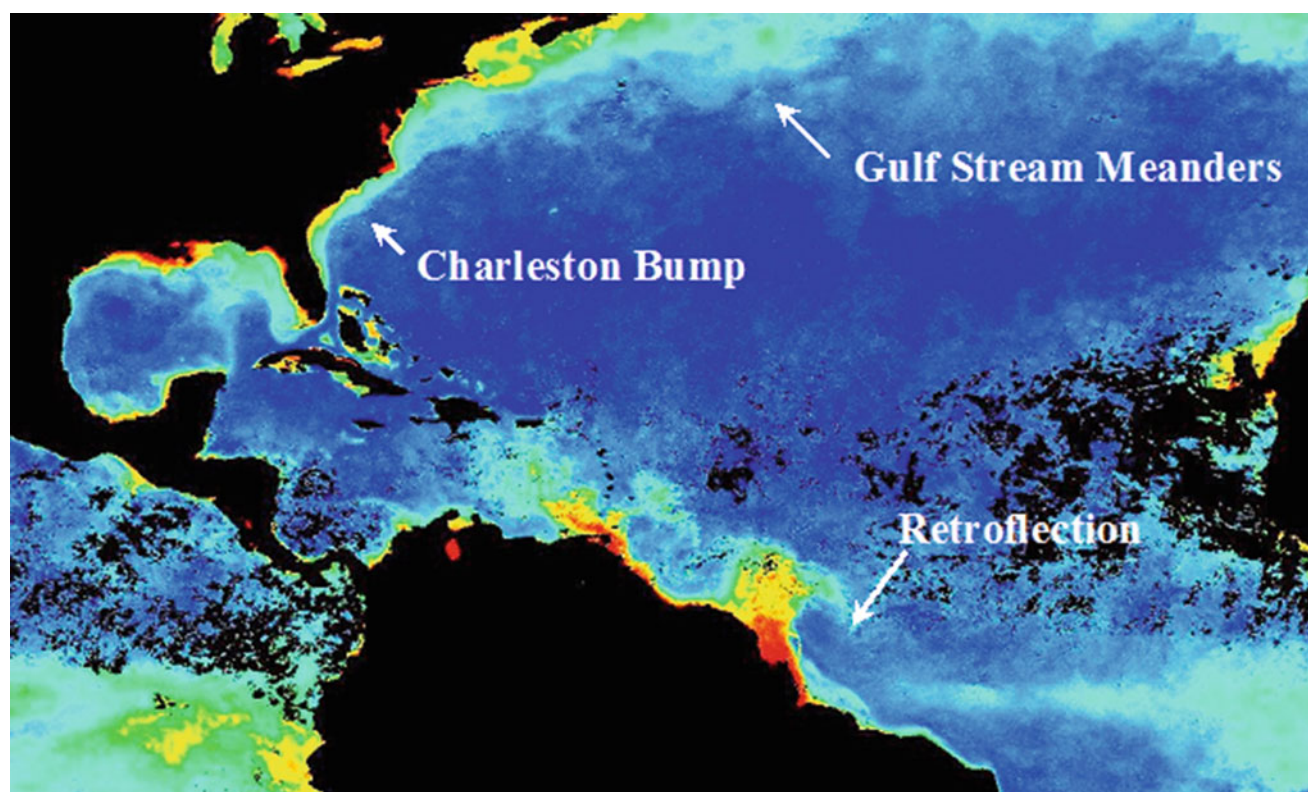
To interpret ζ , imagine a disk rotating cyclonically (counterclockwise in the northern hemisphere) about a spindle ($q.v.$ Fig. 1). Along the x-axis, the farther from the center of rotation, the larger will be the positive v-component of velocity. Similarly along the y-axis, the negative u-component increases as distance from the center increases. On such a disk, both $\frac{\partial v}{\partial x}$ and $-\frac{\partial u}{\partial y}$ are positive quantities, thus defining positive relative vorticity as cyclonic circulation.

Since the ocean and the atmosphere are imbedded on a rotating planet, there is another component of vorticity called planetary vorticity. Planetary vorticity is the familiar Coriolis parameter $f = 2\Omega \cdot \sin \varphi$, where Ω is the rotation rate of Earth, 7.2921×10^{-5} radians per second [or 360° per sidereal day], and φ is latitude. The sum of relative and planetary vorticity, $\zeta_A \equiv \zeta + f$, is known as absolute vorticity.

Absolute vorticity, or more precisely the time change of absolute vorticity, $\frac{d\zeta_A}{dt}$, is related to vertical motion in geophysical fluids. Hydrodynamicists have shown that when following along the path of a parcel of fluid, the so-called Lagrangian perspective, that $\frac{d\zeta_A}{dt} = -\zeta_A D_h$. In this equation, D_h is the horizontal divergence of flow where $D_h \equiv \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}$. This very important expression shows that in the ocean (and atmosphere) changes of absolute vorticity are related to upwelling (rising) and downwelling (sinking), because if a surface flow is divergent, water (air) from below must upwell (rise) to replace the water (air) that is moving away.

Discussion

As an example of the relationship between absolute vorticity time change and vertical motion, consider the Gulf Stream as



Vorticity, Fig. 2 SeaWiFS image of the Intra-Americas Sea showing two regions of potential vorticity conservation: the Brazil Current retroflexion region off the mouth of the Amazon River and the Charleston Bump area of the Gulf Stream

it meanders eastward in the offing of Cape Hatteras, North Carolina. For simplicity, think of the meanders as horizontal waves and that changes in f are small. As a parcel of fluid moves from the crest of a meander where ζ_A is negative toward the trough where ζ_A is positive, the time change $\frac{d\zeta_A}{dt}$ is positive, and thus D_h is negative. Negative divergence is convergence, and thus there is an area of downwelling water between the meander crest and trough. Similarly, as the parcel moves from the trough to the next meander crest, $\frac{d\zeta_A}{dt}$ is negative, and an area of upwelling is observed.

When the ocean is regarded as a barotropic fluid and D_h is integrated from the surface to the bottom where the water is H deep, yet another form of vorticity can be written: $\frac{\zeta+f}{H} = \text{constant}$. This ratio is known as potential vorticity, and it is a conservative property of the ocean and the atmosphere. This very important equation shows that geophysical fluids change direction when either its latitude φ changes or when the water depth H changes.

Two well-known observed examples of conservation of potential vorticity are found in the Atlantic Ocean (Fig. 2). First, as water from the southern hemisphere flows across the equator in the Brazil Current, the Coriolis parameter f changes from negative to positive. If H is constant, then ζ must become negative in order to conserve the sum $\zeta + f$. Negative relative

vorticity is anticyclonic flow, and thus the water must turn eastward as it passes north of the equator. This is the formation of the Brazil Current retroflexion that eventually leads to the North Equatorial Countercurrent in the Atlantic Ocean.

A second example is found in the Gulf Stream off Charleston, South Carolina. As the Gulf Stream flows northward, it encounters a shallow bottom topography feature known as the Charleston Bump. In order to conserve potential vorticity, with f approximately constant, a decreasing H must be balanced by a negative ζ . As with the retroflexion region, negative relative vorticity ζ implies that the Gulf Stream must turn anticyclonically offshore and flow eastward. However, H increases quickly as the current enters the deep water of the continental slope, and ζ changes again, causing the Gulf Stream to meander cyclonically onshore. Thus conservation of potential vorticity initiates meanders in the Gulf Stream that are amplified downstream off Cape Hatteras.

Conclusion

The four components of vorticity in geophysical fluids are relative vorticity (ζ), planetary vorticity ($f = 1.4584 \times 10^{-4} \sin \phi$), absolute vorticity ($\zeta_A = \zeta + f$), and potential vorticity

$\left(\frac{\zeta+f}{H}\right)$. The units of vorticity are s^{-1} , i.e., radians per second.

A related motion is corkscrew-like, called helicity.

Cross-References

► [Gulf Shorelines, Last Eustatic Cycle](#)

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Washover Effects

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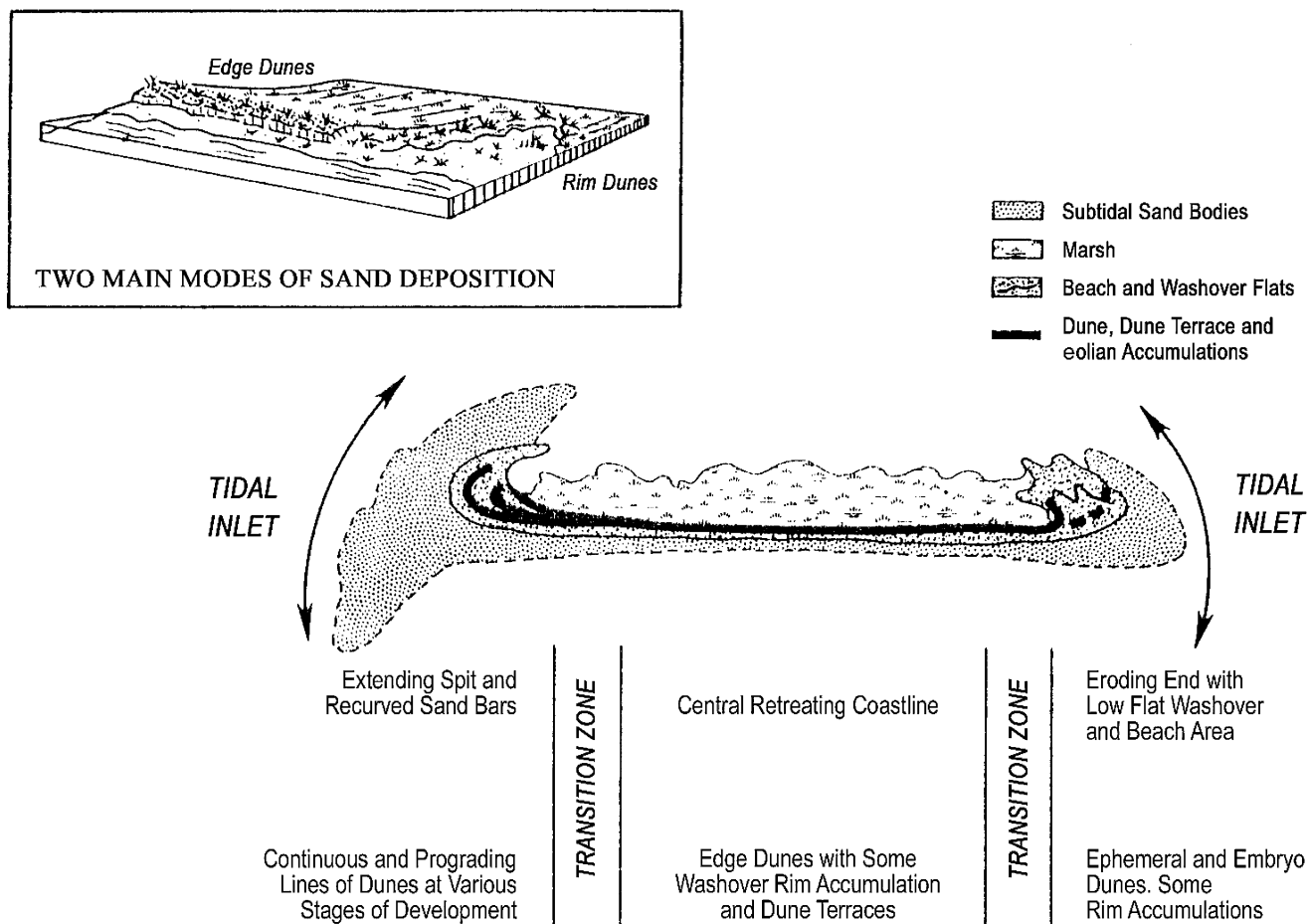
Although there is extensive geomorphological literature on barrier islands (e.g., Schwartz 1973) and a similar volume of work on overwash processes (e.g., Leatherman 1981), studies of the barrier islands associated with present and past positions of the Mississippi Delta (Penland et al. 1988), where there is accelerated subsidence and an absence of a sufficient supply of sand, have demonstrated the critical role of sand dunes in the evolutionary sequence (Ritchie and Penland 1990a; Ritchie 1993). This model may have wider applications as evidence accumulates for greater coastal submergence as a consequence of a possible global rise in sea level (Bird 1993).

In addition to the normal processes of coastal dune development associated with beach-dune exchanges, barrier island dunes and subeolian sand terraces have three characteristics which relate to the typical evolution of barrier islands; these are: rapid change, the importance of position (i.e., at the center or at the flanks of the island, Ritchie and Penland 1990b) and the significance of overwash processes. (Other factors such as sand supply and patterns of vegetation are equally important but are not considered further here insofar as they are not unique to barrier island dunes, although according to Goldsmith (1985), "The most important contribution that humans can make toward the preservation of barrier islands is to prevent damage to the dune vegetation.") Similarly, the evolution of transgressive or regressive barrier islands need not be described. Nevertheless, for reasons such as a rapid reduction in sand supply or tectonic subsidence, a relative rise in sea level will, typically, enhance the frequency and therefore the significance of washover processes (Viles and Spencer 1995).

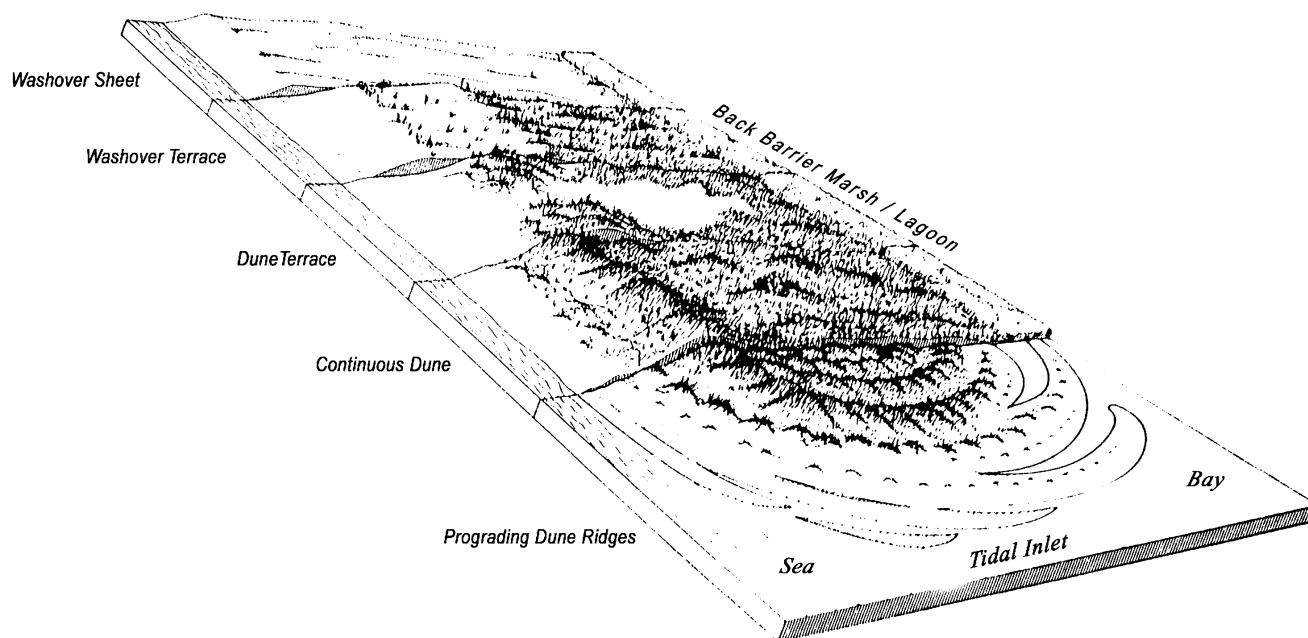
The model barrier island (Fig. 1) will retreat landwards and migrate alongshore depending, primarily, on the strength and direction of coastal currents. Hydraulic processes at tidal inlets also control the shape and rate of change. Larger beaches occur at the ends of the island and are nourished by alongshore transport and from accretionary berms and sand bars which are formed by flood and ebb tidal flows. This pattern of sand movement along the beach toward the flanks of the barriers produce large, sometimes multiple accretionary dunes of several types. In contrast, central areas, unless there is abundant sand supply from the beach or exceptional relict features such as old sand hills or beach ridges (cheniers in Louisiana, Ritchie 1972), are typically eroding. Thus, in the central part of the barrier dunes are often little more than narrow, low, ephemeral accretions along the coastal edge.

Where overwash processes occur, usually as tongue-shaped penetrations leading to a depositional fan, a surface of bare sand is produced which, on drying, provides a local source for wind transport with the sand being trapped at the line of vegetation along the perimeter of the sand flat. In time this surface will revegetate to become a terrace which may have some additional wind-blown sand accumulations from the beach or from local coastal edge erosion and other redepositional processes. Although it is often a fine distinction, the terminology which has been developed consists of "washover terrace" if low and flat and "dune terrace" if higher and undulating. Dune terraces might also originate as thin spreads of sand which form the backslope of most types of coastal dunes. From numerous surveys of barrier island dune systems, including investigations of the rate of change, a generalized model of dune types has been constructed (Fig. 2) for the Louisiana barrier islands.

The rate, frequency, and penetration of overwash events is highly variable and are functions of hurricane, storm, and subtropical frontal passages. The preexisting height of the dune barrier and the width of the beach are the prime defenses against overwash. Thus, the high dunes at the ends of the barrier may be surmounted once every 5 to 10 years whereas



Washover Effects, Fig. 1 Generalized location of eolian accumulation forms



Washover Effects, Fig. 2 Dune types associated with barrier islands in south Louisiana

the low coastal terraces may be crossed by surges several times every year (Ritchie and Penland 1990a). Overwash processes transfer sand from the beach and dunes inland thereby contributing to landwards retreat. If these processes are dominated by extreme events such as the passage of a hurricane or a severe storm, consequential topographic orientations can be at different angles to the orientations which are produced by normal eolian processes. At the north end of the Chandeleur Islands, for example, the biggest, dune ridges run inland at 90 to the coastline, being a combination of post overwash residual features and later eolian deposition. It is often possible on the barrier islands of coastal Louisiana to detect relict washover features including lagoon and bay deposits which can be correlated with known hurricane events over the last 20 to 30 years.

Where severe storm events also produce significant elevation of sea levels which are normally accompanied by very strong storm wave effects, barrier island dunes, of all types, may be eliminated completely and the mass of sand which is stored in these features is translated landwards as sheets and spreads some of which cross the entire width of the island and extend as delta-like deposits into the back – barrier lagoon or bay – and as such are lost to the active coastal beach–dune zone. If these back-barrier spreads of sand are laid down during exceptional high water events (and this is normally true), when the water level subsides they will be colonized by vegetation and remain as visible features until the next extreme event. Complete planation of the barrier island can occur at the penultimate stage of the barrier island cycle (Penland et al. 1988) but in younger, more sand-rich barrier islands, the higher sand dunes, normally at the downdrift end of the island survive, albeit severely eroded and often with local washover penetrations through preexisting low and weaker sections. Accordingly, the importance of dune and similar forms to the maintenance and evolution of barrier islands cannot be over-emphasized but these dunes (unlike coastal dunes in non-barrier coastlines or in higher latitudes or in areas lacking frequent periods of significantly elevated sea levels) change, with great rapidity, not solely as a consequence of eolian processes but also as a consequence of powerful episodic effects of overwash surges. In most locations, however, it is possible to detect cycles of washover activity, with the 7- to 10-year return period of hurricanes of the Louisiana coast being the shortest (Ritchie and Penland 1990a). It is therefore necessary to conclude that eolian coastal dune processes need to be considered alongside overwash processes in regions of rapid barrier island submergence which, if the predictions of global rises in sea level are correct, might be found, increasingly, in areas other than the margins of large deltaic systems.

Extensive surveys of all the barrier islands of Louisiana were undertaken by the Louisiana Geological Survey and published in four regional volumes, that is, Isles Dernieres, 1989; Plaquemines, 1990; Chandeleur Islands, 1992; Lafourche Barrier System, 1995.

Cross-References

- [Barrier Island Landforms](#)
- [Beach Processes](#)
- [Beach Ridges](#)
- [Changing Sea Levels](#)
- [Cheniers](#)
- [Dune Ridges](#)
- [Eolian Processes](#)
- [Sea-Level Fluctuations over the Last Millennium](#)
- [Storm Surge](#)
- [Tidal Inlets](#)

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Water Quality

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Definition

Coastal waters are used for a variety of purposes such as food production, hydroelectricity generation, recreational activities, as a transport medium, and as a repository for sewage and industrial waste. For some of these purposes, primarily, food production and recreational activities, the quality of the water is vitally important. The water environment and

associated habitats such as coral reefs and beaches may offer significant financial benefit to the associated communities and these opportunities provide the motivating factors for creating a management program.

Water, however clean, is an alien environment to man and thus it can pose hazards to human health even when it is of pristine quality. It is therefore necessary to implement and enforce effective management policies in order to minimize and reduce the health consequences of all anthropogenic activities related to water use. In addition to human health implications of poor water quality, anthropogenic risk factors include mortality and reduced growth of the reef-building corals due to their high sensitivity to rising seawater temperatures, ocean acidification, water pollution from terrestrial runoff and dredging, destructive fishing, overfishing, and coastal development (Burke et al. 2011).

Factors Affecting the Quality of Water

Water quality is a multi-faced concept. In addition to microbiological factors, monitoring of coastal waters used for recreational or shellfish harvesting purposes includes the observation of turbidity or transparency, the presence of floating objects, foams on the surface and odor. Apparently clean waters, that is, those that appear clear, may in fact be polluted with microorganisms or chemicals.

Coastal waters have traditionally been considered as the ultimate sink for the by-products of human activities. The key sources of pollutants affecting coastal water quality are riverine inputs of domestic, agricultural, and industrial effluents and direct sewage discharges from the local population. Discharges may be regular through long and short sea outfalls and irregular through storm water and overflow outfalls, and unregulated private discharges. It is the point sources of pollution due to domestic sewage discharges that generally cause most human health concern but is perhaps easier to control than diffuse pollution. Discharge of sewage to coastal waters exert a variable polluting effect that is dependent on the quantity and composition of the effluent and on the capacity of the receiving waters to accept that effluent. Thus, enclosed, low-volume, slowly flushed water systems will be affected by sewage discharges more quickly than water bodies that are subject to fast change and recharge.

Pollution originating from water-based recreation such as leisure crafts, jet skis, etc. includes sanitation discharges, environmentally toxic effects of antifouling paints and general debris. Estuarine water will be particularly vulnerable to pollution due to the enclosed nature of the system and the subsequent accumulation of pollutants. Controls on sanitation discharges from pleasure craft are dependent on suitable holding tanks and port reception facilities (see entry on “► Marine Litter”).

Water quality may also be adversely affected by fuel spillages and discharges from pleasure craft. The oil spill that took place as a result of tanker ship Exxon Valdez in the region of Alaska in March 1989 is supposed to be one of the worst oil spills ever in the history of maritime industry. Out of the total capacity of 58 million gallons, 10.9 million gallons was spilled into Alaskan Coast line. Oil, petrol, and diesel may be spilt at filling barges and bilge waters may be discharged, adding to pollution of the coastal water. In addition to the fuels themselves, the emissions from the engines may be harmful and the oil and gasoline mix in two-stroke fuels has been implicated in the contamination of fish and shellfish products.

Antifouling paints applied to boats and coastal installations to prevent the subsequent attachment and growth of organisms may have an undesirable effect on water quality, especially where the area is closed or where the vessels or installations are concentrated.

Aesthetic pollution of water quality stems from a variety of sources such as recreational users, leisure boats, fishermen, etc. It has been estimated that 70–80% of marine debris comes from land-based sources (Pond and Rees 1994) (see entry on “► Marine Litter”). More recent reports fail to show any degree of improvement in the range and abundance of marine debris (GESAMP 2015).

Environmental factors such as rainfall or sunshine may also significantly influence the microbiological quality of water. Heavy rains can cause sewage overflows thus bypassing sewage treatment facilities, increasing surface water runoff and directly influencing microbiological water quality. In contrast, as the intensity of sunlight increases, some microorganisms (e.g., *Escherichia coli*) die-off rapidly.

Salinity is another factor that has been shown to affect the die-off of microorganisms. Hanes and Fragala (1967) suggest that enterococci, for example, die-off more rapidly in saline water than in freshwaters.

Eutrophication (the enhancement of the natural processes of biological production, caused by increases in levels of nutrients) has been recognized as a growing concern since the 1950s (see for example Diaz and Rosenberg 2008). The most visible indicator of coastal eutrophication is extensive blooms of phytoplankton and/or benthic macroalgae. In many regions, eutrophication in large water bodies has been characterized by blooms of planktonic species such as *Phaeocystis* (the North Sea; Davidson and Marchant 1992), *Noctiluca miliaris* (the Black Sea; Mee 1992), and mucus aggregates (the northern Adriatic Sea; Stachowitsch et al. 1990). Some of the algae are toxic and may cause fish kills while others are aesthetically “nuisance” algae, causing spoiling of beaches, offensive odors, and slimy water. Many cause secondary blooms of undesirable fauna, such as the ctenophore *Mnemiopsis leidyi* (the Black Sea; Mee 1992) which has reached bloom biomass densities of 1 kg m^{-2} .

In extreme cases, toxic species of dinoflagellates, diatoms, nanoflagellates, and cyanobacteria occur and have led to human health impacts after swimming or consumption of shellfish and fish (see for example Funari and Testai 2008). In Tasmania and Victoria, the closure of shellfish beds has resulted from blooms of algae suspected to have been introduced into Australia in ballast water (Hallegraeff and Bolch 1992).

Management of nutrient loading to coastal environments, and any subsequent problems of eutrophication, is based on the reduction of nutrient-rich effluents from the land or better dispersion of existing discharges. This may be achieved by the control of soil erosion, changes in the use or nature of fertilizers, reuse rather than discharge of nutrient-rich effluents, diversion of discharges into less-sensitive or better-flushed environments, engineering works to improve flushing, and nutrient removal from effluents.

Human Health Implications of Poor Water Quality

Epidemiology is the scientific study of disease patterns in time and space (Kay and Dufour 1999). Epidemiological studies have played an important role in providing information to characterize risks associated with swimming in waters of poor quality and of particular relevance, those risks associated with swimming in feces-contaminated recreational waters. Although the water environment is of a highly variable nature, a number of credible studies have been carried out since 1953 (see Kay and Dufour 1999 for a review), which have shown an association between a variety of symptoms and exposure to fecally polluted recreational waters.

Coastal waters generally contain a mixture of pathogenic and nonpathogenic microbes derived from sewage effluents, the population using the water for recreational purposes, industrial activities, and farming and wildlife, in addition to indigenous microorganisms. If an infective dose of pathogens colonizes a suitable growth site in the body of a bather, there is the possibility of disease.

There are a broad variety of illnesses that have been associated with swimming in marine and fresh recreational waters. *E. coli*, *Leptospira*, Norwalk virus, and adenovirus 3 are some of the microbes that have been linked to swimming-associated disease outbreaks (see Pond 2005). A number of outbreaks of *Giardia* and *Cryptosporidium* have been shown to occur in very small, shallow bodies of water that are generally frequented by children. Epidemiological investigations of the outbreaks found that the source was usually the bathers themselves, most likely children (WHO 1999).

Enteric illness, such as self-limiting gastroenteritis, is the most frequently reported adverse health outcome investigated in recreational users of contaminated bathing waters, and it is

now accepted that there is an association between gastrointestinal symptoms and indicator-bacteria concentrations in recreational water. Most of the literature suggests a causal relationship between increasing recreational exposure to fecal contamination and frequency of gastroenteritis (WHO 1999). Higher symptom rates have been reported usually in the lower age groups (Cabelli 1983; Pike 1994). Children are particularly at risk from waterborne diseases compared to other age groups due to weak body defenses, susceptibility, and inadequate or no understanding of how to avoid hazards. Children have a natural curiosity and disregard for safety and thus may play in or drink contaminated water. As children's mobility increases they expose themselves to more hazards thus increasing risk of infection.

Escherichia coli is a well-known indicator of fecal pollution. Enterotoxigenic *E. coli* cause a dehydrating diarrhea in children; enteroinvasive *E. coli* can cause fever and watery mucoid diarrhea in infants; enterohemorrhagic *E. coli* is associated with hemorrhagic colitis, some cases progressing to hemolytic uremic syndrome, particularly in children. Other associations, although fewer, have been reported for other health outcomes, such as ear infections and respiratory symptoms. There is less evidence for the link between recreational water use and more serious diseases although it is biologically plausible. Cholera, for example, can be transmitted through water and food. Environmental sources of the causative agent include sewage, sea, and surface waters. Direct person-to-person transmission is uncommon.

Giardiasis is a common disease throughout the world, affecting all age groups, although the highest incidence is in children (Hunter 1997). Waterborne outbreaks are particularly common and a few outbreaks have been reported from recreational waters.

Meningitis, hepatitis A, typhoid fever, and poliomyelitis could also be acquired through bathing but it is difficult to determine unequivocally.

Limited evidence suggests that local populations bathing in sewage-contaminated water may suffer less illness than visiting populations, probably due to the influence of the immune status (WHO 1999). It has also been suggested that viruses could be transferred from the water to the air and therefore nonswimmers could be at a certain risk (Feachem et al. 1982).

Water Quality Standards

Coastal water quality standards relate primarily to recreational waters and water used for shellfish production. The implementation of water quality standards have had some notable successes in promoting cleanups, increasing public awareness, contributing to informed personal choice, and to a public health benefit. In England, for example 98% of

designated bathing waters met the standards required by the European Commission in 2016 (<https://www.eea.europa.eu/themes/water/status-and-monitoring/state-of-bathing-water/state-of-bathing-water-3>).

Water quality is primarily measured against standards for microbiological parameters and present regulatory schemes for the microbiological quality of recreational waters are primarily or exclusively based on compliance with fecal

indicator counts. A variety of water quality standards exist throughout the world (Table 1). The World Health Organization (WHO) have developed Guidelines for Safe Recreational Water Quality Environments (WHO 1998). To comply with standards of quality, authorities must monitor the water a minimum number of times during the defined bathing season. There is considerable concern about the financial cost of monitoring water quality, especially in the light of the

Water Quality, Table 1 Microbiological quality of water guidelines/standards per 100 ml. (Adapted from WHO 1999)

Country	Shellfish harvesting		Primary contact recreation		
	Total coliforms	Fecal coliforms	Total coliforms	Fecal coliforms	Other
Brazil		100% <100	80% <1,000 ⁱ	80% <1,000 ⁱ	
Columbia			1,000	200	
Cuba			1,000 ^a	200 ^a	
				90% <400	
EEC, Europe			80% <500 (guide)	80% <100 (guide)	Fecal streptococci 100 (guide)
			95% <10,000(mandatory)	95% <2,000 (mandatory)	Salmonella 0/L (mandatory)
					Enteroviruses 0
					PFU/L (mandatory)
					Enterococci 90% <100
Ecuador			1,000	200	
France			<2,000	<500	Fecal streptococci <100
Israel			80% <1,000 ^c		
Japan	70		1,000		
Mexico	70 ^b		80% <1,000 ^c		
	90% <230		100% <10,000 ^g		
Peru	80% <1,000	80% <200	80% <5,000 ^c	80% <1,000 ^h	
	0	100% <1,000			
Poland					<i>E. coli</i> <1,000
Puerto Rico	70 ^d			200	
	80% <230			80% <400	
United States,	70 ^b		80% <1,000 ^{e,f}	200 ^{a,f}	
California			100% <10,000 ^g	90% <400 ^h	
United States,		14 ^a			Enterococci 35 ^a (marine)
USEPA		90% <43			33 ^a (fresh)
					<i>E. coli</i> 26 ^a (fresh)
Former USSR					<i>E. coli</i> <100
UNEP/WHO		80% <10		50% <100 ^j	
		100% <100		90% <1,000 ^j	
Uruguay				<500 ^j	
<1000 ^k					
Venezuela	70 ^a	14 ^a	90% <1,000	90% <200	
	90% <230	90% <43	100% <5,000	100% <400	
Yugoslavia			2,000		

^aLogarithmic average for a period of 30 days of at least five samples

^bMonthly average

^{c,d}Minimum 10 samples per month

^ePeriod of 30 days

^fWithin a zone bounded by the coastline and a distance of 1,000 ft from the coastline or the 30-foot depth contour, whichever is further from the coastline

^gNot a sample taken during the verification period of 48 h should exceed 10,000/100 ml

^hPeriod of 60 days

ⁱSatisfactory waters, samples obtained in each of the preceding weeks

^jGeometric mean of at least five samples

^kNot to be exceeded in at least five samples

precision with which the monitoring effort assesses the risk to the health of water users and the effectiveness with which it supports decision-makers to protect public health. There is therefore a move towards a risk-based approach to management through bathing water profiles. These profiles contain, for instance, information on the kind of pollution and sources that affect the quality of the bathing water and are a risk to bathers, health.

In addition to microbiological hazards there are a number of other diverse physical and chemical pollutants that should be addressed in monitoring programs where they are known or suspected to be locally significant. It is important that standards monitoring and implementation enable preventative and remedial actions that will prevent health effects arising from such hazards.

Water Quality Determination

The wide variety of factors affecting the quality of coastal water (together with the vagaries associated with monitoring) makes it inappropriate to compare the quality between different coastal locations. However, individual studies have highlighted particular problems in various regions. For example, Ozkoc et al. (1997) investigated the specific issues affecting the quality of the Black Sea. Land-based pollutants were considered the main pollution source in this case.

Jensen et al. (1997) investigated the sources of pollution and other impacts for the Mediterranean Sea, the Gulf of Suez and the Red Sea, and Gulf of Aqaba. The study found that industrial wastewater and domestic sewage from residential and tourist areas were a major problem in the Mediterranean Sea. Oil pollution from refineries, ships, and offshore oil platforms were considered the main pollutants for the Gulf of Suez and sewage, oil pollution, landfilling, dredging, and siltation the main contributors to poor water quality in the Red Sea and the Gulf of Aqaba.

Standard methods exist for the determination of the microbiological parameters that identify the quality of the water according to current legislation and methods (see e.g., ISO 1992, 1997). Many waterborne pathogens are difficult or expensive to detect and/or quantify and instead present regulatory schemes for the microbiological quality of recreational waters are primarily based on percentage compliance with fecal indicator organisms (organisms that are always found in feces and which are easy to detect on simple bacteriological media) to show the potential presence of organisms that cause gastrointestinal disease (Table 1). Fecal streptococci are suggested as the recommended indicator for saltwater and either fecal streptococci or *E. coli* can be used for monitoring freshwaters (WHO 1999). Sulfite-reducing clostridia may be suitable as indicator organisms for parasitic protozoa and viruses from sewage-impacted waters (Ferguson et al. 1996).

Other indicator organisms for sewage, include the bacteriophages to *Bacteroides fragilis* HSP40, somatic coliphages, F-specific RNA phages, and fecal sterols. The *B. fragilis* phages appear to survive in the same way as human enteric viruses against a variety of environmental conditions, but they may exist in lower numbers than various coliphages, and it is thought that only between 1% and 5% of humans excrete these phages and therefore they may be unsuitable indicators in areas of small communities.

Aesthetic parameters can act as an immediate indicator of sewage contamination. However, the presence of sanitary plastics, for example, may reflect old sewage contamination and may not be a good indicator of human health threats.

No one indicator organism is suitable for representing all the issues associated with fecal contamination of recreational water.

Monitoring only for microbiological parameters does not provide the most accurate and feasible index of health risk. There are a number of acknowledged constraints associated with the current standards and guidelines:

- Management actions are retrospective and can only be instigated after users have been exposed to the hazard.
- The risk to human health is mainly from human excreta, the traditional indicators of which may derive from other sources.
- There is poor inter-laboratory and international comparability of microbiological analytical data.

As discussed by Fewtrell and Kay (2015), a useful way of finding the source of pollution is quantified microbial source apportionment linked with microbial source tracking methods.

Knowing the possible sources of contamination and their likely influence upon water use provides a fast and robust means to increase the reliability of the overall assessment. This could lead to a classification system based on relative risk as suggested by the WHO and United States Environmental Protection Agency (USEPA) (WHO 1999).

Conveying Information on Water Quality to the Public

The primary reason for monitoring water quality and for informing the public is for the protection of public health. The public is unlikely to want to know the technical details of sample treatment in the laboratory but wish to know only whether the water is safe or not. It is essential that any information relating to water quality is presented to the public in a clear, unambiguous, and easily understood way.

Raising awareness and enhancing the capacity for informed personal choice is an important factor in ensuring the safe use of coastal water environments. It acts directly (by allowing users to choose an area which is known to be safe) and indirectly (the exercise of preference for safer environments will encourage investments in improvements). In order for it to be successful, it is essential that the public is generally aware and that the information is available and easily understood.

Currently, a number of schemes exist which convey information to the public. Unfortunately, as yet the public is often provided with information on the previous bathing season's water quality rather than current results. For example, in the European Community, the EC Directive on the quality of bathing waters requires that Member States submit to the Commission a comprehensive report on their bathing water and the most significant characteristics. The Commission then publishes this information by means of a report just before the beginning of the next bathing season (<https://www.eea.europa.eu/themes/water/status-and-monitoring/state-of-bathing-water/state-of-bathing-water-3>). Real-time information services do exist – generally provided by local authorities for individual beaches.

Other schemes to inform the public of water quality exist; for example, award schemes are common in Europe the best known is European Blue Flag scheme. Many countries have national equivalents. The quality of recreational water is often used in publicity to attract visitors to the area. Award schemes are often used as incentive programs to involve all parties concerned in participating in optimizing beach safety, water quality, and education activities (see entry on “► [Rating Beaches](#)”).

Pollution Abatement and Water Quality

Water quality is often affected by pollution due to sewage and industrial discharges, combined sewer overflows, and urban runoff. Pollution abatement is therefore a key part of coastal zone management aimed at minimizing both health risks to bathers and ecological impacts. The main action to reduce sewage pollution requires a large investment in sewage treatment and discharge. Pollution abatement measures for sewage may be grouped into three alternatives: wastewater treatment, dispersion through sea outfalls, discharge to non-surface waters (reuse). The traditional measures are often prohibitively expensive and therefore in some cases in view of costs of control, it may be preferable for integrated beach zone management to focus on alternative options such as restricting beach use or warning the public of the potential health risks during and after risk events (see entry on “► [Coastal Management Practices](#)”).

Summary

Exposure to fecal pollution through contaminated waters leads to detectable health effects and there is clear evidence of a dose-response relationship linking fecal pollution with enteric and non-enteric illnesses. Evidence also indicates that health effects linked to pollution occur at levels of fecal indicator bacteria which are found in recreational waters globally and which may be below the legal standards in many parts of the world. In order to manage coastal water quality, national authorities should take account of a variety of factors, including social and economic factors, some of which are conflicting, in order to ensure safe waters.

Cross-References

- [Coastal Management Practices](#)
- [Environmental Quality](#)
- [Human Impact on Coasts](#)
- [Marine Debris-Onshore, Offshore, and Seafloor Litter](#)
- [Rating Beaches](#)

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Wave Climate

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Definition

Ocean waves may be defined as periodic undulations about an interfacial reference level, and as such can move horizontally, vertically, or at angles inclined to the horizontal. In the simplest sense, waves transport energy without transporting mass. Sound waves, electromagnetic waves, and sand waves, among others, are also part of the oceanic realm but are not dealt with herein.

Introduction

Most human experience is with those ocean waves visually described as vertical undulations about the air-sea interface.

These are generally known as *gravity* waves because gravity is the restoring force. Oceanographers also recognize internal waves, Rossby waves, tidal waves, capillary waves, solitons, Kelvin waves, and many other complex periodic motions. Table 1 is a brief summary of the commonly recognized wave types in the marine environment. This article however will focus on those waves generally known as *sea* and *swell*, because most of the wave energy in the ocean is contained in such gravity waves – of which sea and swell are the major types.

Waves are characterized by wavelength, height, period, and steepness. Wavelength (L) is the horizontal distance from crest to crest or trough to trough. Wave height (H) is the vertical distance from crest to trough, and steepness is the ratio H/L . When steepness exceeds $1/7$, waves tend to break as is seen routinely at a beach. Period is the time for two successive crests or troughs to pass a fixed point, the reciprocal of which is frequency. Significant wave height ($H_{1/3}$) is the height of the highest $1/3$ of the waves in a large number of waves as estimated by an observer. The human eye seems to pick $H_{1/3}$ as the observed wave height.

The effect of the wind on water is to produce capillary waves (which are seen as ripples), ultragravity waves (which are seen in sunglint), sea (which are the shorter-period gravity waves), and swell (which are the longer-period gravity waves). Tsunami waves and seiches are examples of infragravity waves, storm surges such as caused by tropical storms are generally thought of as long-period waves, and transtidal waves are the realm of Kelvin waves and Rossby waves. The classification boundaries in Table 1 should not be held too rigid as many waves extend across a broad range of periods. The general category of *wind waves* includes capillary and ultragravity waves, sea, and swell. *Sea* is the result of the direct action of wind on the ocean, and *swell* is caused by a nonlinear transfer of energy from shorter-period waves toward longer-period waves.

Table 2 is an adaptation of the Beaufort scale using verbiage from the World Meteorological Organization

Wave Climate, Table 1 Classification of ocean waves

Name	Period	Disturbing force	Restoring force
Capillary waves	< 0.1 s	Wind	Surface tension
Ultragravity waves	0.1–1 s	Wind	Gravity
Gravity waves	1–30 s	Wind	Gravity
Infragravity waves	0.5–5 min	Wind	Gravity
Long-period waves	0.1–12 h	Storms/ earthquakes	Gravity/ Coriolis
Tidal waves	12–25 h	Gravitation	Gravity/ Coriolis
Transtidal waves	> 1 day	Land-air-sea coupling	Gravity/ Coriolis

Wave Climate, Table 2 Beaufort wind scale

Beaufort number	Wind description	Wind speed (m s^{-1})	Wave height (m)	Wave description
0	Calm	0–0.2	0	Calm, mirrorlike
1	Light air	0.3–1.5	<0.05	Glassy, scalelike
2	Light breeze	1.6–3.3	<0.1	Rippled, no white caps
3	Gentle breeze	3.4–5.4	0.1–0.5	Wavelets, few whitecaps
4	Moderate breeze	5.5–7.9	0.6–1.2	Slight, numerous whitecaps
5	Fresh breeze	8.0–10.7	1.2–2.4	Moderate, some spray
6	Strong breeze	10.8–13.8	2.5–4.0	Rough, whitecaps everywhere
7	Near gale	13.9–17.1	4.1–6.0	Very rough, streaky foam forms
8	Gale	17.2–20.7	4.1–6.0	Very rough, marked foam streaks
9	Strong gale	20.8–24.4	4.1–6.0	Very rough, dense foam streaks
10	Storm	24.5–28.4	6.1–9.1	High, visibility reduced
11	Violent storm	28.5–32.6	9.2–13.7	Very high, foam everywhere
12	Hurricane	>32.7	>13.8	Phenomenal, sea completely white

supplemented by the classic “seaman’s” description of wind and sea. This scale is the basis of much of the data oceanographers have regarding winds and waves. It is at best qualitative and depends on the experience and training of the observer for its accuracy. Scientists who study wave climates over extended time need to appreciate the limitations of such data for trend analysis and climate change.

A quantitative theory for wind waves from first principles does not exist. The sea surface is rarely at rest and thus almost always presents an uneven surface for the wind to act upon. Most likely gustiness (turbulence) in air moving over still water has sufficient spatial inhomogeneity in speed to cause capillary waves. Once the sea surface is roughened, small pressure differences on the windward and leeward sides of wave crests transfer more energy from the air to the water, and the waves grow.

Discussion

Stress (τ) between two fluids or between fluid layers is a force per unit area. At the air-sea interface, it is classically given by $\tau = \rho_{\text{air}} C_D v^2$, where ρ_{air} is the density of air, C_D is the drag coefficient, and v^2 is the square of the wind velocity. Surface friction (F_r), a force per unit volume, is the vertical gradient of stress $F_r = \frac{\partial \tau}{\partial z}$ and is related to the square of the wind speed. Since wind velocity is a vector, stress and friction too are vector quantities. Interestingly, linear wave theory is approached by assuming that wave motion is frictionless.

Based on numerous observational experiments, oceanographers and ocean engineers have developed parametric relationships between wind speed (i.e., τ), fetch, and duration leading to quantitative values of wave height and wave period. The maximum wave height and period from a given wind speed, blowing over a minimum distance (fetch) and for a minimum time (duration), leads to the concept of a fully developed sea. Table 3, calculated from parameters in the

Wave Climate, Table 3 Characteristics of a fully developed sea

Wind speed ($\text{m} \cdot \text{s}^{-1}$) at 10 m	Fetch (kilometers)	Duration (hours)	Significant wave height (m)	Period (seconds)
5	60	10	0.6	4
10	235	20	2.5	8
15	530	30	6	12
20	940	41	10	17
25	1475	51	16	21
30	2125	61	22	25

Shore Protection Manual of 1984, summarizes typical values of fetch and duration for generation of the maximum deep-water significant wave height ($H_{1/3}$) and the longest period in a fully developed sea.

The higher the wind speed, the longer must be the duration and fetch for the sea to develop fully. Wave height and period in a fully developed sea will not increase if the minimum fetch and duration required are exceeded. If either the minimum fetch or duration required for a fully developed sea is not met, the resulting waves will have smaller heights and shorter periods. A comparison between Table 2 and Table 3 shows that the two approaches are not fully compatible, but there is general agreement between the seaman’s view of wave height and the ocean engineer’s.

Energy per unit area (E) in sea and swell is a function of the wave height squared (H^2) and the density of the water (ρ_{water}) and is given by $E = \frac{1}{8} \rho_{\text{water}} g H^2$; wave power per unit width of the wave front $P = E \cdot c_g$ is the product of wave energy and the wave’s group velocity (c_g). From linear wave theory, the *group velocity* of a shallow-water wave (a wave whose length L is 20 times or greater than the water depth Z) is $c_g = c = \sqrt{gZ}$ and for a deepwater wave (a wave whose length is less than one-fourth the water depth) is $c_g = \frac{1}{2} c = \frac{1}{2} \sqrt{\frac{Lg}{2\pi}}$. In these expressions, g is the acceleration of gravity, $9.8 \text{ m} \cdot \text{s}^{-2}$, and c is the phase speed or *celerity* of the wave. In the meter · kilogram · second (mks) system, expressions for wave energy

(E) and power (P) are joules/m² and watts/meter, respectively. Longer waves have more energy for a given height than shorter waves (i.e., wave energy per unit width = $\frac{1}{8}\rho g H^2 L$).

As waves transition from the deep sea into shallow water, their period remains the same, but their height increases. This can be seen by equating the shallow-water wave power in one water depth (Z_1) with that in another water depth (Z_2), and solving for the height (H) changes. The resulting equation, $\sqrt{\frac{Z_1}{Z_2}} = \frac{H_2}{H_1}$, quantifies the growth in the height of a wave as it enters shallower water ($H_1 > 0$). The underlying physical reason for this is conservation of energy and power in shoaling wave systems. Another consequence of wave period remaining constant and energy (per unit area) conservation is the steepening of deepwater waves if they encounter an ocean current flowing in opposition to the direction of the waves. Marine meteorologists often use the term “...and slightly higher in the Gulf Stream” to warn mariners of increased risk due to wave-current interaction.

Occasionally, unexpectedly very large waves are encountered; these are called *rogue waves* and have been known to impact coastal communities. On July 3, 1992, an hour before midnight, such a wave struck Daytona Beach, Florida. Many automobiles parked or driving on the hard sands were destroyed, and dozens of beachgoers were nearly drowned. A US Naval officer measured a 34-m-high rogue wave during a typhoon in the South Pacific Ocean. Such rogue waves have been observed more and more frequently as more ships and deep-sea oil platforms populate the oceans, and satellite radar systems mature.

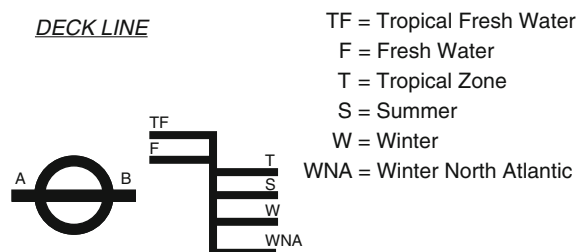
Risk to shipping and marine operations has been of great concern to sailors, marine transportation companies, and insurance firms for well over 100 years (*q.v.* Table 2). In the late nineteenth century, Samuel Plimsoll championed the use of load lines to guide mariners in safety (Maul 2017). An example of load lines (or Plimsoll marks as they are often called) is given in Fig. 1. The A–B designates the American Bureau of Shipping as the issuing agency. The other terms are marks that show how deep a ship may be loaded in different

environments. The marking “WNA” designates “Winter North Atlantic” and requires the vessel operator to allow the maximum freeboard (safety margin) for a ship operating in that environment. From well before Plimsoll’s time, the Atlantic Ocean in winter was known to have one of the most dangerous wave climates on Earth.

Surface wind speeds (u , v) over the ocean are to first approximation given by the geostrophic equations $\rho_{\text{air}} f v = -\frac{\partial p}{\partial x}$ and $\rho_{\text{air}} f u = -\frac{\partial p}{\partial y}$, where the Coriolis parameter $f = 1.4584 \cdot 10^{-4} \sin \varphi$; p is the sea-level air pressure; φ is the latitude; u, v are the east-west and north-south winds, respectively; and x, y is the related horizontal distance. The terms $\frac{\partial p}{\partial x}$ and $\frac{\partial p}{\partial y}$ are known as the pressure gradient and are a force per unit volume. Over the area of the Winter North Atlantic, ∂p is the sea-level pressure difference typically chosen as that at Iceland minus that at the Azores; ∂y is the horizontal north-south distance between Iceland and the Azores. The deep Icelandic Low minus the Bermuda-Azores High gives rise to the large pressure gradients that cause geostrophic winds to exceed $u = 25 \text{ m}\cdot\text{s}^{-1}$. Given that these winds often blow most of the winter and across thousands of kilometers of fetch, a glance at Table 3 will convince the reader that $H_{1/3}=16$ meter waves with >20 s periods is why WNA requires the greatest caution for the mariner operating in those seas.

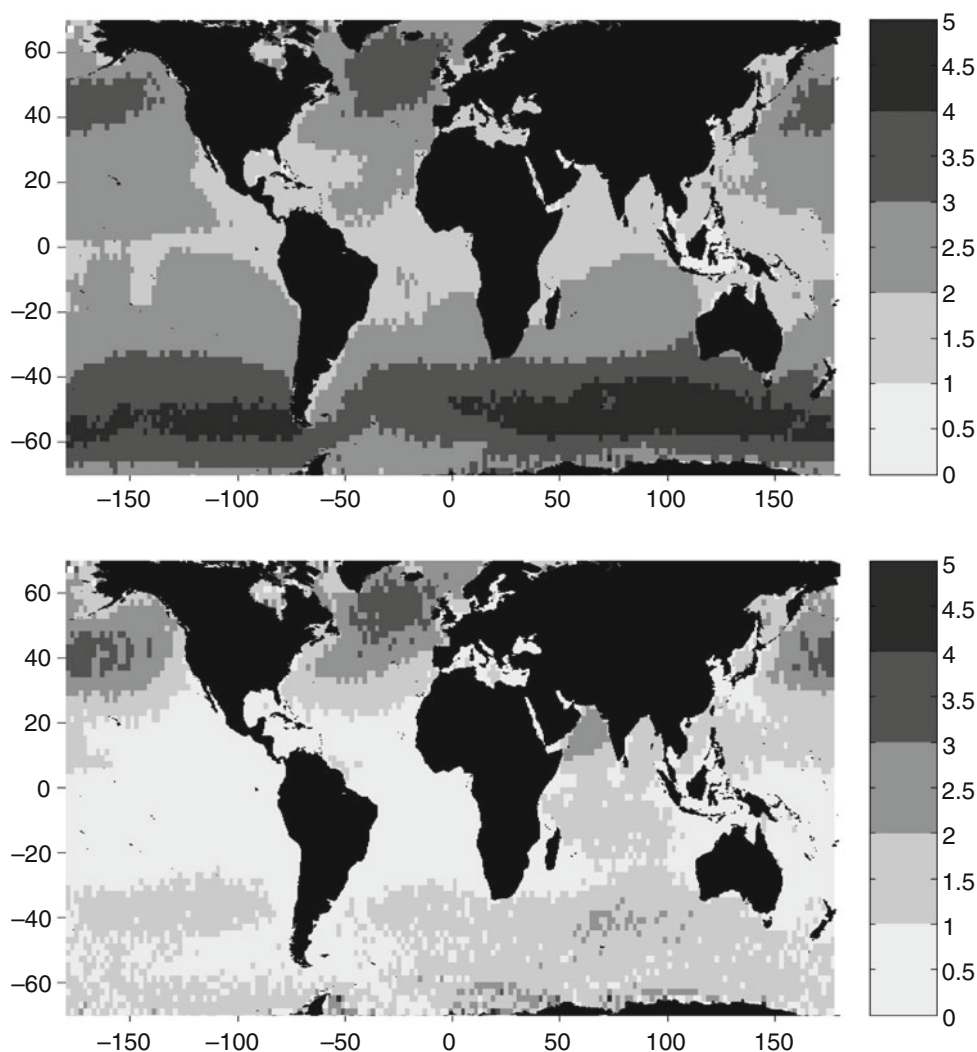
Until the late twentieth century, knowledge of conditions at sea, such as winds, waves, and currents, were from reports by the vessels plying those waters using the codes shown in Table 2. American navigators credit a US Naval Officer, Matthew Fountain Maury, with bringing together much of the information that underlies scientific understanding of winds, temperatures, sea, and swell. Those thousands of individual reports, collated and digested, have resulted in numerous atlases and pilot charts depicting average conditions to be expected. Since ships infrequently visit certain areas (such as the southeast Pacific Ocean), or when shipping routes change (such as after the opening of the Panama Canal in 1914), in certain areas, the data in these atlases and pilot charts are nonuniform, very sparse, or even nonexistent. Satellite oceanography has changed much of this lack of data homogeneity, both spatially and temporally.

Figure 2 was developed from the satellite altimeters flown during the last decades of the twentieth century. The upper panel depicts the average wave heights as measured from the altimeter return-pulse spreading due to surface roughness. While the Winter North Atlantic is obviously an area of large wave heights, it is the region of the Southern Ocean where Earth’s greatest waves are observed. Here, the north-south sea-level atmospheric pressure gradients $\frac{\partial p}{\partial y}$ are large, and the fetch is essentially infinite as there is only one significant land barrier: the Drake Passage between South America and Antarctica.



Wave Climate, Fig. 1 Load line markings for merchant vessels plying the oceans. A vessel loading in tropical freshwater may add cargo until the water is at the TF marking. Then as the ship proceeds out to sea, the greater buoyancy of salt water over freshwater will cause the ship to ride higher and add additional freeboard or margin of safety

Wave Climate, Fig. 2 Average wave height (upper panel) and variability about the average (lower panel) in meters. Dr. David Woolf processed these data at the Southampton Oceanography Centre in the United Kingdom from satellite altimeter data for the period 1985–1997 (Figure used by permission)



In low latitudes, Fig. 2 shows the lower sea states associated with the gentle Trade Winds. While the mean wave heights in the tropics and subtropics is on average small, these are the same regions whose weather is punctuated by tropical storms, hurricanes, typhoons, and cyclones. Sea states in these extreme storms are very high because the winds are oftentimes stronger than anything listed in Table 3, and also because the cyclonic air flow at the sea surface leads to a long fetch. In addition, such storms create a solitary wave known as the storm surge, which can easily raise sea level at least 5 m – on top of which is the sea and swell!

The lower panel in Fig. 2 depicts the variability of sea state about the mean shown in the upper panel. With the large seasonal wind changes in high northern latitudes come the largest changes in wave climate. Again the wisdom of Samuel Plimsoll’s “WNA” designation warns the mariner to take due precautions. In the Southern Ocean, the variability is much lower than in the North Pacific Ocean or the North Atlantic Ocean; the enormous ice mass of Antarctica provides the thermal inertia for fairly constant and large sea-level pressure

gradients $\frac{\partial p}{\partial y}$ and hence large east-west geostrophic winds and resulting waves $H_{1/3}$. The northwestern Indian Ocean (lower panel, Fig. 2) shows the large variability in sea states from the reversing monsoonal winds of Western India, with a smaller effect caused by the Southeast Asian Monsoon seen in the vicinity of the Philippines.

Conclusions

Earth’s wave climate is not static. Although the general picture presented herein is stochastically stable, the same satellite data that allows Fig. 2 to be compiled has shown decadal trends in certain areas. Notably the wave climate of the Norwegian Sea seems to be changing toward higher sea states. This change in wave height is correlated with a deepening of the Icelandic-Azores atmospheric sea-level pressure gradient – a variable known as the North Atlantic Oscillation. Whether this increase in $H_{1/3}$ is long term or cyclical is unknown, but it would be prudent to continue to monitor

sea and swell both from spacecraft and oceanographic observatories. Only through the concerted efforts of scientists such as Matthew Fountain Maury will oceanographers and ocean engineers be able to make statements with certainty about wave climate change.

The coastal science community is embarking on the design of a sustained integrated ocean observing system (IOOS). Conceptually, an IOOS must include measurements, modeling, and forecasts of numerous variables including wave height, direction, and period. Modern issues in coastal zone management require improved forecasts of natural hazards for Earth's burgeoning coastal populations, and wave energy will be of paramount importance not only for the immediate need in emergency management but in the design of future infrastructure. The "S" in IOOS will require effective integration of science, engineering, and management – an intellectual environment where wave climate is a dynamic, not a static, concept. Ocean waves are one component of the Earth system, and it will be from a systems approach that all components will be incorporated into effective decision-making.

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Wave Environments

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Wave environments are those areas of the seas and oceans that have similar wave sources and climatologies. The environments are closely related to wave generating systems, namely, the global and regional wind and cyclonic regimes, and secondly to the shape and orientation of the oceans and their surrounding coastlines.

Introduction

Waves are the most characteristic surface feature of the globe, with a presence over 71% of the surface. They are also the

most transitory feature as they continually form, travel, and ultimately, after a few minutes to days, dissipate their prodigious energy. As ephemeral as waves are, they are ever present, moving across the seas and great oceans and at the shoreline supplying on the order of 2.5×10^9 Kw energy each year (Inman and Brush 1973) to build, maintain, rearrange, and erode the coastal systems. This represents over half the energy in the coastal zone.

The seas, oceans, and water column play a passive role in the formation and movement of waves. While waves do move water, their source of energy and movement comes from external sources. For ocean or gravity waves, that source is wind, particularly the great zonal westerly and trade wind systems and cyclones in all their forms.

Wave Sources

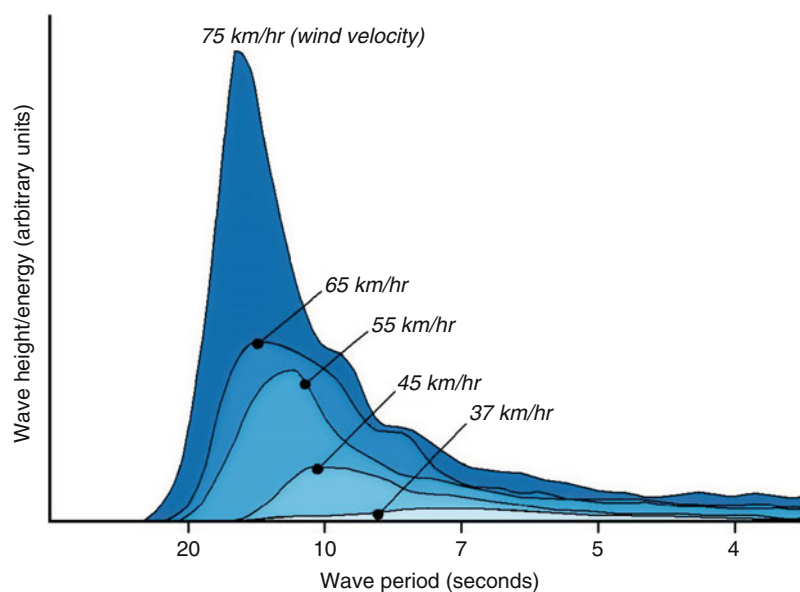
Ocean waves are entirely dependent on wind for their formation and on a global scale can be defined by the location of the zonal wind systems. In the high latitudes are the polar easterlies ($>70^\circ$ latitude), in the mid-latitudes the strong westerlies ($40\text{--}70^\circ$), and in the subtropics the most extensive, through moderate wind system, the trade winds ($30\text{--}10^\circ$), while straddling the equator are the quieter doldrums ($10^\circ\text{N}\text{--}10^\circ\text{S}$). These idealized wind systems and their ocean impacts are however substantially constrained and modified by continents and oceans, seasonal shifts in pressure systems, and in high latitudes sea ice. In addition, cyclones (tropical, east coast, and mid-latitude) while more limited in extent have the strongest winds and generate the world's biggest waves. At a regional scale, the monsoons and locally sea breeze wind systems all generate waves and have important impacts on their regional wave environments. Finally, the Coriolis effect plays a major role in redirecting waves as they travel across the oceans veering wave direction to the right in the northern hemisphere and left in the southern hemisphere. An overview of wave environments must therefore that all these factors into account.

Wave Generation

Waves are generated by the wind blowing over a stretch of water (estuary, lake, sea, or ocean). Wave height and period will both increase with increasing wind velocity and duration, together with length of fetch, the length of ocean over which the wind blows, and the uniformity of wind direction. Water must also be deep (>150 m); otherwise, the higher waves will shoal and break. The largest waves are therefore produced in those parts of the world's deep oceans where strong, unidirectional winds blow for long periods (days) over long stretches of ocean (1,000 km's). The single most important factor is wind speed, with wave energy increasing exponentially with increasing wind speed (Fig. 1).

Wave Environments, Fig. 1

The impact of increasing wind velocity on wave energy and period



Most of the world's wind systems – the polar easterlies, trades, and doldrums – are low to moderate velocity winds with wind speed rarely exceeding 12.5 m s^{-1} , and while they might blow for long periods, over long section of oceans, particularly the trade winds, they are not capable of producing high waves, with trade wind waves less than 3 m high 90% of the year and less than 2 m 50% of the year (Young and Holland 1996). In addition the polar easterlies blow over land in Antarctica and, for much of the year, over sea ice in the Arctic.

Wave Sources

The world's largest and most influential wave environments are those produced under the westerly wind stream between 40 and 70° latitude but focused on 50–60° N and S. In addition to the strong flow of westerly winds are the mid-latitude cyclones or subpolar lows. At any time, between 6 and 15 cyclones are embedded in this westerly stream, continuously encircling the globe at these latitudes. The cyclones form year-round over the Southern Ocean and more in the winter season over the North Pacific and North Atlantic oceans. The cyclones and their associated winds exceed 15 m s^{-1} 10% of the year and 10 m s^{-1} 50% of the year. They generate waves 2–3 m high 90% of the time across the Southern Ocean and in excess of 5–6 m 10% of the time in both hemispheres (Fig. 2; Young and Holland 1996).

Secondary cyclonic sources are the east coast cyclones or lows that form off the east coasts centered around 25–35° N and S. They occur in the USA (the north east storms), south east Australia (east coast lows), south Brazil, and South Africa. In Australia, they form on average ten times a year in all months with an early winter maximum and regularly generate waves in excess of 10 m, with wave recorded up to 19 m. Tropical cyclones (hurricanes, typhoons) are generated

in summer months between 5 and 15° N and S toward the warmer western side of the tropical oceans. While they can generate both high seas and associated storm surges in affected regions (10–30° N and S), their low and seasonal frequency (several per year in each region), and variable source area and erratic trajectory, means that they do not have a significant impact on longer-term wave climates.

Important regional winds include the reversing monsoons of India and Southeast Asia, northern Australia, and central Africa and the local sea breezes that can effect most of the world's coast, through more so in the mid- to lower latitudes.

All the above wind and cyclonic systems vary spatially, shifting with the seasons, and temporally in terms of occurrence and intensity. Some such as polar easterlies and trade winds are relatively uniform in direction and velocity and blow over large areas of oceans. All the cyclonic sources are more variable in their occurrence, trajectory, and wind direction and speed. Both the monsoons and tropical cyclones are seasonally dependent, and all like weather and climate have a high degree of variability, both in terms of location, trajectory, frequency, duration, and intensity.

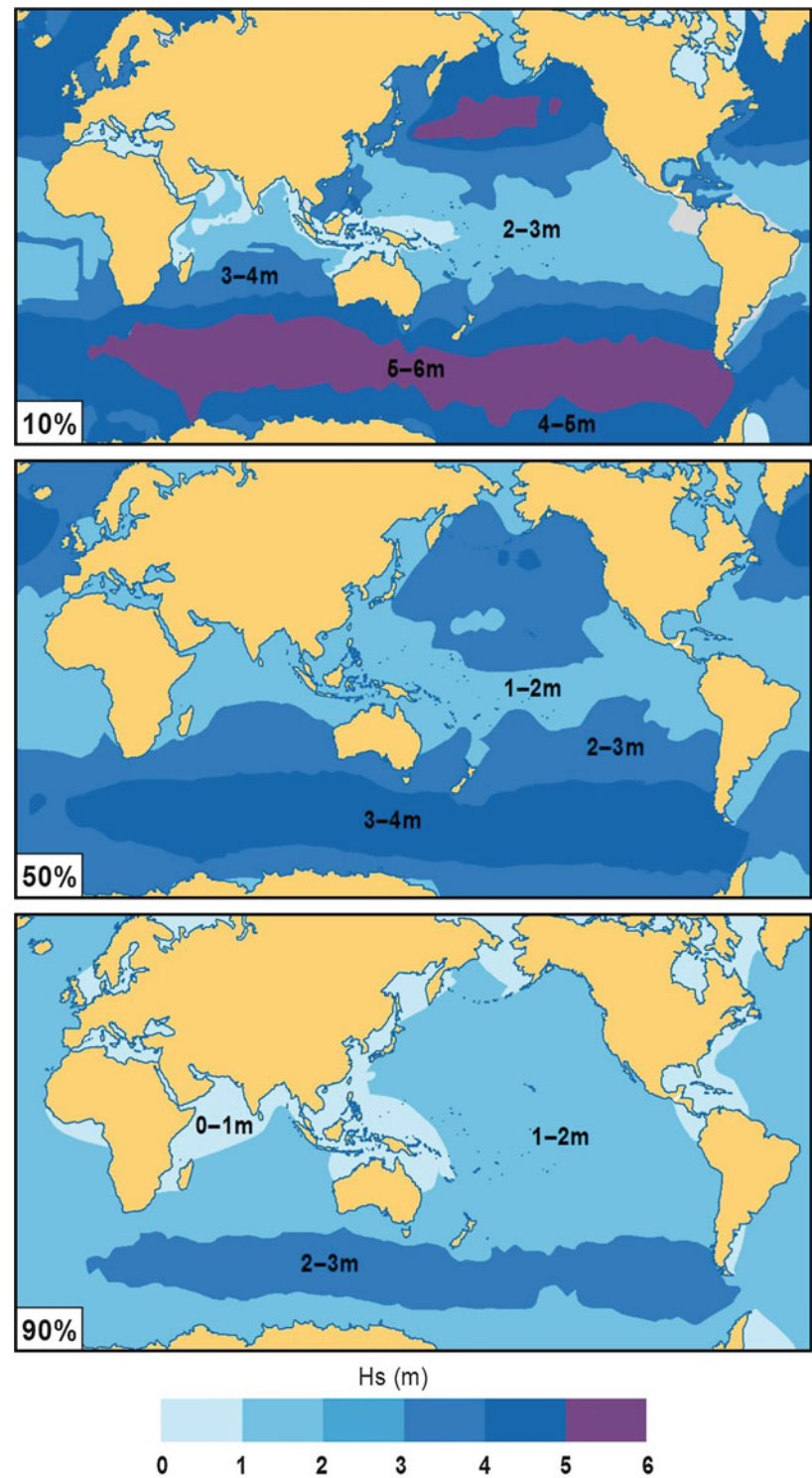
Global Wave Environments

Davies (1964, 1980) was the first to put some sense into this seemingly wide range of wave sources and types. He developed a morphogenetic classification of the world's wave environments, based on their wind sources, their latitudinal location, and the nature and direction of wave travel.

Davies identified four major deepwater wave environments, the high-energy storm wave of the upper mid-latitudes, the west coast swell, the east coast swell, and finally low-energy protected coasts of both polar and tropical regions (Table 1 and Fig. 2).

Wave Environments, Fig. 2

Global values for significant wave height which will be exceeded 10% (*top*), 50% (*middle*), and 90% (*lower*) of the time (From Short and Woodroffe 2009; based on Young and Holland 1996)

**Storm Wave**

The most important and most energetic wave environment is the storm wave environment located between 40 and 60° N and S, under the circumpolar west wind belt and its migratory mid-latitude cyclones. This is a region where all the ingredients for high waves are found. Winds exceed 15 m s^{-1} 10%, 10 m s^{-1} 50%, and 5 m s^{-1} 90% of the

time. While they rotate around the cyclones at any given latitude, they tend to dominate from a relatively uniform direction (south through west). In the northern hemisphere is the wide fetch of the North Pacific and North Atlantic oceans, while in the southern hemisphere is the world's only continuous stretch of ocean the circumpolar Southern Ocean, providing in principle unlimited fetch, all across

Wave Environments, Table 1 Characteristics of major world deepwater wave environments (Modified from Short 1999)

	Latitude/ location	Source – wave type	Seasonality	Deepwater height	Period	Direction
1. Storm wave	40–60° N and S	Mid-latitude cyclones – sea (roaring 40's raging 50's)	N hem – winter	High (2–5 m)	Long (10–14 s)	Westerly to south westerly
	S and W facing coasts		S hem – year-round			
2. West coast swell	40–0° N and S west facing coasts	Mid-latitude cyclones – swell	N hem – winter	Moderate- high (1.5–3 m)	Long (12–14 s)	N hem – NW
			S hem – year-round			S hem – SW
3. East coast swell + east coast cyclones + trade winds + monsoons + tropical cyclones	40–0° N and S east facing coasts	Mid-latitude cyclones – swell	N hem – winter	Moderate (1–2 m)	Moderate- long (8–12 s)	N hem – NE
			S hem – year-round			S hem – SE
	25–35° N and S	East coast cyclones – sea/swell	Year-round winter max	High (2–5 m)	Moderate- long (8–12 s)	N hem – E
	25–0° N and S exposed coasts	Trades – sea monsoons, sea tropical cyclones, sea	Trades in winter monsoons and tropical cyclones in summer	Moderate trades (0.5–1.5 m)	Trades – 6–9 s	Parallel winds
				Low monsoons (0.5–1 m)	Monsoons – 4–6 s	NE and SE trades
				TC – high	TC – short	NW and SW monsoons
4. Protected coasts	70–90° N and S pole facing coasts	Polar easterlies (+ sea ice)	Summer open water season	Low (0.5 m max)	Short (3–5 s)	NE
	Tropics 10–0° N and S	Doldrums	Year-round	Low (<0.5 m)	Short (<5 s)	Variable

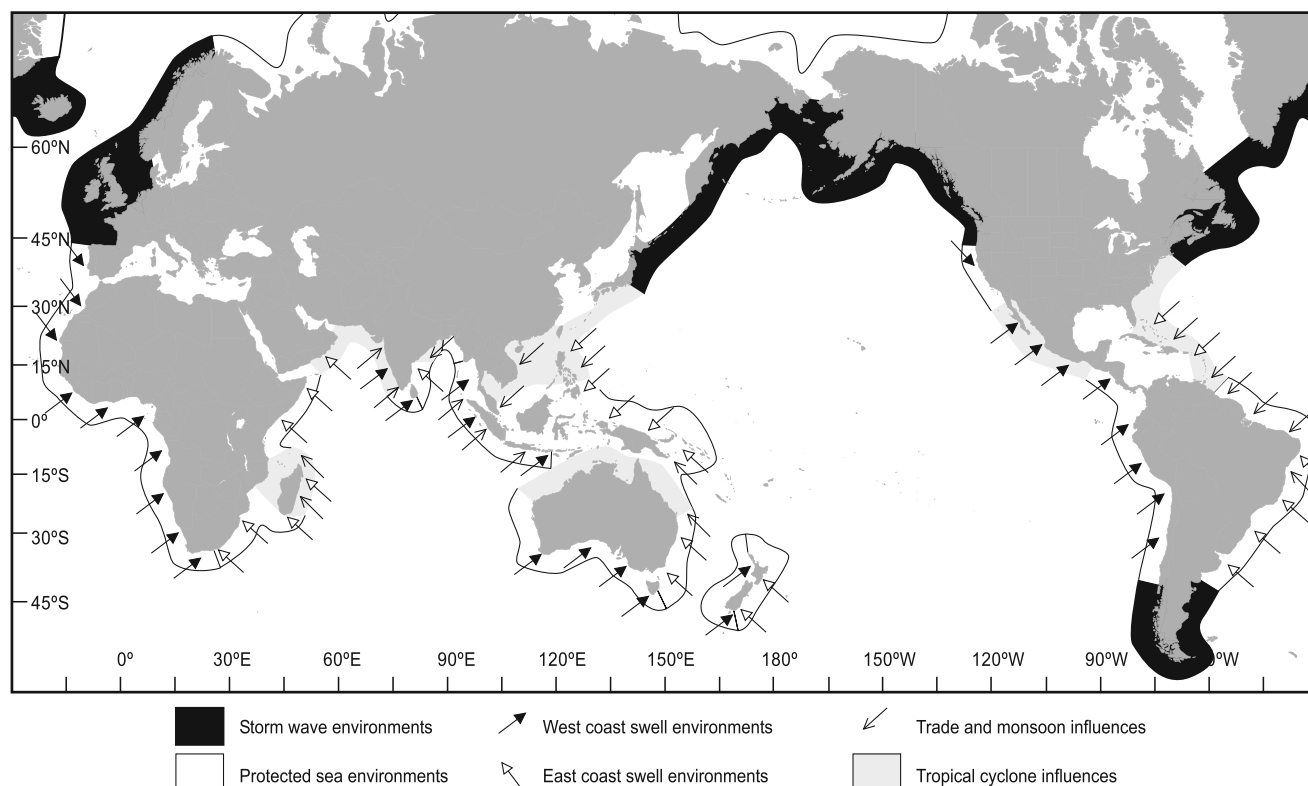
deep oceans with few impediments to wave generation or movement, such as islands or reefs.

Storm wave environments are exposed to persistent strong westerly winds, year-round in the southern hemisphere and in the winter season in the northern hemisphere. They produce the world's highest waves averaging over 2–3 m 90% of the year and 5–6 m 10% of the year (Young and Holland 1996), with extreme waves reaching several meters. Periods are 10–14 s; however being a *sea* environment in the area of generation, waves are high, short, steep, and highly variable in size, shape, and direction (Table 1). Sailors often use the term confused seas when referring to such conditions, with white caps and “rouge” or “freak” breaking wave characteristic of such environments. Wave direction is predominately to the west, with the Coriolis effect deflecting waves equatorward in both hemispheres.

West Coast Swell

West coast swell refers to the long persistent swell that reaches the west coast of many of the world's mid- to low-

latitude coasts, including the west coast of the Americas, Africa, Australia, India and southern Indonesia, New Zealand and many mi-ocean islands. The *swell*, unlike the storm seas, refers to waves that have left the area of wave generation, the storm wave environment, or the wind has stopped blowing. When this happens, the sea waves quickly reformed into lower, longer, faster more uniform swell waves, more uniform in size, shape, and direction of travel. Once transformed into swell, they are able to travel across deep oceans for thousands of kilometers with minimal loss of energy (Fig. 3). The waves are however affected by the Coriolis effect, which turns them equatorward in both hemispheres causing the initially westerly waves to arrive from the northwest in the northern hemisphere and southwest in the southern hemisphere. The swell is higher in the higher latitudes, slowly decreasing toward the equator. They are characterized by moderate to high (2–3 m 50%), long period (12 s+), and uniform swell, with periods of higher swell associated with major

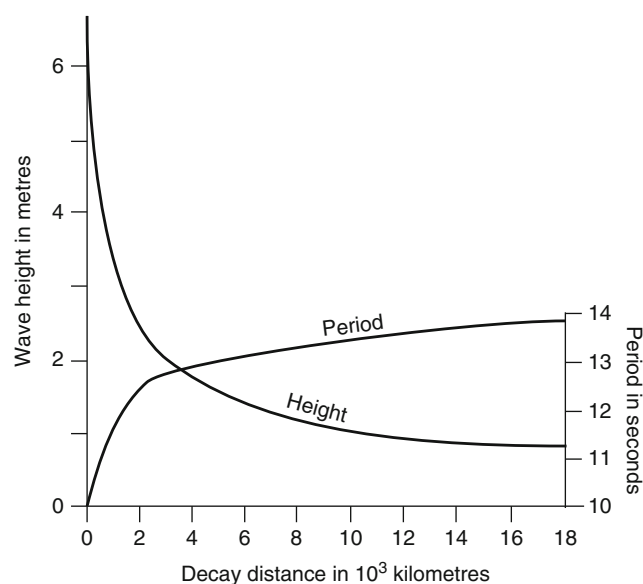


Wave Environments, Fig. 3 The world's major wave environments (Adapted from Davies 1980). Note storm wave and west coast swell environments operate year-round in southern hemisphere and during winter season in northern hemisphere

cyclones and lower swell in between, which in the southern hemisphere arrives on average 350 days a year. It is however distinctly seasonal in the northern hemisphere (Table 1) where it delivers the big winter surf to Hawaii and the California coast. The higher winter swell and lower summer swell along the California coast gave rise to the winter cut-summer fill beach cycle, which, while applicable to California, has limited application in other wave environments (Fig. 4).

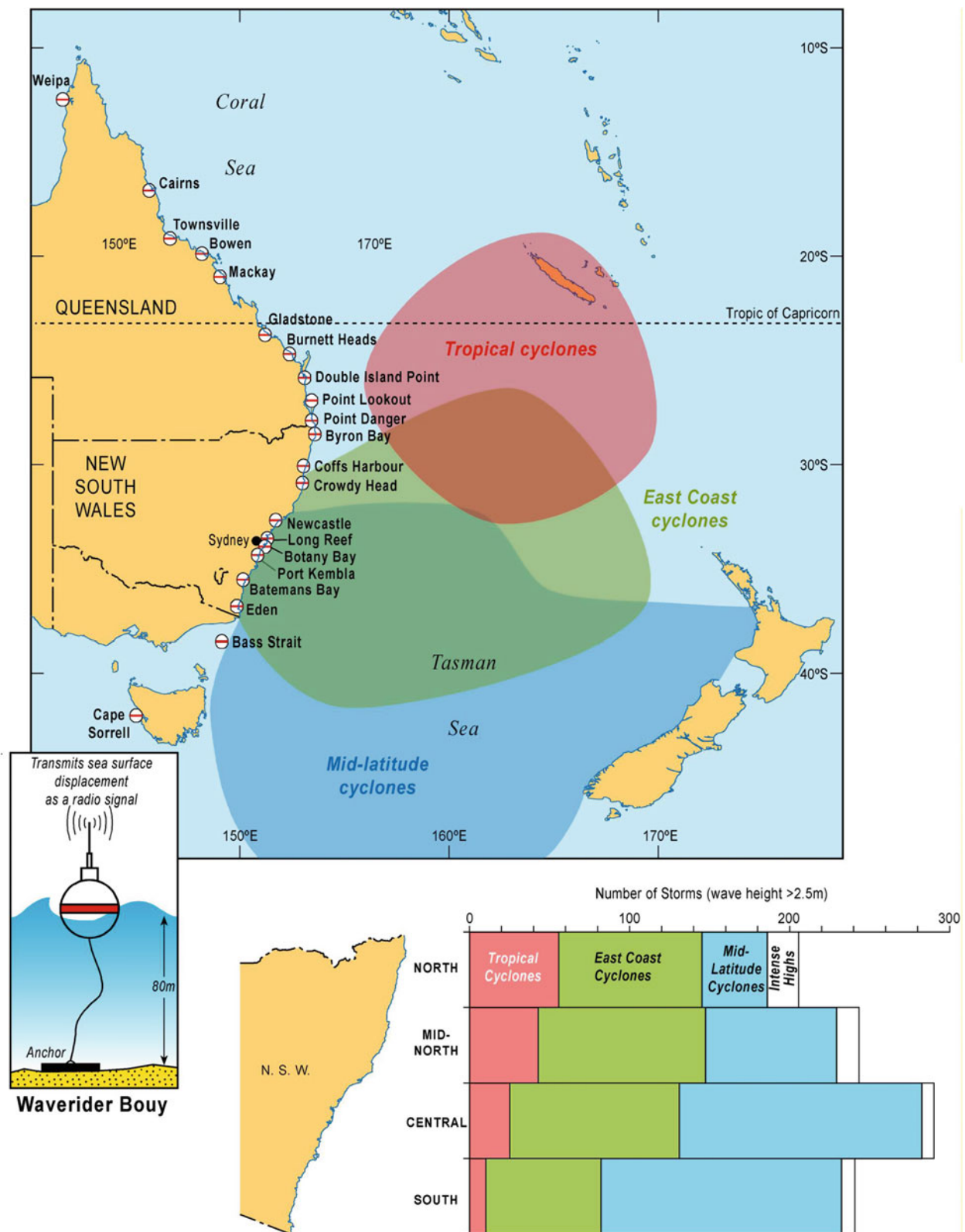
East Coast Swell

East coast swell refers to swell generated in the storm wave environment that travels equatorward but has to undergo more refraction to reach the east coast of the continents below 40° N and S. Because of the initial trajectory and greater refraction required to reach the coast, the swell is lower and arrives less frequently. On the southeast coast of Australia, it arrives 200 days, as opposed to 350 days on the south coast, and average 1.5 m down from 2.5 m on the south coast. It still provides a relatively persistent (60%) low to moderate swell. In addition, the mid- to low-latitude east coasts are exposed to addition wave regimes, which produce the world's most complex wave environment (Short and Trenaman 1992). These include periodic east coast cyclones centered on 40–25°, persistent trade winds peaking between 20 and 5°, summer tropical cyclones-hurricanes originating



Wave Environments, Fig. 4 Transformation of sea (left) to swell (From Davies 1980)

between 15 and 5°, and summer sea breezes. All these wind regimes produce largely sea waves, ranging from lower frequency but high wave conditions associated with the cyclones to moderate short waves with the trade winds and



Wave Environments, Fig. 5 Source of waves arriving at Sydney on the southeast Australian coast, together with location of waverider buoys (top), and latitudinal shift in storm waves arriving along the NSW coast (lower) (Source Short and Woodroffe 2009)

to low, short waves with the sea breeze. In addition all these waves can be generated, while swell arrives from the storm wave environment, resulting at times two wave trains arriving from different sources and directions. East coast swell environments also undergo an equatorward transition from predominately easterly swell conditions in the higher latitudes (40°) to increasing influence from first east coast cyclones ($40\text{--}25^\circ$) and strong trade wind influence between 20 and 0° , together with periodic tropical cyclones between 30 and 10° . Figure 5 illustrates the range of wave sources and their latitudinal shift in impact along the southeast Australian coast.

Protected Wave Environments

Protected wave environments refer to ocean and sea environments protected from ocean swell. They occur in both the tropics where they are protected by location in the doldrums, by distance from major wave sources, and physically by coral reefs, island archipelagos, and smaller seas, as well as polar locations protected by sea ice from the low-velocity polar easterlies. Around the Antarctic continent, the Coriolis effect send most of the westerly tending storm waves equatorward, ensuring further protection of the polar continent. All protected locations are devoid of swell, relying on local winds for generation of short, low seas, interspersed with long calms.

Wave Environments, Fig. 6

Waves arriving at Makapuu Beach, Hawaii, undergo shoaling, leading to an increase in wave height (*right*) and then breaking (*left*) as they reach water depth ~ 1.5 times the wave height (A.D. Short)



Other Influences

Sea ice can form over the ocean surface as around Antarctica and across the Arctic Ocean and seasonally along the shores of high latitude coast such as Labrador and the Bering Sea. The ice both prevents the formation of waves and at the shore either attenuates and/or prevents waves reaching the shore.

Islands and Reefs

Islands and reefs wherever they occur in the mid-ocean or fringing the shoreline will cause wave attenuation and refraction, leading to a reduction in wave height and changes in wave direction.

Breaker Wave Environment

All the above refers to deepwater sea and ocean wave environments. As waves move from deep to shallow water surrounding the continents and shorelines, they undergo a predictable transformation from deep to shallow water waves. In doing so, they may undergo wave attenuation as they shoal across the continental shelf and nearshore zone, wave refraction over variable submarine topography, wave diffraction in bays and ultimately as they enter shallow water they slow, shorten, and ultimately break (Fig. 6). The height of the breaker wave is therefore not only a function of the deepwater wave environment, which will determine the maximum possible wave height, but also the shoaling processes, which can reduce wave height from between 0% and 100%.

Summary

Ocean waves are the most common feature on the earth's surface and contribute most of the energy to do work at the coast. Transitory as they are, they can travel thousands of kilometers to deliver energy to the shore. Waves are generated by winds blowing over the ocean surface – the stronger the wind and the longer its duration, the higher, longer, and more energetic the waves. The major global wind systems each produce a characteristic set of waves – from the low short summer waves of the polar easterlies, the high long storm waves, and the swell of the mid-latitude cyclone belt to the great expanse of moderate trade wind waves. The highest waves are generated by cyclones – mid-latitude, east coast, and tropical – with the mid-latitude cyclone responsible for the high-energy storm wave and the west coast swell coastal wave environments, while they also contribute to east coast swell environments, where they combine with east coast cyclones and in lower latitude trade wind waves. While there global wave environments represent the deepwater wave climate, the breaker waves climate will depend on its deepwater climate, combined with the orientation and topography of the coast and shelf which will determine the amount of wave attenuation, refraction, and diffraction and ultimately the breaker wave height at the shore, which can range from zero to 100% of the deepwater wave height.

Cross-References

- [Accretion and Erosion Waves on Beaches](#)
- [Bay Beaches](#)
- [Boulder Beaches](#)
- [Carbonate Beaches](#)
- [Dissipative Beaches](#)
- [Dynamic Equilibrium of Beaches](#)
- [Gravel Beaches](#)
- [Ice-Bordered Coasts](#)
- [Monitoring Coastal Geomorphology](#)
- [Rating Beaches](#)
- [Reflective Beaches](#)
- [Tourist Beaches](#)
- [Waves](#)
- [Wave Climate](#)
- [Wave-Dominated Coasts](#)
- [Wave Hindcasting](#)

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Wave Focusing

Terry R. Healy

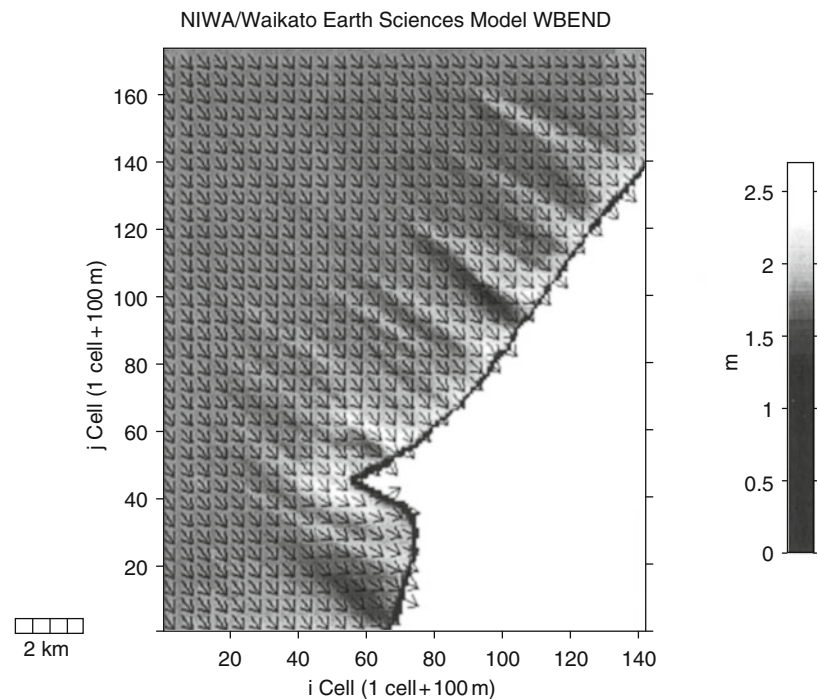
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Swell waves traversing the inner shelf and shoreface undergo shoaling and refraction processes around seafloor topographic highs, such as offshore reefs or transverse sand ridges. These seafloor topographic irregularities act as a “bathymetric lens” (Speranski and Calliari 2001) inducing refraction processes such that the wave orthogonals become compressed in the lee (shoreward) of the seafloor topographic high, that is, energy along the wave front becomes concentrated, resulting in a localized greater wave heights (Healy 1987; Calliari et al. 1998). Away from the zones of wave energy concentration the refraction may spread the wave orthogonals, dispersing the wave energy, and resulting in lower wave heights. As a result of this process, localized zones of higher wave energy may be concentrated or *focused* at the coastline (Finkl and Bruun 1998). This feature of enhanced wave heights is well-known at headlands, but is not often recognized along sandy beach coastlines (Fig. 1). Relatively short swell waves ($T \sim 7$ s) undergo little refraction when crossing the shoreface, but longer period waves ($T \geq 10$ s) may undergo marked refraction and focusing in water depths less than 50–60 m.

Some remarkable erosion and geomorphic effects may result from wave energy focusing along a sandy coastline (Finkl and Bruun 1998). Sectors of beach subject to wave focusing exhibit localized higher breaking waves, resulting in enhanced wave set-up and run up, and thus are the zones where accelerated dune erosion or dune overwash occurs during episodic storm events. Local variation in littoral drift gradients result. Healy (1987) demonstrated from several examples of numerical wave refraction simulation that wave energy focusing exacerbated rates of erosion along sectors of frontal dune on generally long straight beaches. These sectors of “accelerated” erosion created “arcuate duneline embayments”

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Wave Focusing, Fig. 1 Wave refraction simulation commencing with deepwater wave height $H_o = 1.63$ m, and period $T = 10.64$ s. Note the general focusing of energy around the headland where the light patches show wave heights reach 2.5 m, and at various focused sectors further along the coast where waves reach 2.5 m at breaking. In between the major wave-energy focus zones are sectors of lower wave heights. (Source H. Easton, personal communication)



of order 200–500 m long and 10–20 m amplitude which are cut into the dune face (Stephens et al. 1999). Similar effects have been reported along the coast of Brazil (Speranski and Calliari 2001). Wave focusing has also been demonstrated as a fundamental cause of barrier spit breaching and breakthrough during a storm, and was identified as a factor in coastal hazard zone assessment (Healy 1987).

On the inner shelf or shoreface seafloor, the zones of focused wave energy (higher waves) are subjected to greater bottom orbital velocities, and therefore potentially greater scour. These are the zones where sidescan sonar surveys demonstrate that large-scale ripples (often termed “*megaripples*” “*coarse grained ripples*”) are formed in zones of otherwise fine sands (Healy et al. 1991; Bradshaw et al. 1994).

Cross-References

- [Beach Erosion](#)
- [Littoral Drift Gradient](#)
- [Wave Refraction Diagrams](#)
- [Waves](#)

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Wave Hindcasting

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Definition

Wave hindcasting refers to predicting surface waves for a past wind event. The same process is described in different terms – as wave nowcasting and wave forecasting referring to the real-time and future wind-event predictions, respectively. The hindcasting wave parameters of interest are wave height (H , the height from the bottom trough to the top crest) and period (T , simply the passing time of two successive crests).

These two parameters are often called integral parameters – because the individual characteristics such as the shape and the wave regularity or irregularity are not revealed by these two parameters and the hindcasting methods. The wind waves are highly spectromatic – therefore they can best be described in spectral energy terms. The integration of the spectral energy yields a characteristic wave height of the spectrum known as the H_{mo} which turns out to be nearly equivalent to the significant wave height, H_s or $H_{1/3}$ – the average of the highest one-third of the spectrum. On most occasions, H_s is slightly higher than H_{mo} , but they are equivalent in deepwater waves (Goda 1974; Thompson and Vincent 1985; Sorensen 1993). The frequency of the spectrum representing the peak energy is known as the peak frequency, or reciprocally the peak wave period, T_p .

A wind wave is generated by the complex interactions between the blowing wind and the water surface – by the net transfer of energy from the former to the latter. There are two wind wave generation theories pioneered by Phillips (1957) and Miles (1957) – both of them are based on the arguments of resonance phenomenon. The model proposed by Phillips (1957, 1960) relies on the resonant interaction between the forward moving wind turbulent pressure fluctuations and the generated free waves propagating at the same speed of the wind pressure. The pressure fluctuates with the varying wind magnitude and frequency and is generated by turbulent eddies in the wind field. The Phillips model gives rise to a linear growth of the wave spectrum in time, and best explains the initiation and beginning stages of wave generation. A shear flow model proposed by Miles (1957) provides for a useful mechanism of momentum transfer from the wind to the water surface. An elaborate review of the wave generation processes could be found in Janssen (1994) and Holthuijzen (2007).

Basic Hindcasting Parameters

The hindcasting methods rely on three storm-related parameters – the wind speed (U), the storm duration (t), and the fetch (F). For an accurate hindcasting, each of these parameters needs careful definition (USACE 2006). There are five different adjustments needed to define the wind speed of interest from raw measurements: (1) adjusting the length of averaging time, (2) adjusting the measurements to the height of interest, (3) defining the storm duration, (4) adjusting the wind speed to account for the variability in wind direction, and (5) adjusting the wind speed to represent a certain return period or probability of interest using a suitable extremal distribution method. Wind observation stations are usually located in areas of open airflow that is not obstructed by structures or topography. If there are obstructions, additional adjustments

are necessary (Woodward and Pitablo 2010). Measured wind speed time series usually represents a certain sampling frequency or interrogation time and a data-storage time interval. These two timings vary among the instruments and the applied configurations. For an accurate hindcasting, the data must be adjusted to represent the size of the area of interest; this means the requirement for applying a longer averaging time for a larger fetch than a smaller one. Sometimes wind speed is reported as the fastest mile wind speed – this speed represents the applied averaging time equivalent to the time required by wind to travel a distance of 1 mile. Wind speeds are described in different terms such as the gusts (3-s) and the sustained or mean representing an average, typically over a timescale of 1–15 min. At sustained wind speeds, any change of the averaging time has negligible effect on its magnitude.

Measurements of wind speed and direction time series are usually conducted at different heights above surface – for example, close to the sea surface by weather buoys and at different heights but mostly at about 10 m above surface in ground stations. These speeds measured at different heights must be adjusted to represent a certain height. This is necessary because a boundary layer develops near the surface where wind speed is lower than the open-air speed. The height of the atmospheric boundary layer depends on the surface roughness and wind speed and is often modeled as a logarithmic change from near-to-zero at the surface to the asymptote of open-air speed. Most wave hindcasting methods are based on wind speeds representative at 10 m above surface, except the PM method (Pierson and Moskowitz 1964) which uses a height of 19.5 m. The height adjustment can be made either using a logarithmic relation or simply a power law profile. As Eq. 1 shows, the latter known as the 1/7th power law (USACE 2006) can be used to estimate the speed at 10 m (U_{10}) from the wind speed (U_z) measured at any height z above the ground:

$$U_{10} = U_z \left(\frac{10}{z} \right)^{1/7} \quad (1)$$

The definition of the storm duration and adjusting the wind speed to wind-direction variability are related to the definition of fetch. The fetch is defined as the water distance along the wind direction from the shore to the point of interest. In addition to the fetch, another water-related parameter, the definition of depth becomes necessary for shallow-water wave hindcasting. The relation between the fetch and the storm duration (USACE 2006) can best be understood from Eq. 2:

$$\frac{gt_d}{U_A} = 68.8 \left(\frac{gF}{U_A^2} \right)^{2/3} \quad (2)$$

where g is acceleration due to gravity, t_d is the limiting or threshold storm duration, F is fetch, and U_A is the adjusted wind speed, given by

$$U_A = 0.71U_{10}^{1.23} \quad (3)$$

A fetch-limited wave is generated when the storm duration $t \geq t_d$. When $t < t_d$, the wave generation becomes duration limited. The accepted cut-off limit for defining a storm duration is the wind speed variability up to 2.5 m/s (or 5 knot). The directional wind speeds of interest are those spread to 45° about the fetch line – the straight line from the shore to the point of hindcasting (USACE 2006). This wind spreading limit is appropriate in lakes and constricted water bodies where an effective or equivalent fetch is necessary to be defined.

For most practical applications, a storm event of interest should be categorized to belong to a certain recurrence interval or return period. For this purpose, it is necessary to fit the maximum monthly or preferably the annual maximum wind speeds to a long-term distribution. Wind data mostly fit to the Fisher-Tippett I (or Gumbel), the Fisher-Tippett II, and the Weibull distributions (USACE 2002).

In addition to the above adjustments, additional adjustments become necessary because wind speeds measured over water register higher magnitude than the overland measurements. This happens due to the smoothness of sea surface and the consequent development of a thin boundary layer. This and all the necessary adjustments are reviewed by USACE (2002, 2006).

A typical ocean wave climate during a windstorm comprises both wind waves and swells. Wind waves or seas are distinct from swells. Swells are smooth undulations representing decayed and dispersed wind waves that have traveled out of their generating area and are no longer subject to wind input.

In the following sections, three methods of wave hindcasting, empirical relations, spectral models, and numerical models, are discussed with a brief history of the development of wave hindcasting methods. Unless specified otherwise, all units in this entry refer to SI.

Wave Hindcasting Methods

History

Probably the earliest wave hindcasting method started with the pioneering works of Admiral Sir Francis Beaufort of the British Navy in 1805. His devised scale popularly known as Beaufort wind scale provides a qualitative description of winds and visible wave features. The modern history of the quantitative methods of wave hindcasting dates back to World War II. To forecast conditions during the landings with

amphibious vehicles, Sverdrup and Munk (1947) developed methods that related wind speeds to the higher and more distinct oscillations on the sea surface. These oscillations in space and time were characterized using the terms significant wave height (H_s) and significant wave period (T_s), respectively. More works followed with the compilations of wave data, statistical analyses, and derivation of refined relations (Longuet-Higgins 1952; Darbyshire 1952; Bretschneider 1952). It was Neuman (1953) who derived the wave spectral relations relating significant wave heights and periods. This was followed by the works of Pierson et al. (1955) that used Neuman spectra to derive the graphical methods for engineering applications. Further developments are covered in the following sections, and interested readers could consult Sylvester (1974) for more on the historical developments.

Empirical Relations

Using a simple wave energy growth concept and an early prediction method developed by Sverdrup and Munk (1947), Bretschneider (1952, 1958) refined the hindcasting method by an improved calibration with a large amount of field data. It is known as the SMB (Sverdrup-Munk-Bretschneider) method. This method allows graphical determination of deep-water (depth/wave length < 0.5) wave parameters, H_s and T_s using the wind speed, fetch, and duration. In functional relationship, the method is based on a relationship between the dimensionless significant wave heights and periods with the dimensionless fetch and windstorm duration:

$$\frac{gH_s}{U^2} \text{ and } \frac{gT_s}{2\pi U} = f\left(\frac{gF}{U^2}, \frac{gt}{U}\right) \quad (4)$$

In deep water, the celerity of wave is $C = gT/2\pi$, and the ratio C/U is known as wave age – an important parameter defining the wave growth. The deepwater graphical relations were later presented as equations by Bretschneider (1970).

For a circular wind field such as a hurricane, Bretschneider (1957) proposed relations based on an analysis of 13 hurricanes in the Atlantic Ocean off the US East Coast. The relations are given as follows (USACE 2006):

$$H_{sp} = 16.5e^{0.01RAP} \left[1 + \frac{208\alpha V_F}{\sqrt{U_R}} \right], \quad (5)$$

and

$$T_{sp} = 8.6e^{0.005RAP} \left[1 + \frac{104\alpha V_F}{\sqrt{U_R}} \right], \quad (6)$$

where H_{sp} (feet) and T_{sp} (seconds) are peak significant deep-water wave heights and periods, respectively, R is distance (nautical miles) from the center out to the point of maximum wind velocity, ΔP is pressure difference (inches of mercury)

from the center to the periphery of the hurricane, V_F is forward speed (knot) of the hurricane, U_R is maximum wind speed (knots) at R , and α is a correction factor based on the hurricane speed and is equal to 1 for a slow-moving hurricane. The calculated peak H_{sp} and T_{sp} develop in the vicinity of the point of maximum wind velocity.

Spectral Models

Based on empirical fits to measured waves, three well-known one-dimensional deepwater spectral models (Bretschneider, PM, and JONSWAP spectra) have been proposed to hindcast waves. The term one dimensional refers to wave growth in the direction of the wind without indicating the directional distribution of the generated waves. The first is the Bretschneider spectrum (Bretschneider 1959) that provides a period spectrum as a function of wind speed, fetch, and wind duration. The second is the Pierson and Moskowitz (1964) wave frequency spectrum based on the analysis of wave and wind records from British weather ships operating in the north Atlantic. Their analyses considered fully arisen sea for wind speeds (estimated at an elevation of 19.5 m). This wave frequency spectrum known as Pierson-Moskowitz (PM) spectrum, representing only the fully arisen sea, provides wave frequency as a function of wind speed.

The third is the JONSWAP (Joint North Sea Wave Project) spectrum (Hasselmann et al. 1973) developed from wave and wind measurements with sufficient wind duration and provides a fetch-limited spectrum. This spectrum gives a relation between wave frequency and wind speed and fetch. The Shore Protection Manual (USACE 2006) recommended a parametric method based on JONSWAP spectrum for deepwater wave prediction and replaced the earlier recommended SMB method. According to this method, the following relations are used for both fetch and duration-limited conditions:

$$\frac{gH_{mo}}{U_A^2} = 0.0016 \sqrt{\frac{gF}{U_A^2}}, \quad (7)$$

$$\frac{gT_p}{U_A} = 0.286 \left(\frac{gF}{U_A^2} \right)^{1/3}, \quad (8)$$

The significant wave period T_s can be estimated from T_p as $T_s = 0.95T_p$. If the actual wind duration (t) is greater than the calculated limiting wind duration (t_d), a fetch-limited case applies. A duration-limited case applies if the converse is true, and an effective fetch is computed using Eq. 2 by replacing t_d with the actual duration of the storm, t . With the known effective fetch, Eqs. 7 and 8 are used to estimate H_{mo} and T_p .

For a fully arisen sea when $gt/U_A \geq 71,000$ and $gF/U_A^2 \geq 22,800$, the deepwater wave parameters are estimated as

Wave Hindcasting, Table 1 Fetch-limited hindcasted significant wave heights (H_s) and peak spectral wave periods (T_p) for three different Beaufort (BF) scale wind speeds acting on a 20-km-long fetch

Wind	SMB (H_s and T_p)	JONSWAP (H_s and T_p)	Wilson- Goda (H_s and T_p)
Strong Breeze BF 6; 25 knot	1.1 m; 4.3 s	1.1 m; 4.2 s	1.0 m; 3.8 s
Strong Gale BF 9; 44 knot	2.2 m; 6.0 s	2.4 m; 5.3 s	2.1 m; 5.0 s
Storm BF 11; 61 knot	3.2 m; 7.1 s	3.5 m; 6.1 s	3.0 m; 5.8 s

$$\frac{gH_{mo}}{U_A^2} = 0.243, \quad (9)$$

and

$$\frac{gT_p}{U_A} = 8.13. \quad (10)$$

In addition to the above three spectral methods, other methods followed, notable among them is the Goda-Wilson method (Goda 2003; Wilson 1959). Table 1 shows the examples of the hindcasted significant wave heights and peak spectral periods for three different wind speeds acting on a 20-km-long fetch. The SMB, JONSWAP, and Goda-Wilson methods are applied for the purpose. For the fetch-limited conditions of the three cases, the methods predict wave parameters in similar orders of magnitudes. The SMB method predicts somewhat lower H_s , but higher T_p than others. Figure 1 shows the JONSWAP spectrum for the three wind conditions.

JONSWAP spectrum can also be used to estimate wave parameters resulting from a circular wind field such as a hurricane using a method developed by Young (1988). An equivalent hurricane fetch is first estimated using the hurricane forward velocity and maximum wind velocity. This is then used to estimate wave parameters using the above equations.

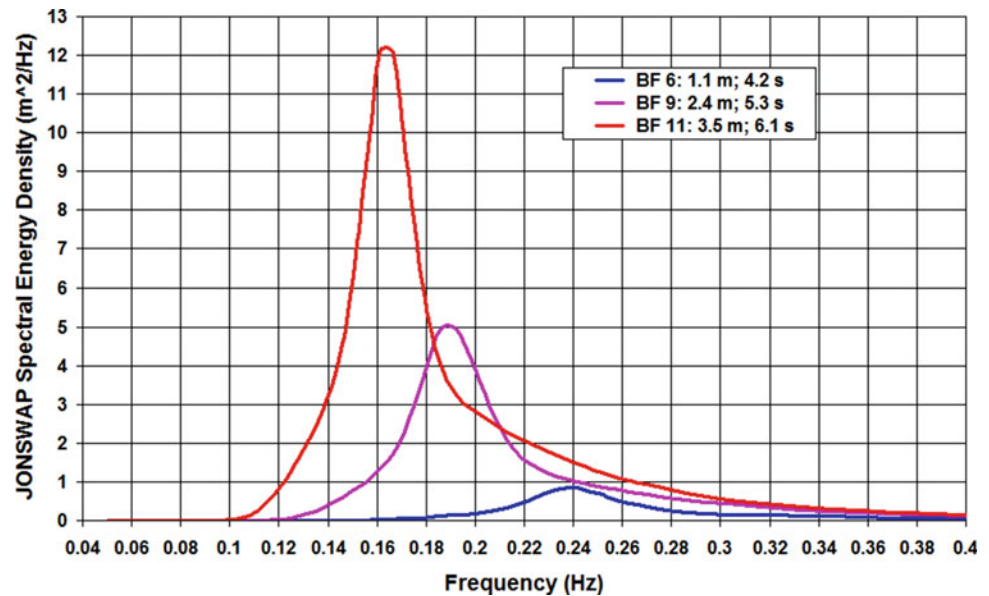
For shallow-water (depth/wavelength > 0.5) applications, USACE (2006) recommended the use of the following equations pertaining to the JONSWAP spectrum:

$$\frac{gH_{mo}}{U_A^2} = 0.283 \tanh \left[0.53 \left(\frac{gd}{U_A^2} \right)^{3/4} \right] \tanh \left[\frac{0.00565 \left(\frac{gF}{U_A^2} \right)^{1/2}}{\tanh \left\{ 0.53 \left(\frac{gd}{U_A^2} \right)^{3/8} \right\}} \right], \quad (11)$$

$$\frac{gT_s}{U_A} = 7.54 \tanh \left[0.833 \left(\frac{gd}{U_A^2} \right)^{3/8} \right] \tanh \left[\frac{0.0379 \left(\frac{gF}{U_A^2} \right)^{1/2}}{\tanh \left\{ 0.833 \left(\frac{gd}{U_A^2} \right)^{3/8} \right\}} \right]. \quad (12)$$

Wave Hindcasting,

Fig. 1 Hindcasted fetch-limited JONSWAP wave spectra with the significant wave heights and peak spectral periods for three different Beaufort (BF) scale wind speeds acting on a 20-km-long fetch



$$\frac{gt_d}{U_A} = 537 \left(\frac{gT_s}{U_A} \right)^{7/3}, \quad (13)$$

where d is water depth. However, shallow-water wave parameters can best be estimated from deepwater waves using wave transformation techniques.

Numerical Models

The above wave hindcasting methods do not give directional distribution of the wave field. The directional spectra are basically one-dimensional spectra corrected by a dimensionless directional spreading function that depends on the wave frequency and direction. The directional spectra can best be hindcasted by numerical models to forecast, nowcast, and hindcast wave climates. The models are based on the numerical integration of the spectral energy balance equation over a spatial grid. The general expressions for the source functions (Hasselmann 1962) are the wind energy input, the nonlinear transfer of energy from one frequency to another by wave-to-wave interactions, and the energy dissipation by wave breaking, turbulence, and bottom friction. These source functions are then allowed to result growth of the wave spectrum as a function of time and space.

The SWAMP (Sea Wave Modeling Project) Group – a group of ten institutes from Germany, Italy, Japan, the Netherlands, Norway, the UK, and the USA – has examined and tested ten numerical models on a common platform and has published their findings in 1985 (SWAMP Group 1985). They classified the existing models into three groups: decoupled propagation (DP) models, coupled hybrid (CH) models, and coupled discrete (CD) models. The DP is a first-generation wave model in which the spectral components evolve

independent of each other in accordance with the linear input of source function. The CH models work on the basis of a hybrid combination of parametrical windsea models with the standard discrete spectral representation for the swell components. The limitations of DP and CH models are eliminated in the CD models, which are based on discrete spectral representation for the entire spectrum of windsea and swell. WAVEWATCH III (Tolman 1999) is a third-generation model of this type developed at NOAA/NCEP in the spirit of wave modeling initiative (WAMDIG 1988; Komen et al. 1994). It represents a further development of the original WAVEWATCH I model developed at the Delft University of Technology (Tolman 1991).

Summary

This entry highlights the state-of-the-art methods that are applied for prediction of wind-generated waves. By defining the basic parameters and outlining the brief history of hindcasting methods, the entry describes empirical, spectral, and numerical methods. The hindcasting techniques known as the SMB method, the JONSWAP spectrum, and the Goda-Wilson method are discussed with examples. The publicly available numerical wave hindcasts from the SWAMP group and the WAVEWATCH are also briefly outlined.

Cross-References

- [Beaufort Wind Scale](#)
- [Coastal Warfare](#)
- [Coastal Modeling](#)

- Time Series Modeling
- Wave Climate
- Wave Environments
- Waves

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Wave Power

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Of the “implemented” systems to extract energy from the ocean, power from the waves is perhaps the one that has received most attention during the twentieth century, from Pacifica (California) to Royan (France). The number of patents taken out, and of systems proposed, is impressive. Projects are being examined in several locations in Asia and the Pacific Ocean, in Canada and in Europe. Waves are considered a “small energy source” that can be profitably tapped by industrialized and less-developed countries alike.

Extraction systems utilize either the vertical rise and fall of successive waves in order to build water- or air-pressure to activate turbines, or take advantage of the to-and-fro, or rolling, motions of waves by vanes or

Roger H. Charlier: deceased.

cams which rotate turbines. Still another approach concentrates incoming waves in a converging channel, thus allowing the buildup of a head of water, which then makes it possible to operate a turbine. Wave refraction effects can be used to focus wave energy and they have played an expanding role in current wave-energy conversion thinking. However, in shallow waters the amount of such energy available is reduced due to shoaling and seafloor friction.

Most recently proposed schemes show economic feasibility. Japanese and British researchers have been in the forefront of research, though the Japanese have been closer to implementation, and Scandinavians have overtaken the British.

Conversion Devices

Systems involve a movable body, an oscillating column, or a diaphragm and the “object” that moves floats is anchored on the seabed. Attenuator type devices in a two-dimensional case will provide only a 50% efficiency, unless dispersion waves strike the object only on one side. A superior apparatus must have a high absorption efficiency for a wide range of wave frequencies. Most currently considered primary wave converters are linear, two- or three-dimensioned. Among the latter are those systems known as Kaimei, Flexible Bag, and National Electronic Laboratory’s Oscillating Water Column (NEL’s OWC). Converters involving a movable body use vertical, rotational, lateral, coupled movable bodies or a raft or float. They include, respectively, the Point Absorber, Salter Duck (Fig. 1), Wave Power Water Turbine, Bristol Cylinder, and Cockerell Raft. There are systems that double as breakwaters, such as the Pendulum and NEL’s OWC. Some apparatus are moored, fixed on the shore or on the seabed. The Focusing Wave Energy System is in a somewhat separate category.

Based on *uses* energy conversion devices can be grouped into four “classes”: propulsion, buoy power supply, offshore plants, shore-based power plants. Another grouping considers rather conversion *methods*: utilization of the rise and fall sequence, of the rolling motion, or convergence to create a hydraulic head. A “*physics principles*” classification recognizes: intervention in wave orbits, utilization of the pressure field, acceleration devices, and use of the horizontal transport from breaking waves. A much older (1892) classification was based on *mechanical concepts*: motors operated by the rise and fall of a float, by to-and-fro wave movement, by the varying slope of wave surface, and by impetus of waves rolling up a beach. With about 40 different *systems* proposed over time still another grouping is possible: surface profile variations of traveling deepwater waves, subsurface pressure variations, subsurface fluid particle motion, and naturally or artificially induced unidirectional motion of fluid particles in a breaking wave.

Systems may intervene in the wave orbits, utilize the pressure field, be accelerative or utilize the water mass horizontal displacement. They may involve flaps and paddles; heaving, pitching and rolling bodies, pressure devices, rotating outriggers, pneumatic or cavity resonators, or be waves focusing, surging, or mixed systems.

Performance of Systems

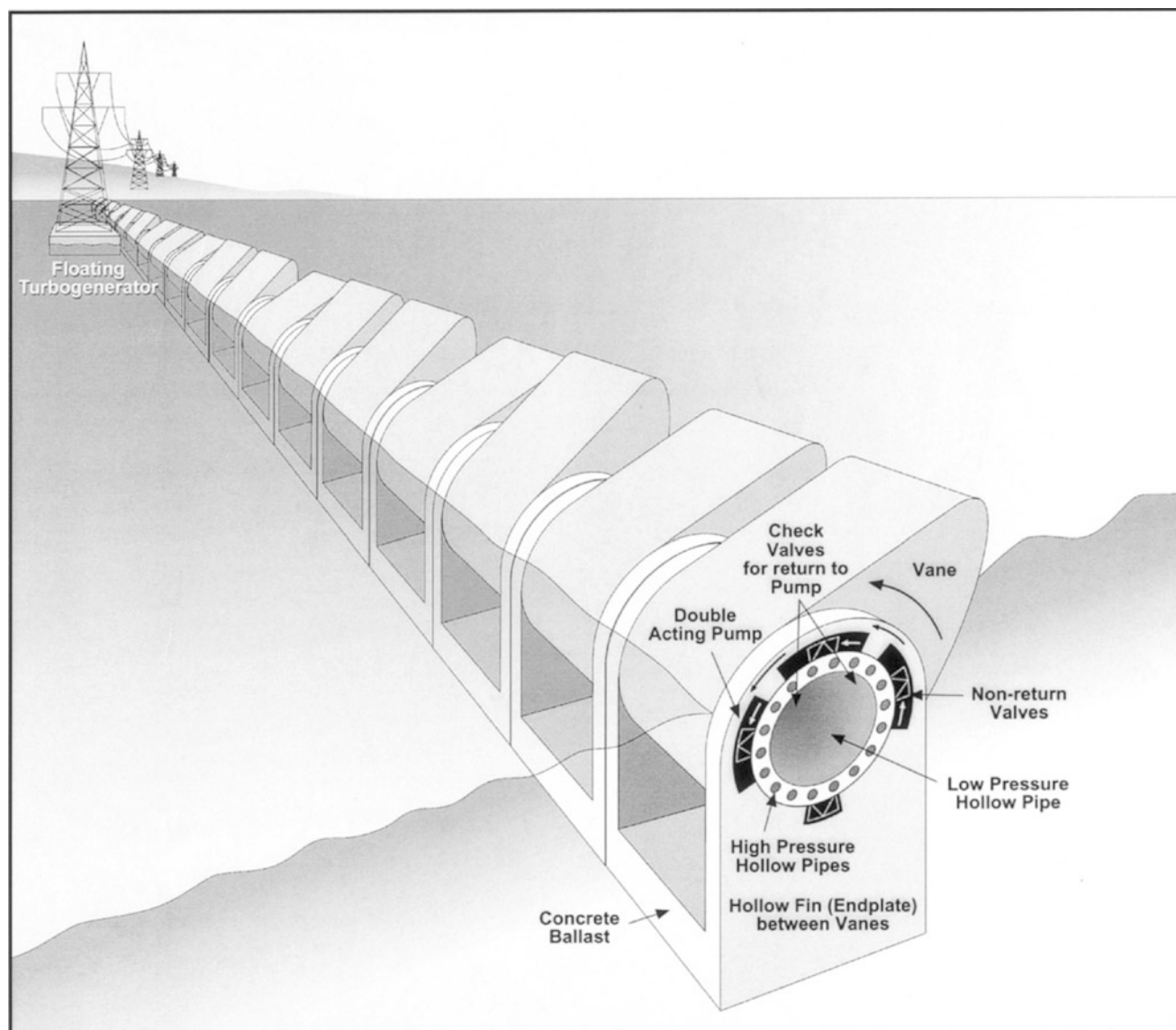
Satisfactory performance and economically sustainable results appear to have been achieved by several systems. Charlier and Justus (1993) list an air turbine driven by water oscillation in a vertical borehole (Royan, France, Gironde River estuary), floats activated by horizontal and vertical motion attached to a pier (Atlantic City, New Jersey and Pacifica, California-USA), a Savonius rotor operating pump (Monaco, Musée Océanographique), pump operated by a rising and falling heavy float (Monaco), low-head hydroelectric plant supplied from a forebay with converging channels (Pointe Pescade and Sidi-Ferruch, Algeria), air turbine buoys (Japan, United States, United Kingdom), air turbine generator (Osaka, Japan), Kamei barge with compressed-air chambers (Japan), hydraulic pumping over pliable strips in a concrete trough (Boston, MA, USA), autobailer bilge pump (Sweden), and a sea-lens concentration scheme (Norway). Still other schemes which either function or show promise are utilized in bouys and desalination plants.

The machine must be able to amplify the relatively low water head, have a broad range of response, be nonresonant, and have the capacity of responding to smaller waves, while simultaneously be able to withstand the effects of large storm waves.

Recent Developments

Just before the dawn of the nineteenth century a patent was taken out in France to turn wave energy into mechanical energy by using a float. Several wave motors were patented in the 19th and 20th centuries which supplied electrical energy for buoys, pumps, lighthouses, and even small residences. Duckers (1989), Scheer and Gandhi (1994), Charlier and Justus (1993), Ross (1985, 1987), and Salter (1975a, b, Salter et al. 1976) among others, discussed the more recent devices tested and put into service. OWCs are perhaps the most promising scheme; indeed, being robust and utilizing relatively conventional, proven technology.

In 1990, Wave Energy Inc. took out patents contracts (hydraulic method) on the heels of several successful tests carried out at Scripps Institution of Oceanography for a wave pump and energy extraction system. Norwave conducted a feasibility study for a 5 MW plant on Java. Sixteen prototypes



Wave Power, Fig. 1 Schematic view of “Salter’s Duck” (from Charlier and Justus 1993, with permission from Elsevier Science). Mighty Whale is a Japanese constructed offshore floating wave power device, 50 m long and 30 m wide, the world’s largest. It was moored in late 1998 off Gokasho Bay. Resembling a whale, it transforms wave motion into high

velocity oscillating airflow, achieved near the “mouth,” by three chambers; waves enter and exit through these ducts at their top, with air turbines driving electric power generators (one each of 10 kW and 50 kW and two of 30 kW). Electricity drives the air compressors

and ten “future prototypes” were listed by Duckers for the period 1965–92; several were operational for a while, but eventually abandoned; they were mostly installed in Japan, with one each in Great Britain, Denmark, Norway, and India, or planned for Japan, Scotland, the United States, among other locations. Eire has concentrated on devices and so have Korea and China; a small power plant (8 KW) was installed in the Pearl River estuary around 1991. China’s interest in wave power has not diminished (Zhi 1993; You and Zhi 1995) and India has conducted surveys and tests (Sivaramakrishnan 1992; Raju and Ravindram 1996). Ocean Power Technology (USA) developed in 1995–96 a power

system using the piezoelectric effect to generate electricity from waves; the bending of laminated sheets suspended from rafts secured on the seafloor generate electricity. The European Union supported in 1994 the construction of a full-scale 2 MW converter of the OWC type in Scotland (You and Zhi 1995).

Environmental Impact

Wave energy is clean, safe, environment friendly, but its harnessing affects beaches, marine life, fishing, shipping,

coastal tourism and recreation, and should be viewed in the overall picture of integrated coastal management.

Though limited, wave-energy utilization has thus some environmental consequences, biological, physical, and social. Marine organisms will be moderately affected, but may pose biofouling problems for the system. Fisheries will be impacted, differently according to species, though floating structures may act as a fish reef. More severe impact on fisheries is caused by seabed fixed apparatus.

Longshore currents may be rerouted, tidal currents influenced, particularly with apparatus fixed on the seabed, and coastlines may consequently be modified. Navigational problems may arise. Wave absorbers alter wave patterns and drift patterns of sand, thus possibly causing coastal erosion.

Production of electricity at an affordable cost, particularly in power deprived developing countries areas, may lead to implantation of new industries, and/or improvement of existing ones. A social and developmental “plus” may flow forth.

The Future

If wave energy is to make a significant contribution to national energy demand, sustained research is needed into its application to the offshore production of hydrogen. Wave power has also been proposed for use in communications and spacecraft propulsion (Korde 1990) and continues to be eyed for desalination plants (Crerar and Pritchard 1991). In some countries, for example, Norway, successful plants have been abandoned because of the availability of huge oil reserves; only when these will have dwindled, will new enthusiasm for alternative sources of energy be rekindled there.

Cross-References

- [Desalination](#)
- [Engineering Applications of Coastal Geomorphology](#)
- [Geohydraulic Research Centers](#)
- [Tidal Power](#)
- [Tide Mill](#)
- [Wave Environments](#)
- [Wave-Dominated Coasts](#)
- [Waves](#)

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Wave Refraction Diagrams

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Wave refraction diagrams are used to illustrate and predict the refraction of waves approaching the shoreline. They are an invaluable tool for understanding coastal morphology and processes, and their construction is practically standard practice in coastal engineering applications. Amongst others, wave refraction diagrams can be used to compute alongshore variations in wave energy and breaker angles, and subsequently longshore sediment transport rates and directions (see “► [Longshore Sediment Transport](#)”). Wave refraction diagrams are also used to indicate the stability of embayed beaches (see “► [Headland-Bay Beach](#)”). A good summary and comparison of techniques to construct wave refraction diagrams is given in Wiegel (1964).

Theoretical Background

The theoretical background to the wave refraction process is discussed in this encyclopedia’s article entitled: ► [Waves](#). Briefly, wave refraction is the bending of wave rays that occurs in intermediate and shallow water depth. As a result of wave refraction, wave crests become increasingly lined up with the bottom contours, thereby reducing the angle of wave

approach. The refraction of waves is similar to the bending of light rays and the change in direction is related to the change in the wave celerity C through the same Snell's law

$$\frac{\sin \alpha_1}{C_1} = \frac{\sin \alpha_2}{C_2} = \text{constant}, \quad (1)$$

where α_1 is the angle of a wave front with the depth contour h_1 over which the wave is passing, α_2 is a similar angle measured as the wave front passes over the second depth contour h_2 , C_1 is the wave celerity at the depth of the first contour, and C_2 is the wave celerity at the second contour (refer to Fig. 7a in this encyclopedia's article entitled: ► [Waves](#)).

Wave Refraction for Coasts with Parallel Bottom Contours

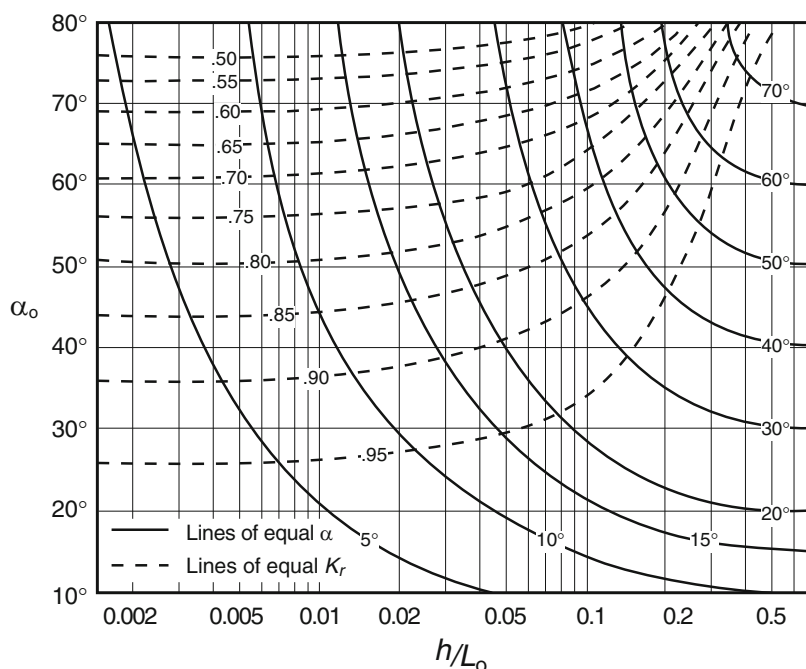
Changes in wave direction due to wave refraction on a straight coast with parallel bottom contours can be determined analytically using

$$\sin \alpha = \frac{C}{C_o} \sin \alpha_o, \quad (2)$$

where α is the angle between wave crest and depth contour at an arbitrary depth, α_o is the angle between wave crest and depth contour in deepwater, C is the wave celerity at an arbitrary depth, and C_o is the deepwater wave celerity. Computing $\sin \alpha$ for different water depths simply involves solving Eq. 2. The relevant equations to do this computation are given in this encyclopedia's article entitled: ► [Waves](#).

Wave Refraction Diagrams,

Fig. 1 Change in wave direction and height due to refraction on a straight coast with parallel bottom contours (after CERC 1984) (h = water depth, L_o = deepwater wave length, α_o = deepwater wave angle, α = wave angle, K_r = refraction coefficient)



Wave Refraction

Diagrams, Table 1 Wave transformation characteristics for an incident wave with a deepwater wave height of 1 m, a wave period of 8 s and a deepwater wave angle of 45°

h (m)	L (m)	C (m/s)	n (—)	α (°)	K_s (—)	K_r (—)	H (m)
100	99.9	12.5	0.50	45	1	1	1
50	99.4	12.4	0.51	45	0.99	1	0.99
30	95.9	12.0	0.58	43	0.95	0.98	0.93
20	88.7	11.1	0.67	39	0.92	0.95	0.88
15	81.8	10.2	0.73	35	0.91	0.93	0.85
10	70.9	8.9	0.81	30	0.93	0.90	0.84
5	53.1	6.6	0.90	22	1.02	0.87	0.89
3	42.0	5.3	0.94	17	1.13	0.86	0.97
2	34.7	4.3	0.96	14	1.23	0.85	1.05

h water depth, L wave length, C wave celerity, n coefficient, α wave angle, K_s shoaling coefficient, K_r refraction coefficient, H wave height

Alternatively, a nomograph can be used to determine changes in wave height and direction due to refraction (Fig. 1). For example, for a deepwater wave α_o angle of 60° and a deepwater wave length L_o of 100 m, the wave angle in 10 m water depth is 38° and the refraction coefficient K_r is 0.79. Table 1 lists an example of wave refraction (and shoaling) over a bathymetry with parallel depth contours. It can be seen that wave shoaling results in an increase in wave height with decreasing depth ($K_s > 1$), whereas wave refraction reduces the wave height ($K_r < 1$).

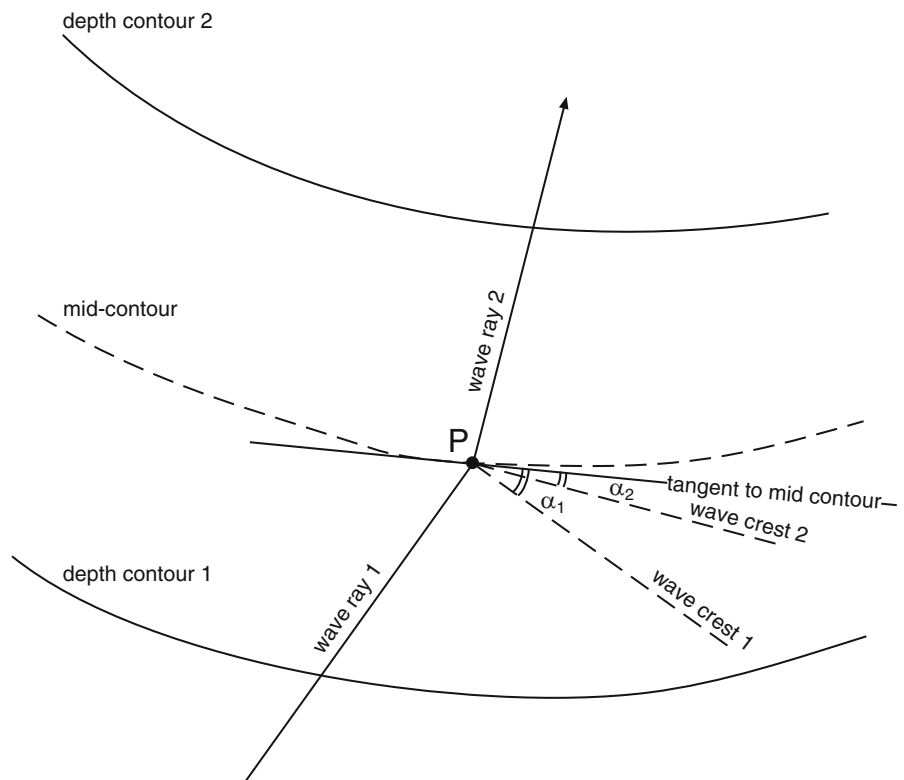
Wave Refraction for Coasts with Complex Bottom Contours

In reality, parallel depth contours rarely exist and in this case the wave refraction pattern has to be determined manually. Refraction diagrams for such conditions can be constructed in two ways. The first, known as the “wave-front method,” is essentially a map showing the wave crests at a given time or the successive positions of a particular wave crest as it propagates shoreward (Johnson et al. 1948). Wave rays or orthogonals can then be drawn perpendicular to the wave crests to indicate the direction of wave travel. In the second method, known as the “orthogonal method,” the wave rays are drawn directly (Arthur et al. 1952). The orthogonal method is considered the most practical, quickest, and widely used technique at present and will be discussed here.

A number of assumptions underly the orthogonal technique: (1) wave energy between wave rays remains constant, that is, there is no leakage of wave energy perpendicular to the direction of wave travel; (2) the direction of wave advance is parallel to the wave rays; (3) the celerity of a wave of a given period depends only on the water depth at that location; (4) changes in seabed topography approaching a coastline are gradual; (5) waves are long-crested, constant period, small amplitude, coastline and monochromatic; and (6) other factors that may affect wave refraction, such as currents, winds, and reflected waves are ignored.

Before constructing the wave rays, a tracing paper overlay is placed on the bathymetric chart. The shoreline and the bottom contours are then traced with all bottom irregularities within a local spatial scale of five wave lengths smoothed out. Mid-contour lines are then drawn between the existing contour lines. Subsequently, a set of evenly spaced and parallel wave rays from the chosen direction of wave approach are plotted up to the bottom contour closest to $0.5L_o$ and extended to the first mid-contour line. The number of wave rays should be selected to sufficiently cover the coastline. The closer the spacing between the rays, the greater the resolution when estimating wave energy at the coastline. Finally, for each mid-contour line the C_1/C_2 ratio should be computed, where C_1 represents the wave celerity of the deeper depth contour and C_2 represents the wave celerity of the shallower depth contour.

Wave Refraction Diagrams,
Fig. 2 Construction of wave rays using a protractor (α_1 = wave angle of incoming wave, α_2 = wave angle of refracted wave)



Starting with any one wave ray the following steps are taken in extending the wave ray from offshore to the coastline (Fig. 2).

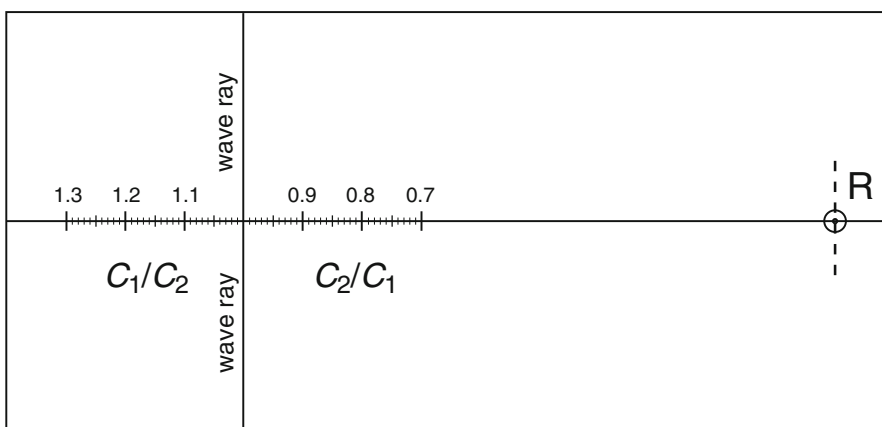
1. Extend the incoming wave ray to the mid-contour and determine the intersection point P.
2. Construct a tangent line to the mid-contour at point P.
3. Measure the angle α_1 between the tangent line and wave crest 1 (normal to wave ray 1) using a protractor.
4. Compute the angle α_2 between the tangent line and wave crest 2 using Eq. 1 and the C_1/C_2 ratio.
5. Using a protractor, construct wave ray 2 which is the line normal to wave crest 2 going through the intersection point.

6. Repeat the procedure for successive contour intervals and all incoming orthogonals.

Manually measuring and plotting wave angles using a protractor can be quite cumbersome. Templates have been constructed that preclude the manual measurement of wave angles. The template shown in Fig. 3 is most commonly used and will be described here. The template should be constructed such that the length of the template is about 25 cm long and should be printed or copied on transparency paper. The following steps are taken in extending the wave ray from offshore to the coastline (Fig. 4).

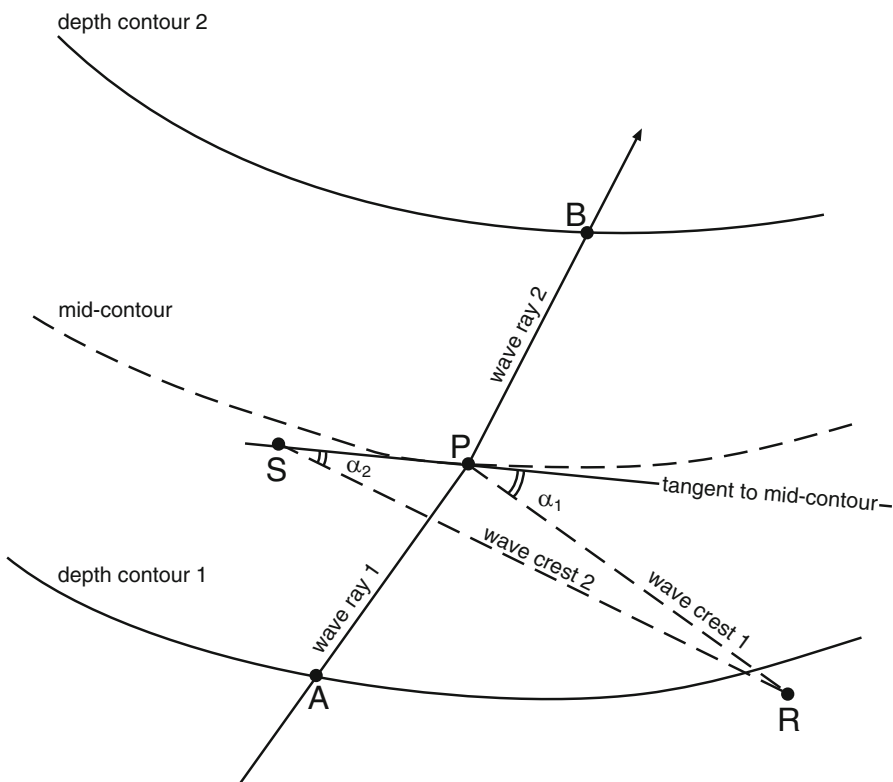
Wave Refraction Diagrams,

Fig. 3 Refraction template (after CERC 1984) (C_1 wave celerity at contour 1, C_2 wave celerity at contour 2)



Wave Refraction Diagrams,

Fig. 4 Construction of wave rays using a template (α_1 wave angle of the incoming wave, α_2 = wave angle of refracted wave)



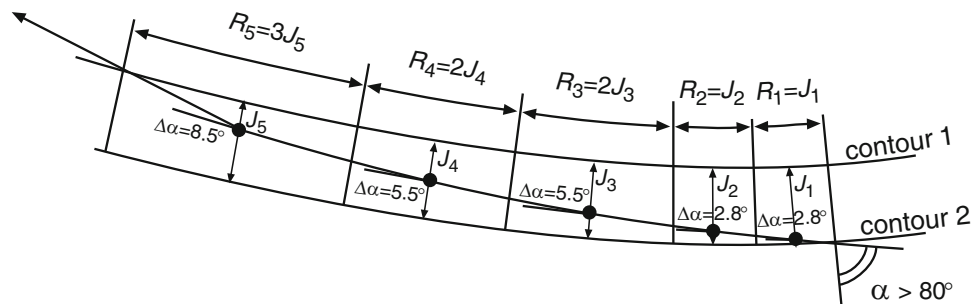
1. Extend the incoming wave ray to the mid-contour and determine the intersection point P.
2. Construct a tangent line to the mid-contour at point P.
3. Place the line on the template labelled "wave ray" along the incoming wave ray with the point representing C_1/C_2 1.0 at point P.
4. Rotate the template about the turning point R until the C_1/C_2 value corresponding to the contour interval being crossed intersects the tangent to the mid-contour at point S. The refracted wave ray now lies in the direction of the turned wave ray on the template.
5. Construct a line parallel to the turned wave ray on the template from P representing the refracted wave ray.
6. Repeat the procedure for successive contour intervals and all incoming wave rays.

Strictly speaking, point P is not the correct intersection point between the incoming and refracted wave ray. The "true" intersection point P is found by moving parallel to the turned wave ray until the distance AP is equal to the distance BP (Fig. 4). This ensures that the distances travelled by the incoming and refracted wave rays within the contour interval C_1-C_2 are equal. However, if the change in depth between the two contours is small and $C_1/C_2 = 1.2$, P and P' will almost coincide and it is permissible to construct the refracted wave ray from P rather than P'.

For wave angles larger than 80° , the orthogonal method is no longer reliable and an alternative method of constructing wave rays must be used. The R/J method is generally used for this purpose and consists of the following steps.

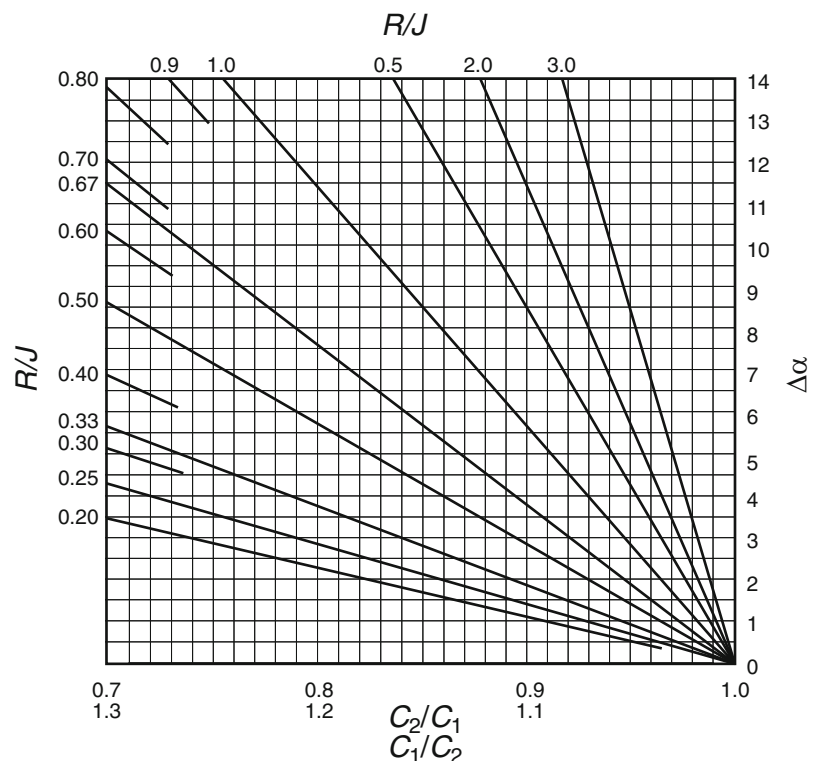
Wave Refraction Diagrams,

Fig. 5 Construction of wave rays using the R/J method when the wave angle α is larger than 80° (after CERC 1984). The wave ray has been constructed assuming $C_1/C_2 = 0.95$ (R = distance along wave ray, J = distance between contours, $\Delta\alpha$ = turning angle of the wave ray)



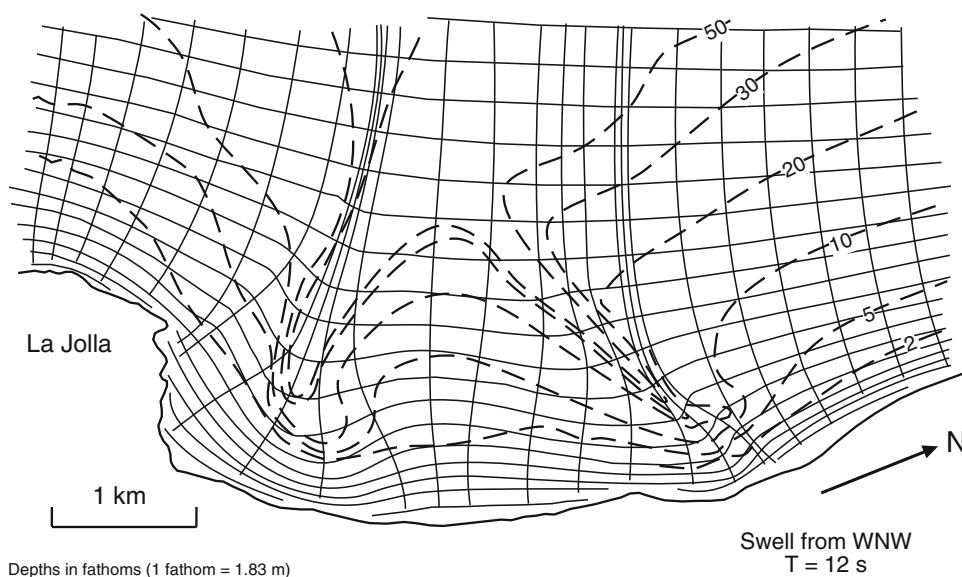
Wave Refraction Diagrams,

Fig. 6 Nomograph to determine the turning angle of the wave ray $\Delta\alpha$ from the ratios R/J and C_1/C_2 (after CERC 1984) (R = distance along wave ray, J = distance between contours, C_1 = wave celerity at contour 1, C_2 = wave celerity at contour 2)



Wave Refraction Diagrams,

Fig. 7 Wave refraction over submarine canyons and along the headland at La Jolla, CA (after Munk and Traylor 1947)



1. Divide the contour interval to be crossed (from h_1 to h_2) into segments by transverse lines drawn perpendicular to the contours (Fig. 5). The spacing R of the transverse lines is arbitrarily set at some small number times the distance J between the contours.
2. Determine the turning angle of the wave ray $\Delta\alpha$ from the R/J nomograph (Fig. 6).
3. Turn the wave ray by the angle $\Delta\alpha$ and extend it to the middle of the segment.
4. Repeat the procedure for each segment in the sequence.

Once the wave ray crosses a new contour and the wave angle α is less than 80° the R/J method should be replaced by the orthogonal method.

Both the orthogonal and the R/J method can be used to construct orthogonals from shallow to deepwater. In this case, the same procedure can be employed, except that the ratio C_1/C_2 (>1) is replaced by C_2/C_1 (<1).

The variation in wave height due to wave refraction can be determined based on the principle that the transported energy between two adjacent wave orthogonals is conserved. By measuring the width between two orthogonals at depth h and in deepwater the refraction coefficient K_r is given by

$$K_r = \left(\frac{b_0}{b} \right)^{0.5}, \quad (3)$$

where b and b_0 are the width between two wave rays at depth h and in deepwater, respectively.

An example of a wave refraction diagram is shown in Fig. 7 which indicates the refraction pattern associated with submarine canyons. Note that the spacing of the wave rays is not constant; where more detail is required, such as in the vicinity of the canyons, the wave rays are spaced closer together. Strong

focusing of the wave rays, known as wave convergence, occurs at the headland. At this location, $K_r > 1$ and relatively energetic wave conditions will prevail. Spreading of the wave rays, known as wave divergence, takes place at the heads of the canyons. As a result, $K < 1$ and the coastline landward of the canyons will experience relatively calm wave conditions.

Numerical Wave Refraction Models

In the past, wave refraction diagrams were invariably constructed manually. Currently, numerical models are often used to construct wave refraction diagrams (e.g., Dalrymple 1988). Such models, which often include the effects of wave diffraction, are much faster and more accurate than the manual method, but require considerable technical expertise and preparatory work (e.g., digitising the bathymetry). Rather than yielding orthogonals, the numerical analyses obtain solutions for wave direction and wave height for a finite number of grid cells that cover the area of interest. The obvious advantage of numerical wave refraction models is that once the model has been set up and found to perform satisfactory, a large number of scenarios (different incident wave direction and period) can be investigated with relative ease.

Cross-References

- [Engineering Applications of Coastal Geomorphology](#)
- [Headland-Bay Beach](#)
- [Longshore Sediment Transport](#)
- [Wave Focusing](#)

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Wave-Current Interaction

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Flows observed in shallow marine environments are frequently combinations of currents and waves. Currents, with variations on timescales of hours or days, are principally tidal, wind-driven, or related to larger-scale patterns of ocean circulation. Wave-generated oscillatory flows, with periods on the order of seconds to tens of seconds are produced by surface gravity waves, though lower frequency internal waves can also be present. When wave-generated oscillatory motion of the water column extends to the seabed, wave-current interactions lead to increases in bed shear stress and the bottom roughness affecting bottom-boundary-layer currents. These effects are important for understanding ocean circulation and sediment transport in coastal environments.

Currents and wave-driven flows interact with the seabed to produce boundary layers where frictional effects are important. The ubiquitous presence of currents in the ocean results in a turbulent bottom boundary layer with a thickness on the order of meters to a few tens of meters. Waves produce a boundary layer only when wave oscillatory motion extends to the seafloor, a condition that is satisfied when water depth is less than half the wavelength of surface gravity waves. The boundary layer produced by waves in the ocean is most often turbulent, with a thickness of the order of 0.1 m. The disparity in wave- and current-boundary-layer thickness has several important consequences, including wave-generated bed shear stresses that are generally higher than current-generated stresses, the confinement of combined wave-current turbulence to the region of the wave boundary layer, and a large gradient in shear stress and

turbulent mixing near the top of the wave boundary layer when wave shear velocities are significantly larger than current shear velocities.

Theory

Analyses of wave-current interactions are commonly carried out on timescales long enough to average over many individual waves and short enough that the currents can be considered quasi-steady. Hourly averaged currents and wave conditions are often used to characterize effects of wave-current interaction. Near the seafloor, the velocity profile associated with the currents is logarithmic under steady, uniform flow conditions

$$u_c(z) = \frac{u_{*c}}{\kappa} \ln \frac{z}{z_{0c}},$$

where u_c is current velocity, u_{*c} is current shear velocity, κ is von Karman's constant (0.41), z is elevation above the seabed and z_{0c} is a roughness parameter. In the absence of waves, z_{0c} is related to the physical roughness of the bed. When waves are present, an apparent roughness associated with the presence of the wave boundary layer dominates the roughness parameter z_{0c} . The applicability of the equation above is limited to nearbed depths greater than the height of the wave boundary layer, δ_w . Wave boundary layer height, assuming turbulent flow, can be estimated as $\delta_w = u_{*w}/\omega$, where u_{*w} is wave shear velocity (discussed below), $\omega = 2\pi/T$, and T is wave period.

The velocity of wave-generated water motion varies throughout the wave period. The wave-generated velocity just above the seabed (at the top of the wave boundary layer), termed the bottom or near-bed wave orbital velocity, u_b , can be related to surface wave conditions and water depth using linear wave theory. For monochromatic waves

$$u_b = u_{bm} \cos \omega t, \quad u_{bm} = \frac{\pi H}{T \sin h(kh)},$$

where u_{bm} is maximum near-bed orbital velocity, t is time, H is wave height, $k = 2\pi/L$, L is wavelength, and h is water depth. When a spectrum of waves is present, as is typical of coastal environments, u_{bm} is most accurately determined by applying the above equation to each frequency band in the full wave spectrum to determine the average or significant maximum near-bed orbital velocity (e.g., Madsen 1994); significant orbital velocity is analogous to significant wave height.

The time-varying wave orbital velocity at the top of the wave boundary layer results in a time-varying bed shear stress. It is common practice in calculations of

wave-current interaction to characterize wave-generated bed shear stress, and associated turbulent mixing within the wave boundary layer, in terms of the maximum value of near-bed wave orbital velocity, u_{bm} . Wave shear velocity, u_{*w} , or bed shear stress, τ_{bw} , can be related to u_{bm} through the equations for an oscillatory boundary layer (e.g., Smith 1977; Grant and Madsen 1979) or a wave friction factor, f_w :

$$\tau_{bw} = \frac{\rho}{2} f_w u_{bm}^2,$$

where ρ is fluid density. Wave friction factor is a function of wave orbital amplitude ($a H/[2 \sin h(kh)]$) and bed roughness, k_s . A number of empirical relationships for f_w are available, such as:

$$f_w = 0.04 \left(\frac{a}{k_s} \right)^{-1/4} \quad \text{for } \frac{a}{k_s} > 50,$$

(Fredsoe and Deigaard 1992). Because of the dependence of u_{bm} water depth, τ_{bw} decreases with increasing depth for a given set of wave conditions.

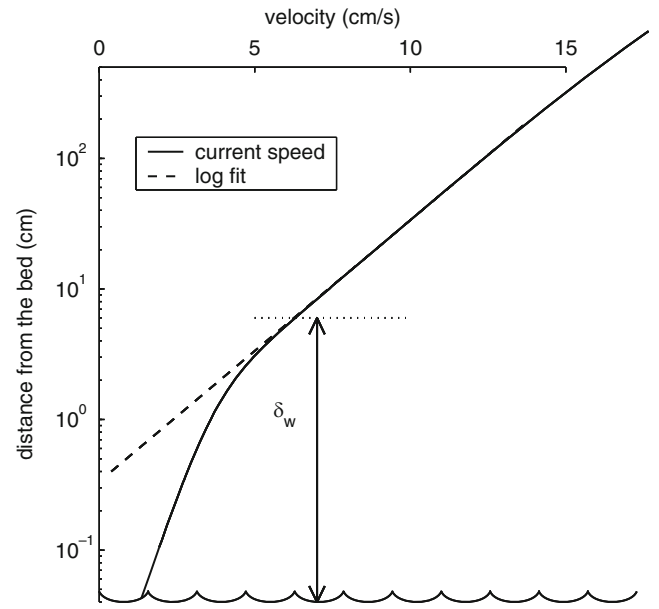
When waves and currents both contribute to near-bed flow, the velocity profile of the currents within the wave boundary layer can be expressed as

$$u_c(z) = \frac{u_{*c}^2}{K u_{*cw}} \ln \frac{z}{z_{0c}}, \quad z < \delta_w,$$

where u_{*cw} is the combined wave-current shear velocity and z_0 is the hydrodynamic roughness related to the physical roughness of the bed. The combined flow shear velocity, u_{*cw} , depends on the shear velocities associated with the current, u_{*c} , and with the waves, u_{*w} . The wave-current shear velocity can be found from the combined-flow bed shear stress as,

$$\tau_{bcw} = [(\tau_{bw} + \tau_{bc} \cos \varphi)^2 + \tau_{bc}^2 \sin^2 \varphi]^{1/2} = \rho u_{*cw}^2$$

where φ is the angle between the direction of wave propagation and the direction of the currents, and τ_{bc} , τ_{bw} , and τ_{bcw} are the current, wave, and combined wave-current bed shear stresses, respectively. In this case, wave boundary layer thickness and turbulent mixing within the wave-boundary-layer scale with u_{*cw} rather than u_{*w} .



Wave-Current Interaction, Fig. 1 Profile of near-bed current illustrating the effect of wave-current interaction. Profile slope changes near the top of the wave boundary layer (δ_w , indicated by the horizontal dotted line) owing to differences in shear velocity above and below δ_w when waves and currents are both present. The slope of a semi-logarithmic line fit to the profile above δ_w is related to current shear velocity, u_{*c} , and the zero-velocity intercept of the line is the apparent bottom roughness felt by the current, z_{0c} . The profile shown in the figure is cut off above the level, z_0 , at which the velocity reaches zero.

Solutions and Applications

Solving the equations given above for $u_c(z)$ requires specification of z_0 and matching the profiles for current velocity above and below the wave boundary layer at the top of the wave boundary layer. Often, particularly in shallow water, wave-boundary-layer flow is fully rough, so that 0 is just proportional to the physical roughness length of the bed, k_s . Wave-formed ripples on the bed typically dominate bottom roughness in shallow, sandy coastal environments, though grain roughness and saltating grains also contribute to bottom roughness, particularly when ripples are very small or absent. Current shear velocity, u_{*c} , which also must be specified when calculating the velocity profile $u_c(z)$, is seldom known in advance. In practice, a measured value of current velocity at some level near the bed is specified, and an iterative calculation is performed to determine the value of u_{*c} that is consistent with the measured velocity given the bottom roughness parameter. The equations for $u_c(z)$ provided here assume a linear eddy viscosity of the form $\nu_e = \kappa u_* z$; within the wave boundary layer $u_* = u_{*cw}$ while above it $u_* = u_{*c}$. This form of eddy viscosity, while yielding a simple analytical solution for $u_c(z)$, is discontinuous at the top of the wave boundary layer. Other formulations for ν_e exist that are continuous across the top of the wave boundary layer (e.g., Wiberg and

Smith 1983), but most of these require a numerical solution for $u_c(z)$. Higher-order closures have also been used.

Solving the equations for current velocity within and above the wave boundary layer yields a characteristic profile exhibiting an inflection near the top of the wave boundary layer (Fig. 1). Current velocity within the wave boundary layer is more uniform than is the profile above it owing to the larger shear velocity (u_{*cw}) in the wave boundary layer. The zero-velocity intercept of a semilogarithmic line drawn through the velocity profile above the top of the wave boundary layer gives the apparent roughness felt by the currents above the wave boundary layer (Fig. 1). The equations for $u_c(z)$ given above can be rearranged to solve for the apparent roughness, z_{0c} , giving

$$z_{0c} = \delta_w \left(\frac{z_0}{\delta_w} \right)^\lambda, \quad \lambda = \frac{u_{*c}}{u_{*cw}}.$$

Resulting values of apparent roughness are often several orders of magnitude larger than the roughness parameter associated with the physical roughness of the seabed.

Field observations of bottom boundary flow motivated the initial formulation of the theory of wave–current interaction, but the first comprehensive wave–current interaction models (Smith 1977; Grant and Madsen 1979) were developed before there were adequate data available to test them. Since that time, many field studies of near-bed waves and currents have been carried out in wave-dominated coastal environments around the world. Some laboratory studies have also been performed. These have demonstrated the importance of waves in increasing bed shear stress above values generated by currents alone. Under storm conditions, waves can dominate the combined wave–current bed shear stress across most or all of the continental shelf. Calculated and measured values of current shear velocity have been shown to agree reasonably well in a number of studies. However, calculated values depend on bed roughness a term that is difficult to predict because of its dependence on bedform height and spacing, which vary in ways that are not fully understood. Nonuniform flows and stratified flows also complicate the application of wave–current interaction models to field data.

Increases in bed shear stress, turbulent mixing within the wave boundary layer, and apparent bottom roughness associated with wave–current interaction at the seabed have implications for a variety of processes. For example, observations from a number of continental shelves have concluded that bottom drag coefficients for near-bed currents can be increased by up to an order of magnitude larger when waves are present. This is important for studies of wind-driven coastal ocean circulation. Effects of wave–current interaction on bottom drag have been incorporated into models of ocean circulation applied to the coastal ocean.

Sediment transport in coastal environments is strongly affected by wave–current interactions. Coastal currents in many locations are insufficient to mobilize sediment outside of the nearshore zone. In these regions, it is only when moderate to high wave conditions are present that bed shear stresses are sufficient to mobilize sediment at the seabed.

Even at depths of 50–100 m, mean winter combined wave–current shear velocities can be considerably larger than current shear velocity. The wave boundary layer, because of its high shear velocity, is able to maintain both higher concentrations of sediment and coarser sediment in suspension than can the overlying current boundary layer. When differences between u_{*cw} and u_{*c} are large, gradients in suspended sediment concentrations near the top of the wave boundary layer can cause significant stratification which inhibits turbulent mixing and modifies near-bed flow and sediment transport.

Cross-References

- [Coastal Currents](#)
- [Coastal Upwelling and Downwelling](#)
- [Cohesive Sediment Transport](#)
- [Rip Currents](#)
- [Ripple Marks](#)
- [Surf Modeling](#)
- [Wave-Dominated Coasts](#)
- [Waves](#)

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Wave-Dominated Coasts

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Price (1955) used the term *wave-dominated coasts* to describe the morphology of depositional shores where consistent, relatively large waves, with their associated strong wave-generated currents, have produced a smoothed shore of sandy sediments. Where such shores are building seaward (prograding), sand beaches associated with long barrier islands and prograding beach ridges are the dominant intertidal habitats. As a generalization, coasts with small tidal ranges (microtidal = <2 m (Davies 1964)) are dominated by wave energy, thus most wave-dominated coasts are microtidal. On the other hand, coasts with large tidal ranges (macrotidal = >4 m) are typically dominated by tidal energy, and were hence termed *tide-dominated coasts* by Price. Coasts with intermediate wave energy and tidal ranges (typically mesotidal = 2–4 m) were termed *mixedenergy coasts* by Hayes (1979).

Wave-Dominated Non-deltaic Coasts

Between major rivers on wave-dominated coasts, the shore is typically occupied by long, uninterrupted barrier islands. Inasmuch as wavedominated barrier islands occur most commonly on microtidal coasts, the barrier islands are typically backed by open-water bays and lagoons. The sediment patterns of barrier islands show a simple gradation of grain size from coarse to fine away from shore as a result of decreasing bottom agitation by waves with increasing water depth.

Barrier islands occur primarily on depositional coasts on the trailing edges of continental plates, or on enclosed tideless seas such as the Baltic and the Mediterranean. Barrier islands are the dominant coastal type along the Atlantic and Gulf

coasts of the United States, with those on the Gulf Coast of Texas, the Outer Banks of North Carolina, and Northeast Florida being wave-dominated. Tidal inlets are widely spaced on wave-dominated barrier islands (e.g., over 40 km apart in North Carolina). See Davis (1994) for a detailed synthesis of the barrier islands of the United States.

Most barrier islands have formed within the last 4,000–5,000 years during a near stillstand in sea level. Two types of barrier islands may be present – those that consistently migrate landward (retrograding) and those that build seaward (prograding). How an island builds depends upon the ratio of relative sea-level change to sediment supply. Where the sediment supply is inadequate or sea level is rising rapidly, the island retreats landward. Sediment supply for an island can be diminished through several natural mechanisms and by dams and jetties.

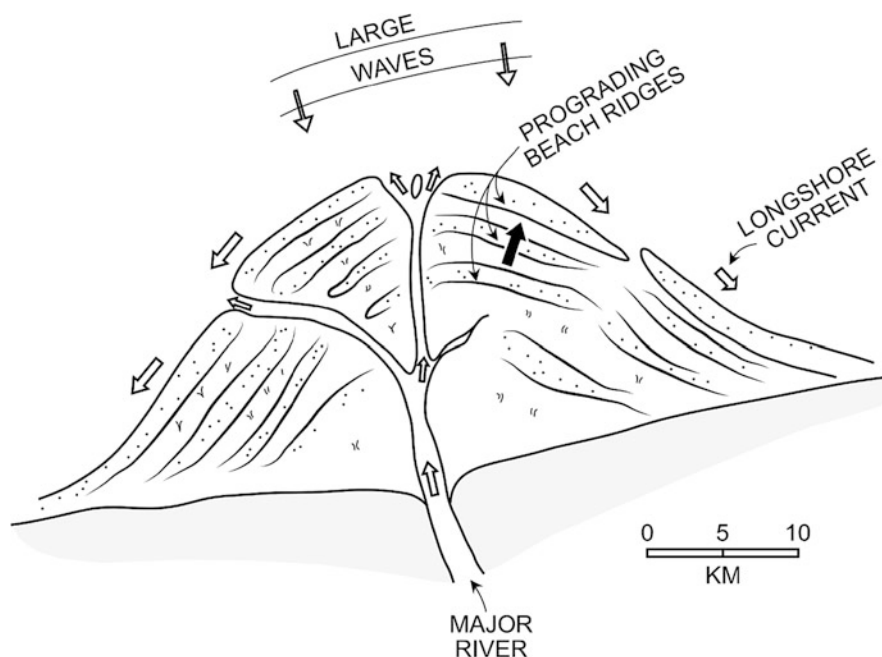
Retrograding barrier islands are composed of coalescing washover fans and terraces that are overtopped at high tides, usually several times a year. Stratigraphically, a relatively thin wedge of sand and shell from the washover terrace overlies back-barrier sediments, which are typically composed of muddy sediments deposited in the lagoons behind the islands. Prograding barrier islands are composed of multiple beach ridges. Stratigraphically, they consist of sand 8–10 m thick which has prograded over offshore muds.

Wave-Dominated Deltas

On wave-dominated coasts, where the rivers have enough sediment load to have filled the antecedent lowstand valley and build a bulge in the shore (within the time frame of the

Wave-Dominated Coasts,

Fig. 1 Schematic sketch in plan view of a wave-dominated delta. Note dominance of prograding beach ridges throughout the delta plain surface



Wave-Dominated Coasts,

Fig. 2 Depositional patterns associated with wave-dominated river mouths: (a) normal wave incidence; (b) oblique wave incidence (after Wright 1977)

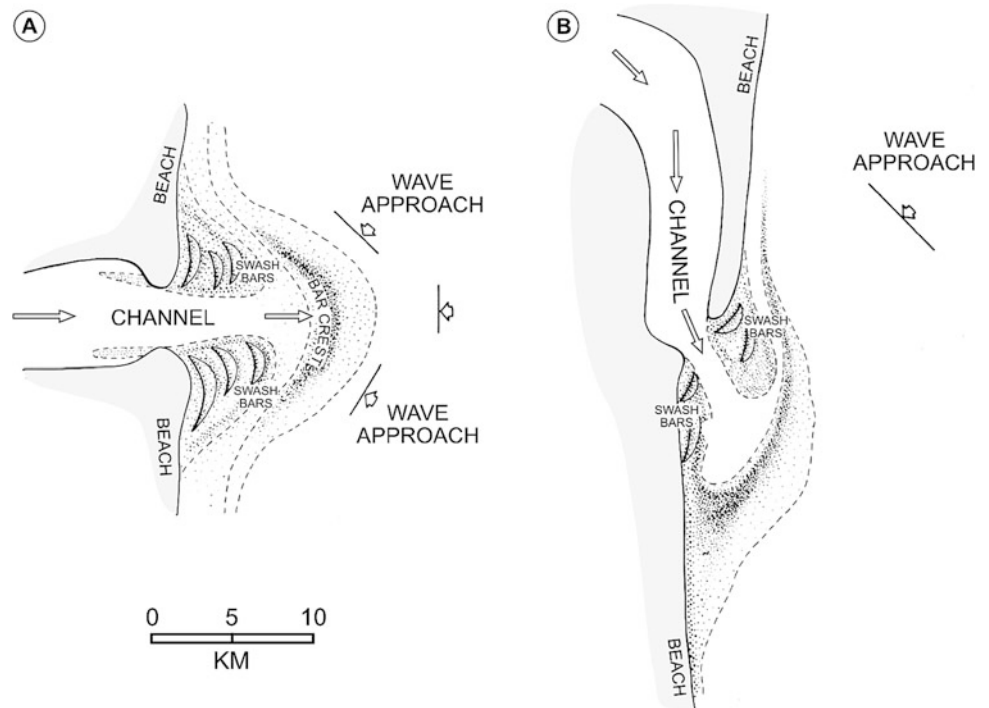
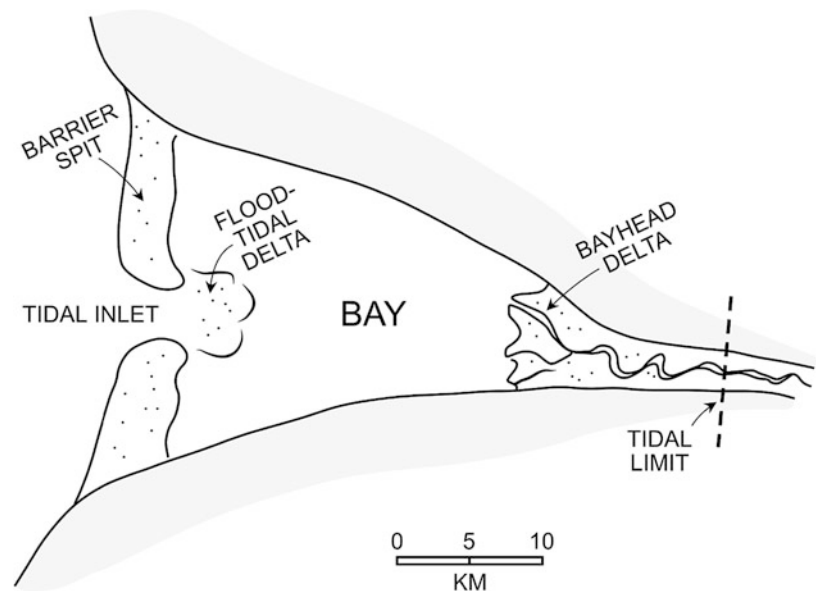
**Wave-Dominated Coasts,**

Fig. 3 Schematic sketch in plan view of a wave-dominated estuary as described by Dalrymple et al. (1992)



present highstand), the resulting river deltas are typically referred to as *wave-dominated deltas* (Fisher et al. 1969; Galloway 1975). A generalized model of a wave-dominated delta is given in Fig. 1. Note the arcuate shape of the delta, which is comprised mostly of prograded beach ridges. The prograded sand sheet associated with such deltas is typically thick because of the massive shorefaces that are created by the large waves.

The depositional patterns of sand at wave-dominated river mouths is illustrated in Fig. 2 (modified after Wright 1977).

On river mouths where waves approach the shore with a parallel orientation (normal wave incidence), the crest of the submerged river-mouth bar is arcuate in shape and projects straight offshore with the intertidal portion of the bar being covered with swash bars that migrate more or less straight onto the shore. With oblique wave incidence, the river mouth, as well as the submerged river-mouth bar, is diverted in a downdrift direction.

The São Francisco delta on the coast of southern Brazil, which bears a close resemblance to the general model given in

Fig. 1, is a spectacular example of a wave-dominated delta. The average tidal range is 1.9 m, which is on the border between micro- and mesotidal conditions. However, the offshore slope is steep, and waves are exceptionally large for deltaic shores. According to Coleman and Wright (1975, p. 131), more wave energy is expended in 10 h on the São Francisco delta front than in 365 days on the Mississippi delta shore. The result is a smooth-faced delta plain dominated by beach ridges and coastal dunes. The composite stratigraphic column for this delta shows a thick sequence of sandy beach, dune, and upper shoreface sediments in the upper half of the stratigraphic column.

Wave-Dominated Estuaries

The concept of tide and wave dominance has also been applied to estuaries by Dalrymple et al. (1992). However, they state that estuaries are unlike many other coastal systems, because they are “geologically ephemeral.” If the rate of sediment supply is sufficient to eventually fill the low-stand valley within which the estuary is located, the filled valley then becomes a delta. The estuaries they term wave-dominated are composed of a sand body complex (barrier/tidal inlet) at the entrance, a muddy central basin, and a bayhead delta system. Their general model for this type of estuary is given in Fig. 3.

Cross-References

- [Barrier Island Landforms](#)
- [Bars](#)
- [Deltas](#)
- [Estuaries](#)
- [Tidal Inlets](#)
- [Tide-Dominated Coasts](#)
- [Wave-Tide-Dominated Coasts](#)

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Waves

Gerhard Masselink

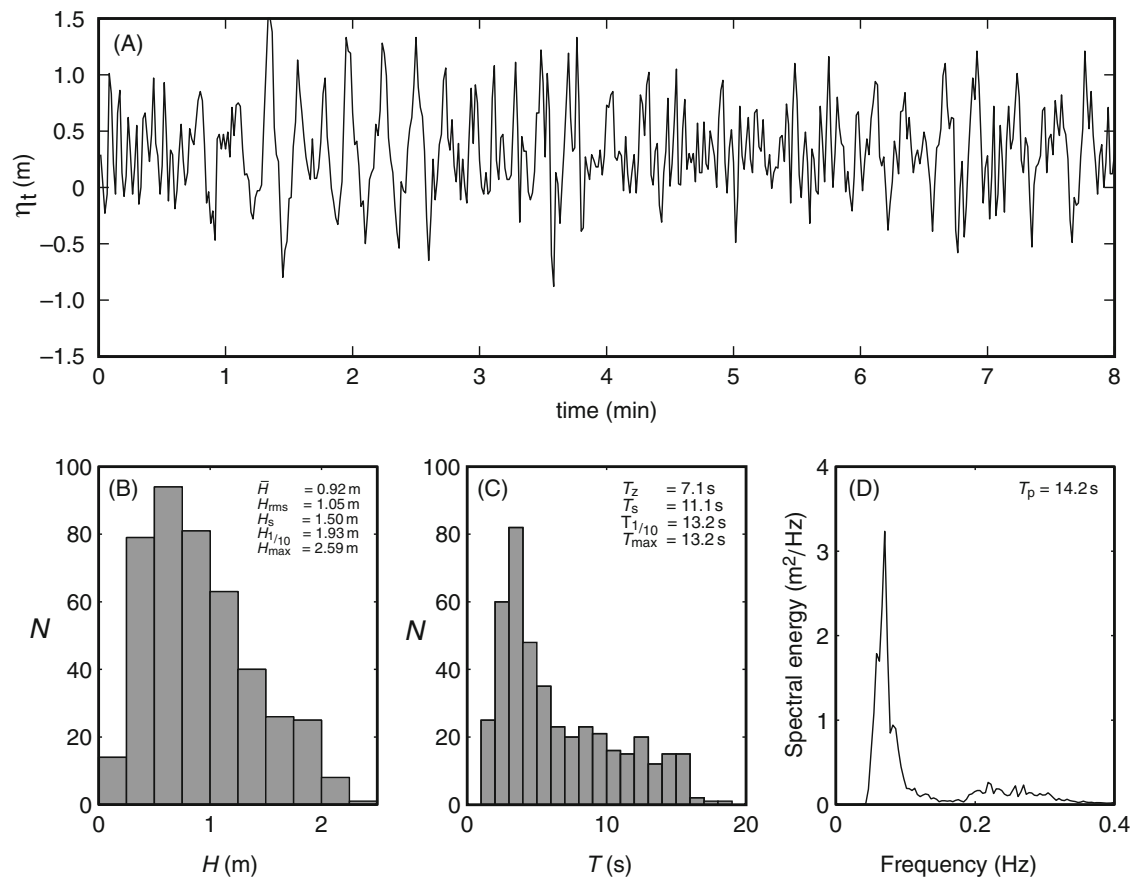
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In most coastal regions, waves provide the dominant source of energy to the nearshore. Although part of the incident wave energy is converted into heat or sound, or is reflected at the shore, a significant proportion of the energy is used for sediment transport and ensuing morphological change. An understanding of wave processes is therefore of fundamental importance when investigating nearshore sediment transport processes and morphology.

The characteristics of natural waves have received much attention in the oceanographic and coastal engineering literature. The classic book by Kinsman (1965) summarizes much of what was known about ocean waves in the early 1960s and remains authoritative. In the last three decades, advances in instrument design and data handling have considerably improved our knowledge of natural wave processes as described in Massel (1996). A succinct, but comprehensive overview of our current understanding of wave processes is provided by Komar (1998).

Wave Analysis and Statistics

An important characteristic of natural waves is that they are often highly irregular and made up of a range of wave heights and periods (Fig. 1a). Hence, statistical techniques are required to properly describe the sea state in quantitative terms. Wave-by-wave analysis is one of the most common approaches to describe irregular waves. This technique consists of identifying the individual waves in the wave record using the zero-downcrossing method (IAHR 1989) and determining representative wave parameters from the subset of wave heights and periods (Fig. 1b, c). An alternative method to describe the properties of irregular waves is spectral analysis (Fig. 1d). The wave spectrum plots wave

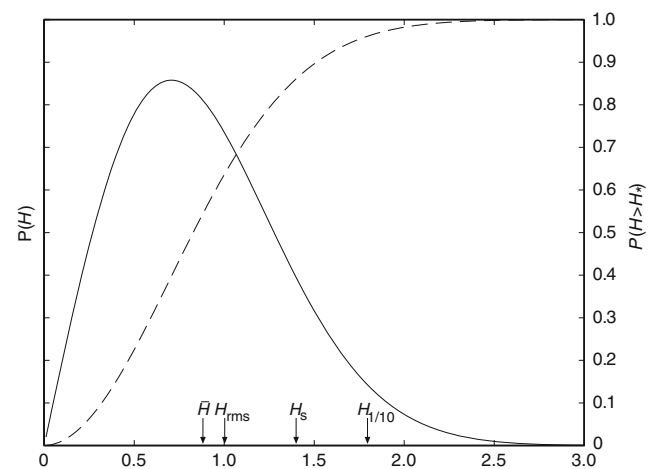


Waves, Fig. 1 Analysis of 1 h of irregular offshore wave data: (a) time series of 8-min section of demeaned water surface elevation η_t ; (b) frequency distribution of wave heights H ; (c) frequency distribution of wave periods T ; and (d) wave spectrum. The wave field represents a

combination of wind waves and swell as indicated by the bimodal wave spectrum. The data were collected in 48 m water depth off the coast of Perth, Western Australia

energy (variance) as a function of frequency (inverse of period) and can readily be used to identify the dominant wave frequencies in the wave record. In addition, the wave spectrum is useful for determining the partitioning of wave energy over distinct frequency bands (e.g., wind and swell waves). The directional wave spectrum can also be computed and plots spectral energy as a function of wave direction and frequency.

The two most common wave parameters to describe an irregular wave field are the significant wave height H_s and the peak wave period T_p . The significant wave height is sometimes denoted by $H_{1/3}$ and represents the average wave height of one-third of the highest waves in a wave record. The significant wave height approximately corresponds to visual estimates of wave heights. The peak period is derived from the wave spectrum and is defined as the wave period associated with the maximum wave energy. Other frequently used wave parameters to be derived using wave-by-wave analysis include $\bar{H}$ (mean wave height), H_{rms} (root mean square wave height), $H_{1/10}$ (average wave height of one-tenth of the highest waves), H_{max} (maximum wave



Waves, Fig. 2 Relative (solid line; Eq. 1) and cumulative (dashed line; Eq. 2) probability density functions for wave heights according to the Rayleigh distribution for a root mean square wave height of $H_{rms} = 1$ m

height), T_z (mean wave period), T_s (mean wave period associated with H_s), $T_{1/10}$ (mean wave period associated with $H_{1/10}$) and $T_{\max}$ (maximum wave period associated with $H_{\max}$).

Longuet-Higgins (1952) suggested that the probability distribution of wave heights in an irregular wave field can be described by the Rayleigh distribution (Fig. 2)

$$P(H) = \frac{2H}{H_{\text{rms}}^2} \exp \left[-\left(\frac{H}{H_{\text{rms}}} \right)^2 \right]. \quad (1)$$

According to Eq. 1, the probability that a particular wave height has a value in the range $H + \delta H$ is given by the product $P(H)\delta H$. Thus, the probability of occurrence in a certain class range increases for increasing width of the class ranges. The probability that a particular wave height exceeds a prescribed value H_* is given by

$$P(H > H_*) = \exp \left[-\left(\frac{H_*}{H_{\text{rms}}} \right)^2 \right]. \quad (2)$$

Based on the Rayleigh distribution, various wave height measures can be determined from the standard deviation of the water surface elevation record σ_η

$$\bar{H} = 2.505\sigma_\eta \quad (3a)$$

$$H_{\text{rms}} = 2.828\sigma_\eta \quad (3b)$$

$$H_s = 4.004\sigma_\eta \quad (3c)$$

$$H_{1/10} = 5.090\sigma_\eta. \quad (3d)$$

Equation 3a–d enables direct determination of representative wave height parameters from the wave record (through the standard deviation) without the need to conduct wave-by-wave analysis.

The periods of individual waves generally show a narrower distribution than that of the wave heights and are in the range of 0.5–2 times the mean wave period T_z . The smallest waves in a record often have the shortest wave periods and site-specific correlations between the wave height and period can be derived empirically. For example, wave records from the North Sea suggest that $T \approx 4H^{0.4}$. One of the main problems with determining the characteristic wave period of a wave field is that natural waves generally consist of more than one wave field (refer to Fig. 1), rendering a single wave period statistic not very useful.

An important application of statistical analysis of wave conditions is the estimation of extreme wave conditions on

the basis of short-term wave records. This involves choosing a suitable probability distribution to fit to the available data and then extrapolating to obtain the probability of occurrence of extreme wave conditions. The probability distribution that is commonly used for this is a lognormal, Gumbel or Weibull distribution (Massel 1996).

Wave Generation

Waves are generated by wind acting on a water surface. In physical terms, the formation of waves constitutes a transfer of energy from wind to waves. At present, the generally accepted theory to account for the growth of wind waves is the combined Miles–Phillips mechanism. The mechanism incorporates two distinct phases of energy transfer from wind to waves: (1) an initial, linear growth phase that accounts for the formation of waves on a calm water surface (Phillips 1957); and (2) an ongoing, exponential growth of the waves due the interactive coupling between wind and waves (Miles 1957). An additional process that occurs within a developing wave field is the transfer of energy from high to low frequencies (Stewart 1967; Longuet-Higgins 1969). This process may occur when short-period waves steepen and break on the crests of the long-period waves. Since the long-period waves are traveling faster in deepwater than the short-period waves, they sweep up the energy and momentum contained within the short-period waves that are being overtaken. Thus, in a developing wave field both the wave energy level and the dominant wave period progressively increase. Wave growth is not infinite and when waves reach their limiting steepness ($H/L = 1/7$ in deepwater, where H is wave height and L is wavelength) they break in the form of white caps. An equilibrium can eventually be achieved whereby the energy losses by wave breaking are balanced by the addition of new energy being transferred from the wind to the waves. Such an equilibrium wave field is referred to as a fully arisen sea.

Wave conditions can be predicted as a function of wind speed, wind duration, and fetch length. This technique is known as wave forecasting when predicted wind data are used, or wave hindcasting when historical wind data are used. There are two main approaches to wave prediction: (1) the significant wave approach; and (2) the wave spectrum approach. The former approach is relatively simple and aims to predict wave statistics such as significant wave height and spectral wave period from wind conditions (CERC 1984). The latter approach is more sophisticated and characterizes the predicted wave field by their spectra. One of the most common spectral formulations to parameterize the growth of a wave field is the JONSWAP spectrum which is based on extensive field measurements conducted in the North Sea (Hasselmann et al. 1976).

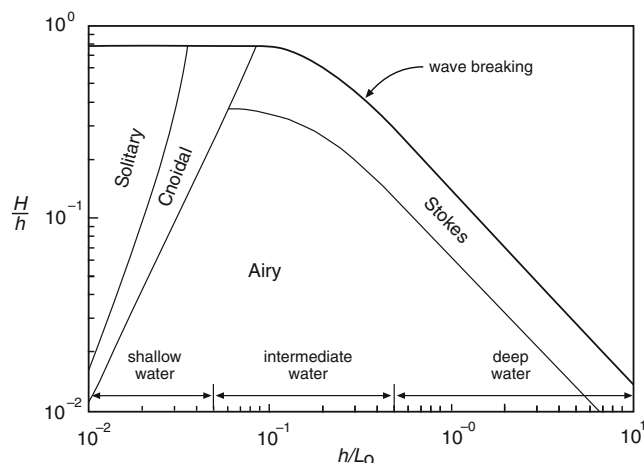
Wave Theories

Wave theories are mathematical formulations that predict the change in wave properties, such as water particle velocity, wave height, and wave energy, with depth. A basic assumption of linear and nonlinear wave theory is irrotational flow, that is, bed friction is neglected and there are no internal shear stresses. Under this assumption, the full momentum equations, the Navier–Stokes equations, reduce to the Euler equation of motion that contains only terms for the water particle velocities and local pressures. A second relation that must be included in the analyses of wave motion is the continuity equation to ensure conservation of mass. To provide an analytical solution to these equations, additional simplifying assumptions are required. The greater the number of assumptions, the simpler the resulting equations, but the greater the departure from the actual wave motion one is trying to describe. A large number of wave theories have been developed with different levels of complexity and accuracy.

Airy (or linear) wave theory (Airy 1845) gives the least complicated expressions for wave motion and is therefore the most widely used wave theory. The simplicity of Airy wave theory is attained by making a number of assumptions. Most significantly, it is assumed that the wave amplitude is negligibly small compared with water depth. This is not a problem for waves in deepwater, but the theory is less suitable for waves in shallow water. Stokes wave theory (Stokes 1847) was developed to overcome this main limitation of Airy wave theory and is a theory for waves of finite height (i.e., wave height is not negligible). Stokes wave theory is essentially Airy wave theory with nonlinear terms added to the equations. The additional terms have the effect of enhancing the crest amplitude and subtracting from the trough amplitude. The resulting wave profile is characterized by narrow wave

crests and wide wave troughs which provides a more realistic wave shape of waves as they enter shallow water. When waves are in very shallow water, their crests peak up and are separated by wide, flat troughs, very much resembling solitary waves. Although solitary waves are not oscillatory waves and do not have a wave length or period, solitary wave theory (Boussinesq 1872) can be used to determine properties of waves very close to breaking. The most sophisticated wave theory available is cnoidal wave theory (Korteweg and de Vries 1895) and has potentially the widest range of application. In deepwater, the cnoidal wave solution reduces to the Airy wave theory, whereas in shallow water the wave period and length become infinite such that the cnoidal waves become equivalent to solitary waves. Cnoidal wave theory provides the most accurate description of the wave profile but is also the most cumbersome to use. A final wave theory that deserves mention is stream function wave theory (Dean 1965) which resolves water motions associated with waves numerically without the need for simplifying assumptions required to obtain analytical solutions.

Selection of the appropriate wave theory depends upon the intended application. Figure 3 suggests the fields of application in terms of the ratios H/h and h/L . In the construction of this graph, the widest possible regions are given to the simpler wave theories. For example, the difficult cnoidal wave theory is given only a restricted region of application within its potentially much greater field. An important consideration is that Airy wave theory predicts water motions that are symmetrical, that is, onshore velocities are identical to offshore velocities. The nonlinear wave theories, on the other hand, predict water flows that are distinctly asymmetric in shallow water depths. This flow asymmetry is consistently directed landward whereby the onshore stroke of the wave motion is stronger, but of shorter duration than the offshore stroke. An accurate description of the flow asymmetry under waves is of fundamental importance for modeling wave-driven sediment transport processes (see “► [Cross-Shore Sediment Transport](#)”).



Waves, Fig. 3 The areas of applications of the various wave theories as a function of the ratios H/h and h/L_0 , where preference is given to the simpler theories such as Airy wave theory (after Komar 1998)

Airy Wave Theory

Airy wave theory is still the most widely used wave theory in oceanographic research and important wave processes such as shoaling and refraction are adequately described by treating ocean waves as Airy waves. According to Airy wave theory the water surface elevation $\eta(x, t)$ is described by

$$\eta(x, t) = \frac{H}{2} \cos(kx - \sigma t) \quad (4)$$

where x is the coordinate axis in the direction of wave advance, t is time. H is the wave height, $k = 2\pi/L$ is the wave number (L is the wave length), and $\sigma = 2\pi/T$ the radian

frequency (T is the wave period). An important relationship to emerge from Airy wave theory is the dispersion equation which expresses the functional relationship between the wave period and the wavelength,

$$\sigma^2 = gk \tanh(kh), \quad (5)$$

which can be rewritten as

$$L = \frac{g}{2\pi} T^2 \tanh\left(\frac{2\pi h}{L}\right), \quad (6) \quad \text{and}$$

where h is the water depth and g is the gravitational acceleration. From Eq. 6 the wave celerity C can be derived:

$$C = \frac{L}{T} = \frac{g}{2\pi} T \tanh\left(\frac{2\pi h}{L}\right). \quad (7)$$

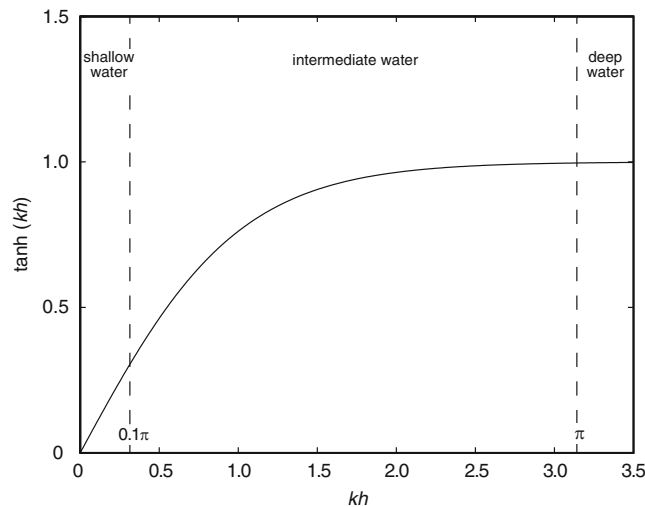
The wave celerity is also referred to as the wave phase velocity and represents the propagation speed of individual waves.

The dispersion equation can not be solved explicitly because it contains L on either side of the equation and therefore has to be solved numerically (iteratively). However, a convenient function which is accurate to 0.1% is given by Hunt (1979).

$$(kh)^2 = y^2 + \frac{y}{1 + 0.666y + 0.355y^2 + 0.161y^3 + 0.0632y^4 + 0.0218y^5 + 0.00654y^6} \quad (8)$$

where

$$y = \frac{\sigma^2 h}{g} = \frac{4\pi^2 h}{gT^2} = 4.03 \frac{h}{T^2}. \quad (9)$$



Waves, Fig. 4 Properties of the $\tanh$ function

At present, the use of computers makes computations involving the general dispersion relationship a relatively painless task, especially if Eqs. 8 and 9 are used. In the past, however, further simplifications to the dispersion relationship were desirable and these were made possible due to the characteristics of the $\tanh$ function (Fig. 4). If $kh = 2\pi h/L$ approaches zero, $\tanh(kh) \approx kh$ and Eqs. 6 and 7 reduce to

$$L_s = T\sqrt{gh} \quad (10)$$

$$C_s = \sqrt{gh}. \quad (11)$$

Equations 10 and 11 are known as the shallow-water approximations, hence the use of the subscript “s,” and are valid for $kh < 0.1\pi$ (or $h/L < 0.05$). If, on the other hand, $kh = 2\pi h/L$ becomes large, $\tanh(kh) \approx 1$ and Eqs. 6 and 7 reduce to

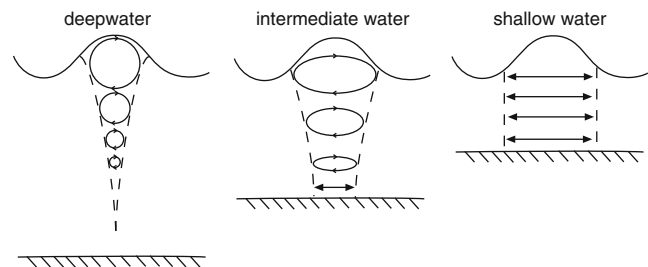
$$L_o = \frac{gT^2}{2\pi} \quad (12)$$

and

$$C_o = \frac{gT}{2\pi}. \quad (13)$$

Equations 12 and 13 are known as the deepwater approximations and are valid for $kh > \pi$ (or $h/L > 0.5$). Commonly, the subscript “o” is used to denote deepwater conditions and is therefore used here. Intermediate depth conditions prevail for $0.05 < h/L < 0.5$ and here the general Eqs. 6 and 7 should be used because the errors in the deepwater and shallow-water approximations will exceed 5%.

Waves have potential and kinetic energy associated with the deformation of the water surface and the orbital motion of the water particles, respectively. According to Airy wave theory, the two energy forms are equal and the energy density E (in Newtons per unit area) is given by



Waves, Fig. 5 Description of the orbital motion of Airy waves in deep, intermediate, and shallow water

$$E = \frac{1}{8} \rho g H^2, \quad (14)$$

where ρ is the density of water. The rate at which the energy density is carried along by the moving waves is the wave energy flux P and is given by

$$P = ECn \quad (15)$$

where

$$n = \frac{1}{2} \left[1 + \frac{2kh}{\sinh(2kh)} \right]. \quad (16)$$

The velocity Cn is the speed at which the energy density is carried along and is referred to as the group velocity C_g since it is the movement of groups of waves as distinguished from individual waves that travel according to the wave celerity C . In deepwater $n \approx 1/2$ but n increases in value as the waves travel into water of intermediate depth, becoming $n = 1$ in shallow water. The implication is that in deepwater, individual waves travel at twice the speed as the wave groups, whereas in shallow water, individual waves propagate at the same speed as the wave groups.

Airy wave theory also provides a description of the movement of water particles associated with wave motion (Fig. 5). In deepwater, water particles move in a circular path with the radius of the circles decreasing with increasing depth beneath the water surface. The wave motion of deepwater waves is not felt at the seabed. In intermediate water depths the water particles follow an elliptical path with the ellipses becoming flatter and smaller as the seabed is approached. At the bottom, the water particles undergo a to-and-fro motion with a horizontal extent of the water motion d_0 .

$$d_0 = \frac{H}{\sinh(kh)} \quad (17)$$

and a maximum near-bed velocity u_m

$$u_m = \frac{\pi H}{T \sinh(kh)}. \quad (18)$$

In shallow water, Airy wave theory predicts that all water motions consist of to-and-fro horizontal movements uniform with depth with

$$d_0 = \frac{HT}{2\pi} \sqrt{\frac{g}{h}} = \frac{H}{kh} \quad (19)$$

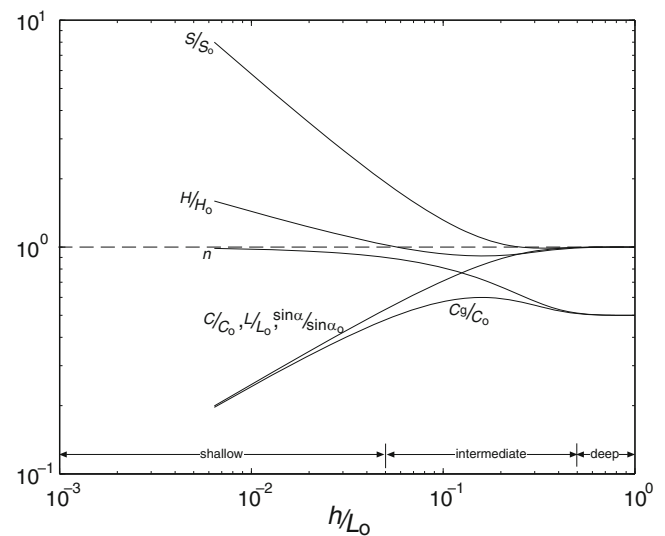
and

$$u_m = \frac{H}{T} \sqrt{\frac{g}{h}}. \quad (20)$$

Parameters d_0 and u_m are important with respect to the bed morphology (wave ripples) and sediment transport processes. It should be emphasized that nonlinear wave theories such as Stokes theory provide a better description of water particle motions in shallow water than Airy wave theory.

Wave Dispersion

A developing wave field is characterized by a broad-banded wave spectrum, indicating that a large range of wave periods are represented in the wave field. Such waves are referred to as sea. According to Airy wave theory, the wave celerity in deepwater C_0 increases with wave period T (Eq. 13). In a broad-banded wave field, therefore, waves propagate at a range of wave celerities with the longer-period waves traveling faster than the short-period waves. Given sufficient time, the longer-period waves will outrun and leave behind the shorter-period waves. This sorting of the waves by period is termed wave dispersion and results in the transformation of broad-banded and confused sea into regular swell. The longer the distance of travel from the area of wave generation, the more effective the wave sorting process and the narrower the wave spectrum becomes. If a wave field is generated by an offshore storm, wave dispersion causes the longest-period waves within this storm-wave field to reach the coast first. These first waves, the so-called storm-forerunners, will then be followed by waves of progressively decreasing shorter periods.



Waves, Fig. 6 Shoaling transformations for Airy waves as a function of the ratio h/L_0

Wave Groups

Ocean waves often occur in successive groups of higher or lower waves. Using linear wave theory, Longuet-Higgins and Stewart (1964) demonstrated that a small lowering of the mean water level accompanies groups of high waves, whereas a small rise in mean water level occurs under groups of small waves. This leads to the development of the group-bound long wave which has the same wave length and period as the incident wave group, but is 180° out of phase with the group. Offshore-directed currents of the long wave are maximum under groups of high waves and may constitute an important mechanism of offshore transport under shoaling waves (see “► Shelf Processes”). Wave groups and the group-bound long wave are also of interest through their link with infragravity wave motion in the surf zone (see “► Surf Zone Processes”).

Wave Shoaling

As waves propagate from deep to shallow water they undergo a number of transformations which can be predicted using Airy wave theory. Some of these variations in wave properties have been plotted in Fig. 6 and indicate that both the wavelength L and the wave celerity C systematically decrease with decreasing water depth. The wave period is the only variable that remains constant. Similarly, there are variations in n and hence changes in the wave group velocity C_g as the depth decreases.

The variation in the heights of the shoaling waves can be calculated from a consideration of the energy flux. Assuming energy losses due to bed friction and reflection can be ignored, the wave energy flux $P = ECn$ remains constant during wave propagation. This can be expressed as:

$$P = E_2 C_2 n_2 = E_1 C_1 n_1 = \text{constant}, \quad (21)$$

where the subscripts “1” and “2” indicate two different locations along the path of wave travel ($h_1 > h_2$). Inserting $E = 1/8 \rho g H^2$ in Eq. 21 yields

$$H_2 = \left(\frac{C_1 n_1}{C_2 n_2} \right)^{0.5} H_1. \quad (22)$$

The ratio of the local wave height H (i.e., H_2) to the deepwater wave height H_o (i.e., H_1) can easily be derived from Eq. 22

$$\frac{H}{H_o} = \left(\frac{1}{2n} \frac{C_o}{C} \right)^{0.5} = K_s, \quad (23)$$

where K_s is referred to as the shoaling coefficient. Equation 23 is plotted in Fig. 6 and it can be seen that during wave shoaling, the wave height initially decreases slightly while entering intermediate water depth followed by a rapid increase.

The wave steepness $S = H/L$ also varies in the shoaling waves. Because the increase in H during shoaling is accompanied by a decrease in L , the wave steepness shows a dramatic increase during shoaling, especially over the shallow water region (Fig. 6). It is the steepening of the waves that ultimately results in wave breaking.

Wave Refraction

When a wave approaches the coast with its crest at an angle to the bottom contours, the water depth will vary along the wave crest. If the wave is in intermediate or shallow water, the wave celerity C will also vary along the wave crest with the deeper part of the wave propagating at a faster rate than the shallow part of the wave (Eqs. 7 and 11). This results in a rotation of the wave crest with respect to the bottom contours, or in other words a bending of the wave rays or wave orthogonals. This process is known as wave refraction (see “► Wave Refraction Diagrams”) and is of great relevance to nearshore currents, sediment transport, and morphology (see “► Beach Processes,” “► Headland-Bay Beach,” “► Longshore Sediment Transport,” and “► Surf Zone Processes”).

The refraction of waves is similar to the bending of light rays and the change in direction is related to the change in the wave celerity C through the same Snell's law

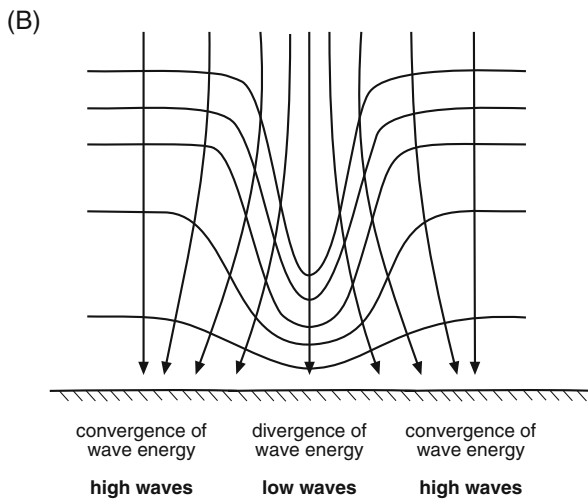
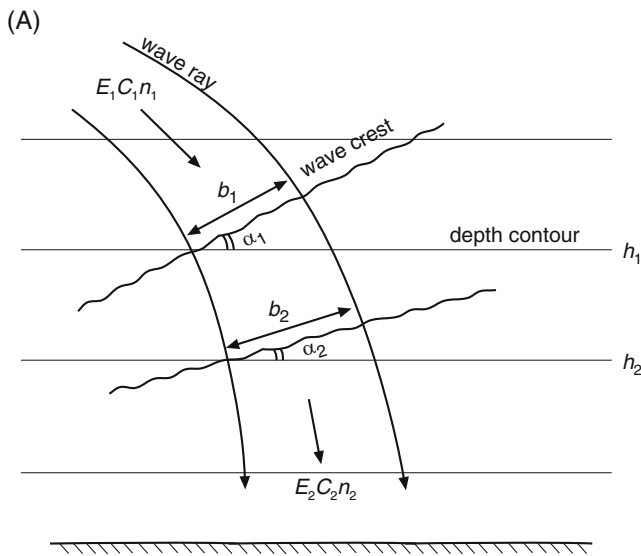
$$\frac{\sin \alpha_1}{C_1} = \frac{\sin \alpha_2}{C_2} = \text{constant}, \quad (24)$$

where α refers to the angle between the wave crest and the bottom contours and the subscripts “1” and “2” are used to indicate two different locations along the path of wave travel ($h_1 > h_2$) (Fig. 7a). For a straight coast with parallel bottom contours, the angle at a given depth can be related to the angle of wave approach of the wave in deepwater

$$\sin \alpha = \frac{C}{C_o} \sin \alpha_o. \quad (25)$$

The expression shows that as the wave celerity decreases in shallow water, the angle made by the wave with the bottom contour also decreases.

The refractive bending of the wave rays also causes the wave rays to spread out, that is, the distance between rays increases as the waves are being refracted. If b_1 and b_2 are the spacing of the rays at two consecutive depths then the energy



Waves, Fig. 7 Wave refraction in case of: (a) parallel bottom contours; and (b) complex bathymetry

flux between the wave rays at these two depths should be constant:

$$P = E_1 C_1 n_1 b_1 = E_2 C_2 n_2 b_2 = \text{constant}. \quad (26)$$

Inserting $E = 1/8 \rho g H^2$ in Eq. 26 then yields

$$H_2 = \left(\frac{n_1 C_1}{n_2 C_2} \right)^{0.5} \left(\frac{b_1}{b_2} \right)^{0.5} H_1. \quad (27)$$

For straight coasts with parallel contours, simple geometry considerations give

$$\frac{b_1}{b_2} = \frac{\cos \alpha_1}{\cos \alpha_2}. \quad (28)$$

The ratio of the local wave height H to the deepwater wave height H_o can easily be derived from Eq. 27

$$\frac{H}{H_o} = \left(\frac{1}{2n} \frac{C_o}{C} \right)^{0.5} \left(\frac{b_o}{b} \right)^{0.5} = K_s K_r, \quad (29)$$

where K_r is referred to as the refraction coefficient.

Irregular bottom topography can cause waves to be refracted in a complex way and produce significant variations in wave height and energies along the coast (Fig. 7b). Waves diverge over relatively deepwater (e.g., depression in the seafloor), resulting in a spreading of the wave rays ($K_r < 1$) and a decrease in wave energy and wave height. In contrast, wave rays converge over relatively shallow water (e.g., shoal on the seafloor), resulting in a decrease in the spacing of the wave rays ($K_r > 1$) and an increase in wave energy and wave height.

Waves, Fig. 8 Plunging breaker



Wave Diffraction

Wave diffraction is the process of wave energy transfer in a direction other than the wave propagation direction and occurs irrespective of water depth. As such, diffraction is fundamentally different from wave refraction, although both processes often operate in concert. Wave diffraction occurs when an otherwise regular train of waves encounters an impermeable structure, for example, an island. A wave shadow zone will be created behind the obstacle, but wave diffraction will cause wave energy to leak into the shadow zone. Wave diffraction also enables wave energy to enter into narrowly confined bays and harbors.

Wave Breaking

At some point during wave transformation, the water depth becomes too shallow for a stable waveform to exist and the waves break (Fig. 8). The mechanism of wave breaking is that the horizontal velocities of the water particles in the wave crest exceed the phase velocity of the wave. Consequently, the water particles leave the waveform, resulting in a disintegration of the wave into bubbles and foam. The breaking wave height is related to the water depth by the breaker index γ according to

$$H_b = \gamma h_b, \quad (30)$$

where H_b is the height of the breaking wave and h_b is the mean water depth at the point of breaking. When the height-to-depth ratio of the wave exceeds the breaker index, breaking will occur. Solitary wave theory prescribes that $\gamma = 0.78$ and this provides a good first-order approximation.

Komar and Gaughan (1972) obtained a useful expression of the breaker height as a function of the deepwater wave height and steepness by applying linear wave theory (and ignoring energy losses by bed friction) to evaluate the wave energy flux from deepwater up to the breakpoint and using $\gamma = 0.78$,

$$\frac{H_b}{H_o} = \frac{0.563}{(H_o/L_o)^{1/5}}. \quad (31)$$

The value 0.563 was determined empirically from laboratory and field data. Due to wave shoaling, the height of the breakers always exceeds that of the deepwater waves and the ratio H_b/H_o increases with decreasing wave steepness H_o/L_o .

A continuum of breaker shapes occur in nature. However, three main types of breakers are commonly recognized:

spilling, plunging, and surging (Galvin 1968). Spilling breakers are characterized by a gradual peaking of the wave until the crest becomes unstable, resulting in a gentle forward spilling of the crest. Plunging breakers are distinguished by the shoreward face of the wave becoming vertical, curling over, and plunging forward and downward as an intact mass of water. In surging breakers, the front face and crest of the wave remain relatively smooth and the wave slides directly up the beach without breaking. Spilling and plunging breakers occur in the surf zone, whereas surging breakers are restricted to the swash zone. The transition from one breaker type to another is gradual and in natural surf zones, a mixture of breaker types is often present.

The breaker type principally depends on the beach gradient and wave steepness, which may be expressed by the surf similarity parameter

$$\xi_b = \frac{\tan \beta}{\sqrt{H_b/L_o}}, \quad (32)$$

where $\tan \beta$ is the beach gradient and the subscripts “b” and “o” indicate breaker and deepwater conditions, respectively. Spilling breakers occur for $\xi_b < 0.4$, plunging breakers occur when $\xi_b = 0.4$ –2 and surging breakers occur for $\xi_b > 2$ (Battjes 1974).

Cross-References

- Beach Processes
- Cross-Shore Sediment Transport
- Headland-Bay Beach
- Longshore Sediment Transport
- Nearshore Wave Measurement
- Surf Zone Processes
- Wave Climate
- Wave Environments
- Wave Focusing
- Wave Hindcasting
- Wave Refraction Diagrams

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Wave-Tide-Dominated Coasts

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Definition

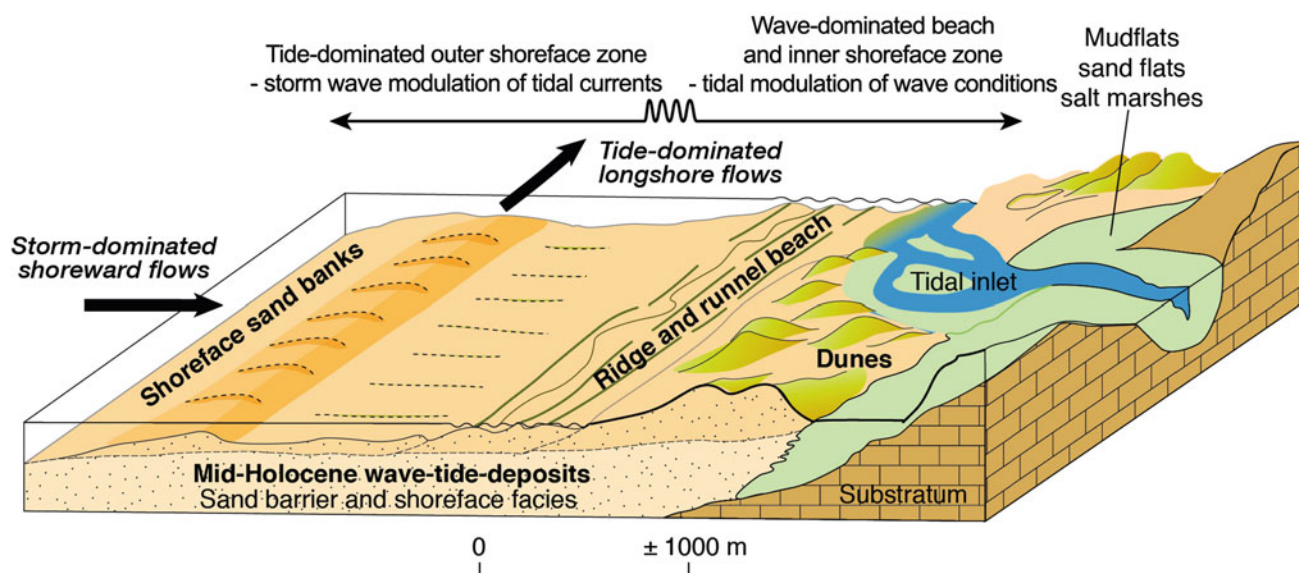
Wave-tide-dominated coasts may be defined as coasts the morphology of which is shaped by short- to long-term background hydrodynamic conditions generated jointly by waves and tides. Few of the world's coasts are devoid of wave action (Davis and Hayes 1984), such that waves are generally considered as the dominant source of processes of coastal change. From first principles, tidal forces are felt universally on all coasts, although their effect is modulated by position. However, waves are variable in time and space. As a result, waves

might mute, modulate, obliterate, amplify, or dominate the tidal signal. Coasts where waves are absent or where waves obliterate the tidal signal are the extremes previously recognized as tidal- or wave-dominated coasts (Anthony and Orford 2002). Inevitably, the rest must be mixed wave-tide-dominated coasts. Between the wave-dominated coast and the tide-dominated coast therefore exists a wide range of mixed wave-tide-dominated coastal types that comprises a considerable proportion, if not the majority of the world's coastline.

Wave-Tide-Dominated Coastal Settings and Characteristics

Understanding how tides and waves interact is essential to many areas of research and resource management, including marine renewable energy, sediment transport, long-term seabed morphodynamics, storm surges, and the impacts of climate change (Hashemi et al. 2015). This interaction is also fundamental to the stability and management of many coasts. The importance of tides in determining coastal processes and morphology was long overlooked, mainly because of lack of studies that properly evaluate their role, notably in settings deemed as wave-dominated. As far as beaches are concerned, by far the largest number of process studies has concerned areas of low tidal range where the tidal process signature is weak or negligible. The importance of tidal modulation of the hydrodynamics, sediment transport patterns, and resultant morphology of beaches was emphasized in a number of pioneer studies (Wright et al. 1982; Short 1991; Masselink and Short 1993; Levoy et al. 2000). Identification of the joint influence of waves and tides is important in terms of understanding the physical processes involved in coastal change and its implications for coastal management. Any tidal influence in shaping the coastal morphology occurs directly through tidal currents and their asymmetry, through enhancement of tidal currents due to shoreline configurations, as in the case of estuaries and tidal inlets, through tidal interactions with waves, and indirectly through the effects the large vertical and horizontal tidal translations have on wave processes.

Between the wave- and tide-dominated coastal extremes, the broad spectrum of wave-tide-dominated coasts range from settings with high wave energy and perceptible tidal energy associated with a low tidal range to settings with low wave energy and large tidal ranges. Whereas tides show a regular predictable temporal cycle, wave conditions generally exhibit large temporal variability expressed by irregular, but short-term (order of days to weeks), to seasonal variations in energy and period. As a result, the wave-tide-dominated spectrum is not constant in power level and is probably very irregular, characterized by a few key spikes associated with certain coastal types (Anthony and Orford 2002). Examples are the stubby barrier coasts with frequent tidal inlets associated with



Wave-Tide-Dominated Coasts, Fig. 1 Block diagram of a wave-tide-dominated coast showing schematic spatial and intensity variations in wave- and tide-induced processes and deposits. This example is drawn from the shoreface and beach-dune system of the southern North Sea, which is characterized by numerous shoreface sand banks, some of

which are progressively driven alongshore and onshore by storm waves and tidal and wind-induced currents. Eventual welding of a submarine bank to the shore results in the formation of an intertidal sand platform that feeds sand for the accretion of ridge and runnel beaches and eolian dunes

low- to moderate-energy waves and moderate tidal ranges (Hayes 1979), beaches with very large tidal ranges and low- to moderate-wave energy, and shorefaces characterized by sand banks and ridges molded by storm waves and tides (Fig. 1). Further variability occurs in wave-tide-dominated settings where wind forcing plays an additional important role in modulating the power level. Waves and tides contribute together to extreme water levels and flooding and erosion hazards along the coast (Serafin et al. 2017).

In the high wave energy-low tidal range part of the wave-tide-dominated spectrum, strong tidal currents are only generated where local bathymetric/topographic factors lead to tidal strengthening and slackening. A classification of the Australian continental shelf on the basis of predicted sediment threshold exceedance from tidal currents and swell waves has, for instance, been proposed (Porter-Smith et al. 2004), while examples of coastal variations in tidal and wave transport in adjacent environments have been documented (Harris et al. 2002; Ryan et al. 2007). Variations in shoreface gradient and width can be a large-scale control on spatial patterns of wave versus tidal dominance of shoreface dynamics, as in the southern North Sea (Anthony and Orford 2002). Along certain coasts, this situation may be typical of both fixed and migrating tidal inlets linked to estuaries or lagoons. A case of wave-tide-dominated coast that is often overlooked is that of alongshore migrating inlet systems associated with barriers (see entries on “► Tidal Inlets and ► Barrier”). Such migrating inlet systems, especially active in areas with low tidal ranges favorable to barrier development, are maintained by

the tidal flux. The tidal inlets may capture and store significant amounts of sand transported alongshore by wave-induced longshore drift as tidal-inlet fill, flood-tidal delta, and ebb-tidal delta deposits. It is noteworthy that the sediment record of barrier coasts comprising inlets may exhibit important longshore sequences of wave-deposited barrier sand that overlies such tidal deposits (Hayes 1980; Reinson 1984). Another situation where this may occur is in the vicinity of headlands where sand banks may accumulate as a result of tidal eddies (Pingree 1978; Ferentinos and Collins 1980). It must be noted, however, that this headland-tidal current relationship may span the whole spectrum, depending on tidal range and wave-energy conditions. The high wave-energy end of the wave-tide-dominated spectrum may comprise a spatially narrow or limited zone of tidal influence vis-a-vis a spatially wide but temporally fluctuating zone of wave influence.

At the other end of the spectrum are coasts with low net wave energy and strong tidal action. Coasts subject to low wave action are found where fetch or wave-energy conservation conditions are such that significant waves are either not generated regularly or are not energetic enough as to have the dominant impact on the coast. The former situation may be found in protected settings such as large bays and sheltered epicontinental seas. Wave conservation depends on nearshore morphology which largely determines frictional energy dissipation and energy spread through refraction and diffraction. The highly periodic nature of wave generation or extreme wave attenuation may result in open fine/cohesive clastic-

dominated coasts, the process signature and morphology of which are dominantly shaped by tidal currents. However, for this to happen, these currents need to be sufficiently strong to mobilize or entrain sediments. This end of the wave-tide-dominated spectrum may comprise, in contrast to the high wave-energy end, a spatially narrow but widely temporally fluctuating zone of wave influence vis-a-vis a spatially wide zone of tidal influence, the temporal fluctuations of which are regular, being hinged on the lunar cycle. Swell wave attenuation may be expected on coasts fronted by wide, shallow continental shelves that are themselves favorable to tidal range amplification (Clarke and Battisti 1981). This is the typical situation of tide-dominated deltas. Tidal influences in deltas are particularly important where river mouths debouch into shallow embayed settings adjacent to the open ocean, such as the Ganges-Brahmaputra-Meghna (GBM) in the Bay of Bengal, and where deltaic sedimentation over millions of years has resulted in a large shallow shelf that favors

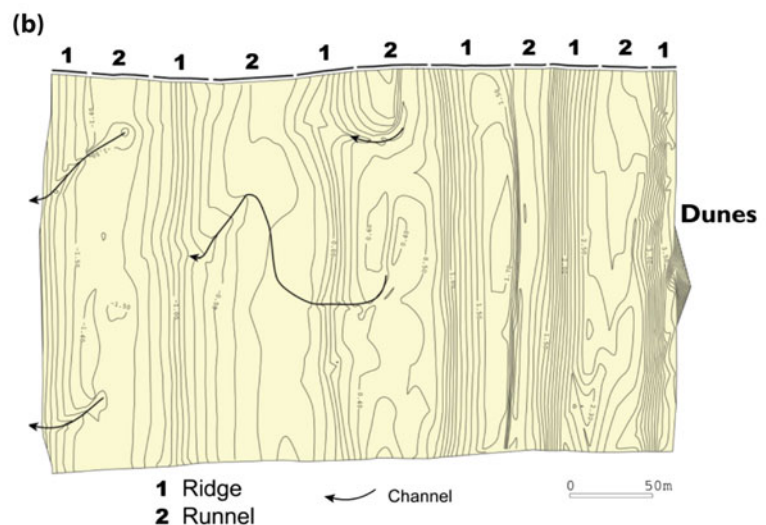
amplification of tides, as off the mouths of the Amazon and the GBM (Anthony 2016).

Schematically, the wave-tide relationship may be presented in terms of a spatial band wherein the tidal influence tends to increase in the offshore zone and the wave influence toward the shore (Fig. 1). As wave energy increases, waves would tend to mute the tidal signal and even dominate. This relationship would, however, vary both spatially and temporally as a function of the temporal fluctuations in wave energy and in tidal power during the lunar cycle.

On modally low wave-energy coasts, especially in protected wave environments, the tidal signal would tend to dominate over a wider area of nearshore zone. This tidal domination would be maintained up to the high-tide shoreline in the periodic absence of waves, or, in areas where attenuation of swell wave energy is complete, up to the high-tide shoreline, or would diminish toward the mid- to high-tide part of the shoreline where wave energy still attains the shoreline.

Wave-Tide-Dominated Coasts,

Fig. 2 Oblique aerial photo (a) and digital elevation model (b) of a typical macrotidal beach (Merlimont, France, eastern English Channel) characterized by a complex intertidal morphology of ridges (bars), runnels (troughs), and drainage channels, in response to a combination of wave- and tide-induced processes



This mixed energy regime may also be expressed by depth variations in protected settings characterized by short-period waves, with wave action being efficient over the shallower coastal and nearshore deposits, and tidal action in the deeper areas.

Wave-tide-dominated coasts may vary from sandy beaches to mudflats and mangrove-colonized tidal flats (see entries on “► Tidal Environments”, “► Tidal Flats”, “► Mangroves, Ecology”, “► Mangroves, Geomorphology”, “► Mangroves, Remote Sensing”, “► Muddy Coasts”, and “► Salt Marsh”). Many deltaic shorelines, such as parts of the Mekong, and areas adjacent to the multiple outlets of the Niger delta, show morphologies, hydrodynamics, and facies typical of mixed wave-tide-domination (Anthony 2015). Within the moderate to high wave- and high tidal-energy end of the spectrum, beaches may exhibit characteristics typical of their counterparts in settings with low tidal ranges. However, tidal modulation of their wave hydrodynamics and sediment transport patterns still occurs, together with significant tidal current activity offshore. Examples of this are the sandy Cable Beach in Western Australia (Wright et al. 1982) and the gravel Loe Bar beach in Cornwall (Almeida et al. 2015). Other examples include beaches on the Aquitaine coast of France. Low- to moderate-energy beaches in settings with very large tidal ranges (above 8 m at mean spring waters) such as those of Normandy, France (Levoy et al. 2000), show even stronger tidal modulation of their morphology and dynamics, essentially through large fluctuations in water level due to the important tidal excursion. Ridge and runnel beaches (Fig. 2) occupy the modally low to moderate wave energy part of the spectrum in settings with short-period waves and moderate to large tidal ranges (Reichmuth and Anthony 2007).

The topography of the nearshore zone or inner shoreface on wave-tide-dominated coasts may be very regular and akin to that of typical wave-dominated coasts exposed to moderate- to high-energy waves, such as the Aquitaine coast. As tidal action increases, this topography may, where abundant loose sediment is available, be characterized by numerous banks and ridges (see entry on “► Off-shore Sand Banks and Linear Sand Ridges”) that are typical of a tide-dominated process signature, such as in the eastern English Channel and southern North Sea, and the Yellow Sea.

Muddy wave-tide-dominated coasts may be considered as being close to the tide-dominated end of the spectrum of coastal types and are generally found in the vicinity of major river mouths with large tidal ranges and that drain catchments with fine-grained sediments such as the Amazon, Huanghe, Yangtze, GBM, and Mekong. They may also be found in tropical areas where several small high-discharge rivers form coalescing estuaries along significant stretches of coasts with moderate to large tidal ranges, as on the Guinea coast of West Africa (Anthony 2006). On these

coasts, significant swell wave-energy dissipation may occur over muddy inner shorefaces and over alongshore migrating mud banks. The longshore transport of particulate mud and of mud banks on these coasts is generally assured by strong tidal currents aided by synoptic wind-forced currents. In all such settings, periodic storm wave activity over the muddy shores may lead to the reworking of disseminated shells, sand, or gravel into a distinctive type of beach ridge called cheniers that rest on muddy deposits (see entry on “► Cheniers”).

Examples of Wave and Tide-Generated Processes and Interactions

The fundamental process feature of wave-tide-dominated coasts which differentiates them from pure wave- or tide-dominated coasts is the way the waves and the tides interact to give distinct process signatures, sediment transport patterns, and coastal morphologies. This having been said, it is by no means easy to actually distinguish wave- from tide-generated processes, mainly due to the difficulties and high computational cost of developing coupled wave-tide models (Hashemi et al. 2015, 2016). Depending on the position in the wave-tide-dominated spectrum, the inner shoreface on moderate- to high-energy swell wave coasts may show a regular bathymetry related to long-term planation by waves, with tidal and wind-induced currents aiding in the transport of sediment that is put into suspension by the high-energy waves. On low to moderate wave-energy coasts with strong tidal action, the inner shoreface may exhibit either an irregular, poorly reworked topography comprising inherited drowned features, or may be reworked, when loose sand is abundant, into tidal current ridges and banks. Off high sediment discharge river mouths, the inner shoreface may be draped with mud that may be in equilibrium with waves and tidal and wind-induced currents. On macrotidal shorefaces and shallow epicontinental tidal seas such as the eastern English Channel, the southern North Sea, and the Gulf of Bohai, where tidal currents exceed about 0.5 m s^{-1} , and where sand and mud are abundant, tidal and storm-induced currents may be responsible for significant redistribution of these facies (e.g., Vincent et al. 1998; van de Meene and van Rijn 2000; Kleinhans and Grasmeijer 2006; Héquette et al. 2008, 2013; Baeye et al. 2011; Bricheno et al. 2015). In these settings, the hydrodynamic regime of the shoreface appears to exhibit a dominant tidal signal that may be modulated by storms, the influence of which tends to dominate toward the shore, thus, imprinting a cross-shore spatial pattern of wave-tide modulation (Anthony and Orford 2002). Such tidal modulation of wave power on a macrotidal shoreface has been confirmed from Eulerian field measurements of wave height (Davidson et al. 2008). Waves are considered as an important

stirring mechanism on macrotidal shorefaces, with the mean currents from tidal, wind- and density-driven currents generating bedload transport and suspension transport of the finer fractions. Bedload transport is an important portion of the net annual sediment transport over many shorefaces and shelf environments where tides are important, but it is virtually impossible to isolate the fraction of bedload transport respectively under waves without currents, and under currents without waves (Kleinhans and Grasmeijer 2006). Forms associated with these processes are generally sand ridges and banks (Dyer and Huntley 1999). Many linear current banks, ridges, and dunes are generally oriented more or less parallel to the shore in response to longshore tidal currents. From detailed monitoring of sand dunes off the Belgian coast, Van Lancker (1999) showed a vertical pattern of wave-and-tide reworking. Dune mobility in shallow depths (<5 m), whatever the distance offshore, was due mainly to wave action while deeper water mobility was caused by tides. In the shallow eastern English Channel and southern North Sea, storm waves tend to drive subaqueous banks inshore, a process that sometimes tends to be countered by the longshore tidal currents (Fig. 1). This may result in stretching and eventually division of the banks (Tessier et al. 1999). Banks that do get close inshore may eventually become attached to the beach, leading to significant accretion and onshore feeding of eolian dunes (Héquette and Aernouts 2010; Anthony 2013). On coastal sectors where the morphodynamic balance is such that the banks do not move inshore, this may eventually deprive beaches and dunes of sand in spite of the abundant nearshore sand stocks.

The nearshore residual tidal current signature may become strengthened or weakened by synoptic winds, thus supporting the differentiation of tidal asymmetry patterns that determine, at various timescales, residual sediment transport. Ebb- or flood-dominated flows lead to well-defined sand transport pathways such as those that run along both the French and English coasts in the eastern Channel and Dover Straits (Beck et al. 1991; Grochowski et al. 1993). As a result of various hydrodynamic conditions, notably orientation of the synoptic wind field and Coriolis deflection of the water mass, tides are larger on the French coast and the tide-dominated sand transport pathway on this coast has been much more active than that on the English side. It has been suggested that this sand-rich pathway has fed the large Holocene dune fields on the French coast via storm reworking of the nearshore sand stocks, and shoreward transport by winds over the wide dissipative beaches (Anthony 2013).

The relationship between tides and waves has been most readily invoked in explaining the influence of large tidal ranges on beaches (Davies 1980), and on beach morphodynamics (Wright et al. 1982; Carter 1988; Short 1991; Masselink and Short 1993; Masselink and Turner

1999; Levoy et al. 2000; Masselink et al. 2014). Generally, when sediment supply conditions are favorable, coasts with large tidal ranges adjust, over the long term, to the important vertical tidal excursion by having a significant intertidal volume of sand or gravel that necessitates relatively high wave-energy levels in order for large-scale morphological changes to occur. In northern France, ridge and runnel beaches (Fig. 2) with mean spring tidal ranges of 5–9 m have intertidal volumes of up to $1,200 \text{ m}^3/\text{m}$ of coastline, compared to the moderate volumes of beaches in settings with low tidal ranges (generally less than a tenth of this volume). As a result, whatever the background wave-energy level, the rates of sediment transport and beach morphological change are retarded. The greater the tidal range and the lower the modal wave energy, the more retarded are these changes. Daily wave reworking of beach morphology in such cases may become limited to minor changes in bedforms that are themselves strongly dependent on the tidal cycle. In storm-wave settings, such as those of northwestern Europe, change becomes significant only during storms. This effect may extend to the shoreface. From hydrodynamic data collected on several megatidal beaches (the term was used by the authors to designate beaches with tidal ranges exceeding 8 m), Levoy et al. (2000) identified significant beach and shoreface tidal modulation of wave heights, with wave heights being lower at low tide than at high tide. Similar conditions of tidal modulation have been reported in a plethora of studies of other beaches, among which recent examples are Masselink et al. (2014), Almeida et al. (2015), and Kang and Kim (2015). The effect is exacerbated over shallow offshore areas where the tidal fluctuation induces corresponding changes in water depth.

The large beach volume and the important horizontal tidal translation also imply a reduction in overall beach gradient. A common feature of the tidal influence on beaches with large tidal ranges is that the various wave zones migrate rapidly across this wide, low-gradient profile during the tide, resulting in significant cross-shore variations in the hydrodynamics and resulting morphology (Kroon and Masselink 2002; Anthony et al. 2004). The wave influence is toned down because the sum of energy spent per unit area over time is much less than on beaches with low tidal ranges. Whereas entrainment thresholds are the same, the volume of sediment transport is thus much less on beaches with large tidal ranges. The beach sediment volume and slope being equal, the larger the tidal range the greater the variations in morphodynamic behavior between the lower beach and the upper beach. On megatidal beaches in northern France, the morphodynamic domains may range from extremely dissipative at low tide on the lower beach to moderately reflective on the upper beach at high tide. At high tide, the extreme lower beach is subject to a combination of shoaling waves and strong longshore tidal currents. On similar megatidal gravel

beaches fronted by sand flats in northern France, the upper beach is extremely reflective and is dominated by sub-harmonic gravity wave motions whereas the lower beach is highly dissipative and shows infragravity edge wave motions, and strong, tidally induced longshore currents. Where wind, sand size, and supply conditions are favorable, wide, low-gradient beaches associated with large tidal ranges may provide significant surfaces for eolian reworking of beach sand into coastal dune fields. These wide beaches are also sometimes characterized by strong groundwater table fluctuations that may actively affect beach face hydrodynamic conditions and stability (Turner 1993; Masselink and Turner 1999).

On wave-tide-dominated beaches, waves induce sediment suspension while transport is assured by strong tidal currents (Davidson et al. 1993; Masselink and Pattiaratchi 2000), although the most active tidal phase appears to differ on different beaches, depending on the effects of friction and on whether the tidal wave is progressive or standing. Tidal flow asymmetry on such beaches may determine preferential sediment transport patterns and directions. On some French sandy beaches in the eastern English Channel, the progressive tidal wave is associated with strong longshore currents at high water that are highly effective in terms of sand transport because they coincide with higher high-tide waves that lead to suspension of fine sand. Such currents are always directed northward or eastward, and are therefore very important in terms of the net long-term bedload and suspended sediment (including pollutants) transport in these directions. Low-tide waves and tidal currents on the lower beaches are generally much weaker, probably because the shallower water results in both enhanced wave dissipation and tidal retardation by increased bed friction, whereas tidal current strength is generally greater seaward in deeper water.

Management Aspects of Wave-Tide-Dominated Coasts

One rationale for recognizing distinct wave-tide-dominated coasts is that such coasts may be characterized by specific management aspects that require an understanding of the mixed process regime and sediment transport and accumulation patterns. The commonly wide beaches in settings with large tidal ranges may be utilized for a wide range of activities (notably space-consuming leisure activities such as speed-sailing) that may be mutually exclusive and whose practice may therefore require set regulations and space allotments in highly frequented zones. The distinctive coastal profile of such areas may require the clear recognition in management strategies of the specific cross-shore and longshore dynamics due to both wave action and tidal currents. The classic management notion of coastal sediment cells (e.g., van Rijn 2011)

based on longshore wave-energy gradients when coasts with low tidal ranges are considered is not easy to apply on wave-tide-dominated coasts because of the significant modulation of the energy spectrum by tidal currents (Sedрати and Anthony 2014). Sediment-cell definition criteria need to integrate the tidal dimension in sediment transport when wave-tide-dominated coasts are considered. Wave and tidal sediment transport may work over wide low-gradient intertidal and subtidal profiles, and this dimension needs to be taken into account in the design of coastal structures such as groins, but also in monitoring coastal pollution. Port breakwaters, for instance, are much more easily bypassed by tidal transport of sediment than on simple wave-dominated coasts, thus requiring constant dredging operations down-drift.

In areas where the nearshore zone is characterized by tidal sand banks and ridges, such as the English Channel and southern North Sea, changes in the volume and positions of these features in time may lead to navigation hazards as well as exposure or destabilization of cables and pipelines. These banks and ridges are also important in dissipating storm waves and are therefore significant in terms of shoreline protection, while being also a potential source of sand for beaches and dunes, as well as of dredged aggregate.

Summary

Understanding how tides and waves interact is essential to many areas of coastal and marine research and resource management, including a better understanding of coastal response to storm surges, and to the impacts of climate change. This interaction is also fundamental to the stability and management of many coasts. Wave-tide-dominated coasts may be defined as coasts the morphology of which is shaped by short- to long-term background hydrodynamic conditions generated jointly by waves and tides. Whereas waves are generally considered as the dominant source of processes of coastal change, the importance of tides was long overlooked, mainly because of lack of studies that properly evaluate their role, notably in settings deemed as wave-dominated. Although it has become clear that tides play a significant role in settings associated with waves, thus warranting the recognition of mixed, wave-tide-dominated coasts, more studies are needed to elucidate the various ways tides interact with, and modulate waves, and overall sediment transport. Such studies are also important in understanding how waves and tides contribute to extreme water levels and flooding and erosion hazards along the coast, especially with the perspective of sea-level rise.

Cross-References

- [Barrier Island Landforms](#)
- [Beach Processes](#)
- [Dissipative Beaches](#)
- [Shelf Processes](#)
- [Tide-Dominated Coasts](#)
- [Tides](#)
- [Wave–Current Interaction](#)
- [Wave-Dominated Coasts](#)

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Weathering in the Coastal Zone

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Weathering, anywhere on the earth's surface, is a destructive process whereby preexisting rocks, boulders, pebbles, or smaller particles are altered in composition, texture, color, and/or hardness with little or no transport from their original *in situ* positions, except as mediated by chemical solutions. In soil science, it is the initial step in a long and complex process (Merrill 1897/1921; Carroll 1962; Barshad 1972; Yatsu 1988; Birkeland 1999). As a process in physics, this rock decay follows the Second Law of Thermodynamics, as positive entropy, in which systematic ordering of chemical elements is progressively broken down. In a mineral petrological sense, the Bowen reaction series is reversed (Colman and Dethier 1986).

The *agencies of weathering* are multiple, often operating in a synergistic way, and rarely in total independence. The net product contributes to the “*Mantle*” or “*Regolith*,” the term

coined by Merrill (1897/1921, p. 299) for unconsolidated material without specifying genesis (Fairbridge 1968; Gale and Hoare 1991). The agents of weathering are threefold: (1) *Physical weathering*, such as thermoclastic processes (“thermoclasty”), involving expansion and contraction, operating both in low latitudes (notably the hot deserts) and mid-to high-latitudes (due to freeze and thaw). There is also “Fretting,” a process of weathering by wetting and drying (especially by sea spray). The scale ranges from granular-sized disintegration (“granulation”) to the splitting of giant boulders, 3–5 m in diameter. (2) *Chemical weathering*, operating principally in the humid climates, both warm-wet and cool-wet types. The weathered material is termed *Saprolite* (literally “rotted rock”). Chemical susceptibility depends partly on the nature of the solvent (the pH of the ground or seawater, acidic or alkaline), and on the minerals themselves, the solubility of which may be positively or inversely related to temperature, for example, the calcite or aragonite of limestone dissolves more freely as the temperature falls, but the K-feldspars of granite dissolve more readily with a rise in temperature and at low pH, but quartz (SiO₂) dissolves faster at a high pH. The progressive nature and stage is referred to in terms of “*Weathering Maturity*” (Garner 1974, p. 269). Under warm freshwater conditions the ultimate mineral product of granitoid weathering is *kaolinite* (which has a group formula of Al₂Si₂O₅(OH)₄) and has the lowest cation exchange capacity). In the warm coastal zone the littoral belt is often marked by mangrove swamps. There, the decaying leaves and organic debris forces the normally high pH seawater to an acidic range (pH 4–6), creating mud-filled pools. This acidification favors removal of SiO₂ and concentration of *bauxite* (ideally Al₂O₃ · 2H₂O; but more often a mixture of aluminous laterite). Commercial bauxites on the seaward side of the Guyana Shield and of the West-Australian Shield appear to mark former (pre-Quaternary) eustatically high sea levels. (3) *Biological weathering* relies largely on the presence of liquid water (at equable temperatures), and is thus inhibited by the hot-dry climates of deserts and by the cold-dry climates of the very high latitudes. The synergism of plant roots is particularly significant, for example, the thermoclastic crack in a mineral is exploited by root penetration. But the physical expansion of the root opens the crack still further, accelerating the disintegration process. Contributing also is the chemical (biochemical) role of decaying organic matter, and root “exudation” or granulation. Mineral diagenesis that involves hydration leads to expansion and thus further accessibility for groundwater (and roots). Animals play an important role, especially in limestones, which often favor the roots of plants or algae that crabs, gastropods, bivalves, and ecinoids like to eat. “Lithophagous” mollusca do not actually “eat” the limestone, but mechanically bore into it for the algae living in it. The byssus of *Mytilus*, and related bivalves, is an anchoring device, but is often torn away to expose fresh

rock to weathering. In the coastal zone, weathering is an essential forerunner to the various processes of *Erosion* (*q.v.*) which shifts or displaces the original material that has been made softer or more “erodable” by the loosening or softening effects of weathering. The process may be (1) subaerial, or (2) marine in its action, but for most rock material the rates involved are generally about five or six orders of magnitude less for the latter. The boundary between these two realms is accentuated by the action of tides, waves, and spray, and as a rule-of-thumb the level of mean low tide may be taken as the lower limit of subaerial processes. As an exception to this rule may be the occasion of a tropical hurricane on a coral-reef coast, when a meteorological blocking leads to a dead calm with 3 days or so of steady rain creating an aerated freshwater layer up to about 60 cm deep, effectively killing the surface reef.

Normally speaking, the low-tide boundary means that coastal rocks are made systematically more erodable, while their submarine extensions are differentially preserved. Depending upon the lithology, there is often a littoral bench or platform produced, the loose weathering products having been removed by mass wasting, wave action, or floating ice. *Lithology* of the coastal belt plays a critical role in the effectiveness, style, and rate of weathering. Bearing in mind that since the last glacial phase most of the world’s coastlines in the more stable regions have only been exposed to marine forces for about 6,000 year, the rate of weathering is often critical. Important for this rate three lithic categories may be noted:

1. Massive plutonic rocks of granitic, dioritic, and gabbroic types, together with deeply crystallized metamorphics of granitoid types.
2. Feldspathic volcanics, pyroclastics, and derivative sandstones, including some graywackes.
3. Limestones, beachrocks, calcareous eolianites or calcarenites, algal and coral breccia.

These three categories need to be taken in conjunction with the long-term geological history of a given coastline, in other words, its paleoclimatic inheritance. Among the three categories listed above, rates of weathering are difficult to determine but certainly vary greatly with latitude in view of the physical processes introduced (Loughnan 1969; Ollier 1969a; Colman and Dethier 1986; Birkeland 1999).

Shore platforms develop mainly in the (2) and (3) categories (Stephenson 2001). Important also is the role of inheritance, former climates and eustasy, and also the role of sea ice which develops *well before* the glacier buildup causing sea-level fall (Fairbridge 1971, 1977).

For category (1), the massive plutonic rocks, weathering of a chemical nature is most evident in equatorial and temperate latitudes, where it penetrates the rock along major joint planes which in uncomplicated structural examples tend to produce cube-shaped blocks. Sharp corners and intersections became

gradually rounded off until a so-called “woolsack” is created. Around the fractured borders of Gondwanaland, much of this weathering began in the early or mid-Mesozoic (200–150 Myr ago). In the tropics and subtropics the repeated eustatic fall of sea level during the Quaternary glaciations led to brief episodes of torrential precipitation and giant mudflows, which transported the unweathered crystalline blocks to the coast, where recent and present-day wave action is removing the weathered “rind,” exposing totally unweathered, rounded boulders (Fairbridge and Finkl Jr. 1984). Fine examples can be seen around Africa, Brazil, and Australia. Where the massive crystalline rocks are still *in situ*, for example, in the Recherche Archipelago, off S.W. Australia, the smooth surfaces of unweathered material plunges directly from the coastal hills well far below present mean sea level (MSL). Little trace of an intertidal notch is seen, and present rate of intertidal coast retreat (over the last 6,000 year) is almost zero.

In the same category of massive plutonic rocks, but in the intermediate latitudes of the high-pressure belts, a strong alternation of wet and dry seasons prevails. During the Quaternary glacial stages, however, total aridity often prevailed, with seasons of cold-fog (“Atacama Desert” type). Also seen in the Namib desert of SW Africa, weathering pits develop on various scales (Goudie and Migon 1997). The process is called “exudation.” This is also the setting for a peculiar geomorphic form of pitting, *tafoni* (Fairbridge 1968). Its classic setting is on the island of Corsica in the Mediterranean; the mountains there were not glaciated but subjected to severe frost action and cold dry winds blowing off the Alps. Fine examples may be seen on the south coasts of Australia, South Africa, and South America. The mechanism appears to be a process of weathering pit development. On the shady side of the outcrop, the daily dew precipitation leads to chemical weathering of feldspar minerals. Wind action removes the insolubles but diffusion and evaporation leads to the strengthening of a hard rim. The pit progressively deepens in size to cave where a human being can stand up in it.

Yet a third type of weathering of the first category appears on the coasts of the high latitudes, where alternate wet-dry, freeze-melt conditions apply. These have been studied particularly in Quebec and eastern Canada (Dionne and Brodeur 1988); as well as in the southern Baltic where the clays become winnowed to concentrate the boulders. Winter freeze-up combined with tides and current action creates a mixture of frost debris, till clays and boulders that the French call *glaciel*. Freeze-thaw, combined with shore-ice pushed by currents and tides, is also seen on the coasts of Norway, where a prominent feature, the *strandflat* (*q.v.*) has long been a problem, a product of multiple cycles, involving both glacio-isostasy and glacio-eustasy.

The second weathering lithic type is that dominated by feldspathic minerals, which are susceptible to slow dissolution in acid groundwater (high levels of CO₂). This form of

weathering is most apparent around volcanic centers that have erupted with the andesitic-rhyolitic lava suites, that are typical of the circum-Pacific “Ring of Fire.” Weathering platforms are typically swept clean by wave action, and in places display the very evocative patterns of truncated pillow lavas.

Arkosic sandstones (i.e., rich in K-feldspars) are particularly common in the *flysch*-type facies and graywackes that are a feature of the great orogenic belts near former plate boundaries. These rock types are often heavily folded, but in coastal sectors they may be dramatically transected by contemporary (or at least Quaternary) marine erosion. That erosion involves a differential truncation made possible by the subaerial “preparation” by chemical weathering (Stephenson 2000). The role of subaerial weathering in the development of the Kaikoura Peninsula shore platforms on the south island of New Zealand has been studied in depth by Stephenson and Kirk (2000), who have comprehensively studied the roles of both animal and plant types as well as their interactions with the results of wave energy. A useful review of the whole topic of shore platforms was issued separately by Stephenson (2000), with the question of platform width (Stephenson 2001).

Finally, there are the limestones, along with beach rock, calcareous eolianites (calcarenes), coral and algal reef rock, and their breccias (Fairbridge 1968; Miller and Mason 1994). These are susceptible to solution in CO₂ rich (acid) waters, but seawater is alkaline (pH ca. 8), so at first sight this would not appear to be an effective process. The explanation lies in the development of biochemical microenvironments. Tide pools disclose an extraordinary range of pH (5–10) in 24 h periods. During hours of darkness, CO₂ production and cooling cause the pH to fall into the acid range and the pool margins become undercut (helped by gastropoda). The next morning, the pool warms and solar evaporation shifts the pH quickly to the alkaline range. A white precipitate (aragonite) is often seen, and the pure spaces in the country rock are tightly cemented. The limestone shores are commonly marked by nearly horizontal intertidal platforms with undercut notches (and possibly overhanging visors) on the landward side (Fairbridge 1968, p. 653). Tidal action is constantly saturating this undercut which is populated by unicellular blue-green algae which bore into the rock to depths up to 2–3 mm. The platform, for its part, is differentially hardened by interstitial precipitation of CaCO₃, in a plane approximating mean low tide (higher on exposed points). A limestone shore platform is thus a product of differential biochemical weathering and lithic hardening. An outer rim of algal construction (the so-called “Lithothamnion Rim” often forms a massive protective barrier on the seaward edge of the platform. Measurements of Holocene platforms suggest their landward growth is often at a rate of 1 mm/year. When measuring Holocene changes of relative sea level it is found the limestone shores provide the clearest evidence, even to cm-precision in favored places.

As to the *agents of weathering*, to say mechanical, chemical, and biological is simplistic, because in most settings there is synergistic interplay, so that there is commonly a highly complex sequence of reactions. In many standard works on coastal science, the subject is not even listed in the index. (Nevertheless, it gets 20 citations in Fairbridge 1968.) It became a topic of debate during the mid-to late-nineteenth century, particularly in Britain, when the physiographic questions of landscape came under consideration. The prominent action of marine (wave) erosion in western Europe tended to overwhelm the associated agencies and was favored by major physical geographers of the nineteenth century (e.g., Von Richthofen 1870–72). The role of weathering particularly chemical weathering, was able to gain ground in some circles (e.g., Whitaker 1867), and notably in New Zealand during the next century (Bartrum 1926). In the South Pacific, the “Old Hat” phenomenon, observed by Dana in the previous century on the US Exploring Expedition, was a critical factor, although little noticed in the Northern Hemisphere. Thus, it was mainly in Australia and New Zealand during the twentieth century that weathering became recognized as the key agency in coastal erosion.

Anomalies abound in coastal weathering situations. Disharmonic climatogenetic features are commonly sources of controversial interpretations. A classic example is at the Giant’s Causeway in Northern Ireland, to be seen repeated on the opposing shores of Scotland (Fingel’s Cave on Staffa), where the same Cenozoic basalts display spectacular columnar jointing. Paradoxically, the boulder beaches nearby are not characterized by sharply edged pentagonal or hexagonal blocks, but by well-rounded boulders. Their source is evident in the deeply weathered tropical soil that forms a transition zone above the fresh basalt, showing the progressive weathering of the columns into spheroidal “core stones” analogous to those of Devon’s well-known granite tors.

Differential hardness of the parent rock material is the explanation for the sorting and concentration of the Upper Cretaceous chert (flint) horizons of the Chalk, as notably developed along the English Channel coast of Dorset where it forms a massive boulder beach, the Chesil Bank. The soft chalk has been lost to weathering. Another process involving Upper Cretaceous chalk is observed on the Baltic coastal cliffs of eastern Germany and Poland. Frozen spray in winter constitutes a powerful weathering agent.

Salt spray is effective in all latitudes, leading to “haloclasty”, or what E.S. Hills called “fretting,” besides simple wetting drying, that is a grain-by-grain disintegration of the coastal rocks, most typical of sandstones. Due to inhomogeneities to the rock, cavities develop and the result is a network or lace-pattern, that is generally known as *honeycomb alveolar weathering* (see numerous illustrations in Fairbridge 1968). As in the case of tafoni, noted earlier, extra-

large examples are attributed to a dew-plus-chemical action that can operate far inland from the coastal zone (e.g., in the mid-Sahara), but the typical spray-activated alveolar weathering is most typical of the salt-crystal expansion in the supratidal zone. Ice-crystal expansion takes its place in the high latitudes (e.g., the “dry valleys” of Antarctica). Yet another agency in the coastal zone is biological where crabs and gastropods loosen mineral grains as they browse on algae.

Landsliding in the coastal zone is particularly evident in the “*Zone of Weathering*,” label introduced by Ruxton and Berry (1957) after a survey undertaken in the territory of Hong Kong which displayed weathering on a scale unfamiliar to eyes accustomed to the landscapes of NW Europe or eastern North America where repeated Quaternary glaciations and periglacial activity has greatly modified the warm-wet products of what is “normal” weathering in much of the world far removed from the direct influences of glaciation over the last 1–2 Myr. In the deeply dissected coastal areas of Hong Kong, likewise in the Seychelles and eastern Brazil, particularly in granite or granitoid rocks, the zone of weathering is commonly 50–80 m thick. Deforestation and other human activities have helped to accelerate landscape modification by massive landsliding. The latter, while not a weathering process itself, may be greatly enhanced by the preparation and lubrication provided. Even in a temperate zone like the south coast of England, landsliding of coastal cliffs is particularly prevalent, thanks to the action of groundwater in the porous stratified formations.

A combination of deep weathering and landsliding is also observed in the region of Rio de Janeiro on the coast of Brazil, but here it has been accelerated by differential uplift following the plate-tectonic separation from Africa. Granite plutons form “sugar-loaf” mountains (inselbergs) that in a relative sense have grown progressively higher through time, aided by eustatic fluctuations and secular lowerings of the water table (Bremer 1999).

Weathering profiles exposed in deeply dissected volcanic rocks such as in Fiji or in New Guinea (Ollier 1969b) may frequently exceed 100 m. However, if the volcanic necks are traced eastwards into the Louisiade Archipelago, they are often seen to have been completely isolated from their weathered mantle and the waves beat against a vertical wall of unweathered igneous rock.

An unusual volcanic landform results from the weathering of a porous pyroclastic lithology that is rich in K-feldspars. The result is a so-called “volcano-karst” which includes weathering pits, caverns, and gullies reminiscent of limestone karst.

Important for shore platform development is the “*Weathering Front*,” a concept introduced in Australia by Mabbutt (1961), and now adopted worldwide for the interface between fresh and weathered rock, to replace an earlier term “basal surface.” In crystalline rocks it represents the ultimate

saturation depth of groundwater, and therefore of atmospheric gases such as CO₂ and O₂. In porous, sedimentary sequences its definition is less straightforward, and in artesian basins the term has little use. On the contrary, in the ancient cratons it is very instructive, especially where such features reach the present-day coast.

For the Gondwanaland cratons such weathering fronts may be traced back many million years in time to the initiation of subaerial exposure. Over broad areas this zero date was in the Late Permian with the retreat of the last Late Paleozoic ice sheets about 250 Myr ago. A comparable date in the regions of the Laurasian glaciations would be only about 6,000 year ago, but nevertheless in some exceptional spots, for example, in eastern Quebec pre-Cretaceous weathering fronts are known. In extensive areas of South America, South Africa, Antarctica, and Tasmania the last major defining events were the basaltic lava and diabase intrusions that ended in the Early Jurassic about 200 Myr ago. In the Indian peninsula the same type of eruption, marked the Cretaceous/Tertiary boundary, was 65 Myr ago.

In each region the “geomorphic clock” started ticking at a different point in time. With the far-reaching taphrogeny that marked the breakup of Gondwanaland came isostatic adjustments and drainage system reorientations. Rivers do not rapidly abrade their beds in fresh crystalline rocks, but where the “bedrock” has been preconditioned by a deep weathering front, that stream bed will be rapidly incised.

One might contrast the interior Guyana Shield in South America (Garner 1974) and its limited weathering of sandstones, with the deep weathering front of the West Australian Craton where it is often situated at a depth of about 100–120 m (Finkl and Fairbridge 1979). In this case, the boundary can be directly inspected by taking one of the gold mining elevators in the region of Kalgoorlie. The mean surface elevation of the craton is about 400–500 m, but the highest eustatic level of the Late Cretaceous was nearly 300 m, so there has been progressive lowering of the streambeds in the major rivers, but only near the cratonic margins. A mean rate of cratonic lowering over much of Gondwanaland has been 10–100 cm Myr. This snail’s pace constitutes the “primary weathering cycle” on Gondwanaland. Only in the water-saturated stream beds approaching the grade dictated by sea level is chemical weathering feasible. And that MSL has fluctuated markedly, but shown a secular fall during the last 65 Myr.

Where the deeply weathered margin of the Deccan Plateau in India can be seen at the present-day coast, for example, around Goa and Kerala, the unsilicified latosol is terraced by marine erosion. The soft paleosols in the intertidal and supratidal zones there is subject to a *secondary weathering cycle* under its exposure to sea spray, tidal action, and biological agencies, further supplemented by the mechanical role of wave action.

This exemplifies the predominant role of weathering, on every scale, in predicting the erosion of coastlines.

Cross-References

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- [Cliffs, Erosion Rates](#)
- [Ice-Bordered Coasts](#)
- [Marine Terraces](#)
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- [Shore Platforms](#)
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- [Tafone](#)
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Wetland Restoration

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Wetlands are usually thought of as vegetated habitat on land that is flooded for at least part of each year. However, specific definitions vary. For example, the Canadian Wetland Registry defines wetlands as “land having the water table at, near, or above the land surface or which is saturated for a long enough

Wetland Restoration, Table 1 Definitions of terms related to wetland restoration

Constructed wetland: Increasingly used to refer to wastewater or stormwater treatment wetlands, but occasionally used interchangeably with “created wetland.”

Created wetland: Usually refers to a wetland built in an area that did not previously support a wetland.

Enhancement: An attempt to improve a specific wetland function, such as the ability of the wetland to support ducks.

Mitigation: Literally refers to the reduction of some impact, but is commonly used to describe a wetland that was restored or created to replace a wetland destroyed by development, especially under Section 404 of the US Clean Water Act.

Rehabilitation: Sometimes used to refer collectively to restoration, creation, and enhancement, but also used to refer to improving aesthetic qualities of a wetland.

period to promote wetland or aquatic processes as indicated by hydric soils, hydrophytic vegetation, and various kinds of biological activity which are adapted to the wet environment” (Mitsch and Gosselink 1993, p. 26). In contrast, the Convention on Wetlands of International Importance Especially as Waterfowl Habitat (also known as the Ramsar Convention) defines wetlands as “areas of marsh, fen, peatland or water, whether natural or artificial, permanent or temporary, with water that is static or flowing, fresh, brackish, or salt including areas of marine water, the depth of which at low tide does not exceed 6 meters” (Mitsch and Gosselink 1993, p. 26). While the first definition includes only vegetated habitats, the second definition includes nonvegetated habitats such as mudflats and shallow reefs. In a narrow sense, then, wetlands include salt marshes, mangrove forests, swamps, bogs, fens, marshes, muskegs, playas, moors, vernal pools, wet meadows, and similar areas, but in a broader sense wetlands also include sea grass beds, mudflats, coral reefs, shallow lakes, and shallow rivers (see “► Bogs,” “► Mangroves, Geomorphology,” “► Mangroves, Remote Sensing,” “► Salt Marsh,” and “► Wetlands”).

Restoration can be defined as “the process of repairing damage caused by humans to the diversity and dynamics of indigenous ecosystems” (Jackson et al. 1995, p. 71). Thus, wetland restoration is the repair of anthropogenic damage to habitat that is saturated or flooded by shallow water for at least part of the year. Related terms include construction, creation, enhancement, mitigation, and rehabilitation (Table 1). Approaches to restoration include activities such as planting hydrophytes, control of invasive plants, grazing control, removal of fill, removal of structures that restrict water flow, and placement of dredged material to provide elevations suitable for wetland establishment.

Why Restore Wetlands?

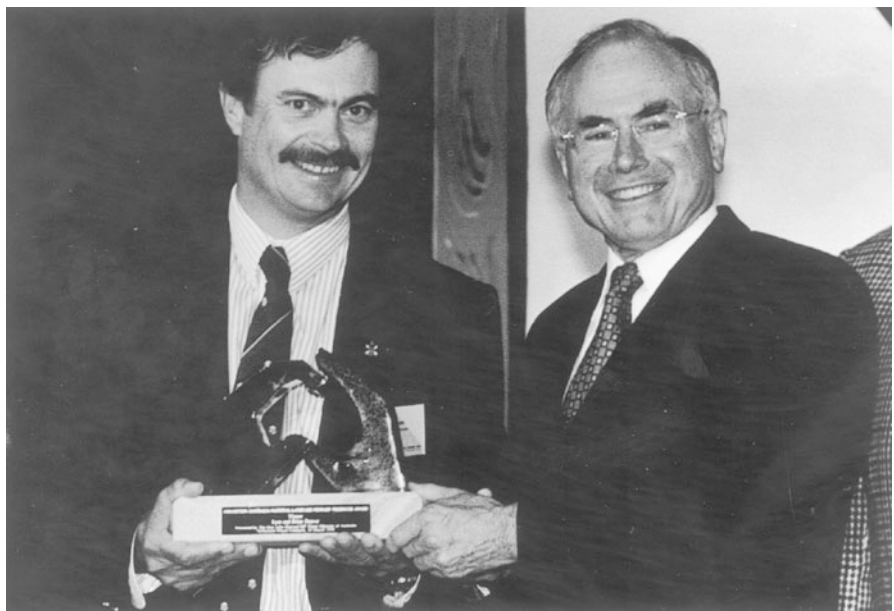
In some areas, such as California in the United States and Java in Indonesia, more than 90% of the historic area of wetlands has been lost (Dahl 1990; Crisman and Streever 1996). Although estimates are influenced by many factors, most estimates suggest that about 50% of the world’s wetlands have disappeared in the past few hundred years. Recognition of these losses, along with recognition of the economic and ecological value of wetlands and a general increase in environmental awareness, has inspired interest in wetland restoration. This interest is reflected in community action, incentive measures, and laws that require restoration under some circumstances.

Community action is often catalyzed by the efforts of nonprofit organizations, including relatively small organizations that operate locally or within a single country, such as the Yadfon (“Raindrop”) Association in Thailand, and larger organizations that operate internationally, such as the Mangrove Action Plan and Wetlands International (Quarto 1999). In some cases, community action manifests itself in the efforts of individuals undertaking restoration on their own property. For example, Kym Denver and his family have made extensive efforts to restore wetlands on their 1200-ha farm near the mouth of the Murray River in South Australia (Denver 1999) (Fig. 1).

In some nations, government-backed incentive plans provide the impetus for restoration. These incentive plans usually involve direct payments or tax relief for landowners participating in restoration programs. For example, in southwest England the Environment Agency has paid about £450 per ha per year to landowners for participation in the

Wetland Restoration,

Fig. 1 Australian Prime Minister John Howard (right) awards Kym Denver the 1998 National Landcare Primary Producer Award, in recognition of Denver’s efforts to restore wetlands on his family’s property in South Australia. (From Streever 1999; with permission of Kluwer Academic Publishers)



Levels and Moors Strategy, which seeks to restore and maintain wetland habitat (Jenkins and Sturdy 1999). Similarly, the Wetland Reserve Program in the United States pays farmers to restore wetlands on marginal croplands (Gaddis and Cabbage 1998). For government projects, policies, and funding packages provide incentives for restoration. Examples include the Commonwealth Wetlands Policy in Australia, which encourages government agencies to consider restoration opportunities, and the Coastal Wetlands Planning, Protection, and Restoration Act in the United States, which provides millions of dollars for restoration of wetlands. At the international level, Resolution VII.17 of the Ramsar Convention encourages signatory nations to incorporate wetland restoration as an element of national planning.

Some nations also have laws that can require wetland restoration to compensate (or “mitigate”) for destruction of wetlands that occurs as the result of development. In New South Wales, Australia, *State Environmental Planning Policy Number 14 – Coastal Wetlands* can require restoration as compensation for the destruction of wetlands. Similarly, Section 404 of the Clean Water Act in the United States can require restoration to compensate for wetland losses caused by filling of wetlands.

Approaches to Wetland Restoration

There are dozens of approaches to wetland restoration, ranging from planting hydrophytic vegetation to constructing substantial earthworks to using innovations that set the stage for natural processes to restore wetlands. Often,

different approaches are used together, in a single project. In all cases, restoration should begin with consideration of the impacts that caused degradation followed by development of specific objectives for the restoration project. In addition to ecological knowledge, successful restoration requires an understanding of hydrology and engineering as well as coordination among stakeholders that may include landowners, funding sources, nonprofit organizations, government agencies, and the local community. Although many handbooks provide guidelines for restoration, these guidelines are neither universally applicable nor definitive – every restoration project is to some degree unique, and every project requires some level of innovation and creativity.

One approach to restoration involves little more than planting. In the tropics, thousands of acres of mangroves have been planted, leading to reestablishment of mangrove forest that had been cut for production of charcoal or for building materials (see “Mangrove Restoration”). When vegetation has been removed from a wetland but hydrological conditions remain unchanged, planting may be all that is required to meet restoration objectives. Often, however, construction of dikes or other activities change hydrological conditions, rendering planting alone insufficient as a means of restoration.

If hydrological conditions have been changed, restoration should address the hydrological change. Complete restoration of hydrology is not always possible because of constraints related to urban infrastructure, flood control, and other factors, but partial restoration of hydrology can contribute substantially to wetland restoration (Streever 1998). For example, industrial and agricultural development has significantly affected the entire Hunter River Estuary in New South

Wetland Restoration,

Fig. 2 Bill Streever points to peat exposed by removing gravel from an abandoned work pad near the Beaufort Sea on the North Slope of Alaska. Gravel removal is the first step in tundra rehabilitation



Wales, Australia, but replacement of small-diameter culverts with bridges in selected locations has led to increased tidal flushing and subsequent increases in wetland area (Streever and Genders 1997). Similarly, in Florida, USA, dikes constructed as part of a mosquito-control program led to severe degradation of wetland habitat, but breaching of dikes to allow some tidal exchange led to rapid recovery of vegetation and fish populations (Brockmeyer et al. 1997). However, even if hydrological conditions can be restored, impacts resulting from drainage may be difficult to overcome. In some coastal areas that have been temporarily drained for farming or other purposes, oxidation and subsidence can lower elevations or oxidation of naturally occurring pyrite can result in soil acidification so severe that substrates will not support plants even after the land is reflooded (White et al. 1997).

Removal of fill has been undertaken at several locations to restore wetlands. In the Gog-Le-Hi-Te Wetland, in the Puyallup River estuary in Washington, USA, removal of about 55,000 m³ of dredged river sediments and solid waste contributed to partial restoration of a 2.2-ha area (Simenstad and Thom 1996). Other examples include the Tourle Street Bridge site near Newcastle, New South Wales, Australia, where slag from the iron-smelting industry was removed to restore intertidal wetlands (Day et al. 1999), the Sweetwater River Wetlands Complex near San Diego, California, USA, where fill was removed to mitigate for wetland losses resulting from highway construction (Pacific Estuarine

Research Laboratory 1990), and numerous sites on the Alaskan North Slope, where gravel fill has been removed to promote recovery of tundra near the Beaufort Sea (Jorgenson and Joyce 1994) (Fig. 2). In some cases, natural habitat is excavated to create wetlands. For example, on the Grand Bay-Bangs Lake Estuary in Mississippi, USA, about 10 ha of salt marsh was created by excavating sediment from a pine forest (LaSalle 1996).

In contrast to fill removal for wetland restoration, sediment dredged from navigation channels is sometimes placed in shallow water to restore or create vegetated wetlands (Streever 2000). Typically, hydraulic dredges remove sediment from a navigation channel and pump it through a pipeline to the wetland project site. Fine sediments must be contained within earthen berms or other structures to allow consolidation, while sandy sediments can be placed on shallow bay bottoms without confining structures. In the 1970s, when technology for dredged material wetlands was in the early stages of development, success was typically defined by establishment of vegetation, but in recent years increased emphasis has been placed on maximizing all aspects of similarity between natural and dredged material wetlands. For example, in the Atchafalaya Delta, Louisiana, USA, attempts have been made to build dredged material marshes with geomorphology similar to that of naturally accreting deltaic marshes (Fig. 3).

Restoration sometimes requires removal of contaminants, including contaminants from oil spills and industrial

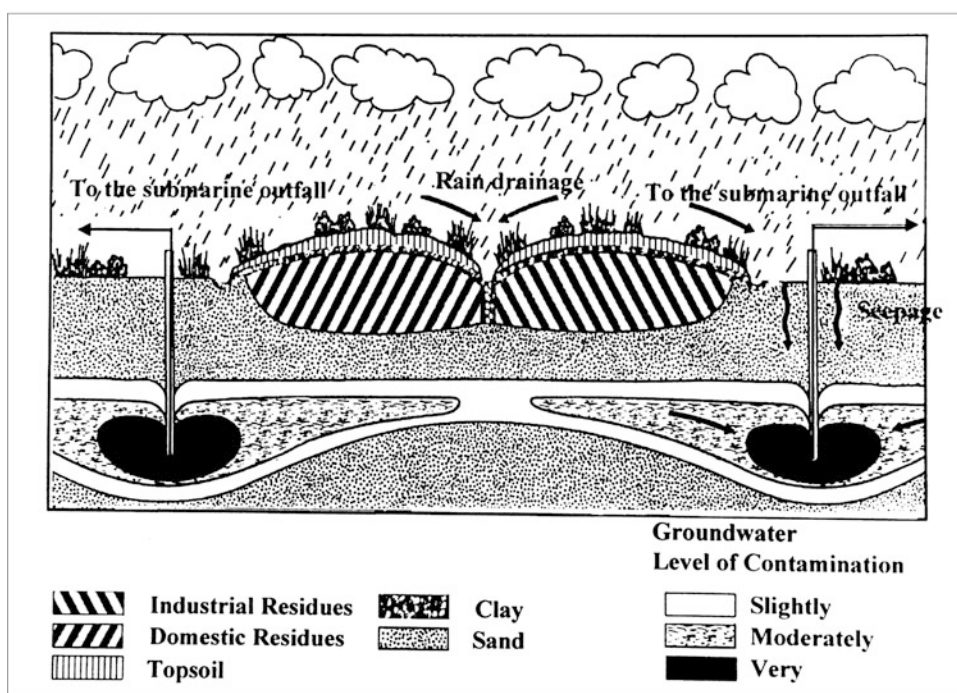
Wetland Restoration,

Fig. 3 Arrows point to recently created dredged material wetlands built to mimic the natural geomorphology of the Atchafalaya Delta, Louisiana, USA



Wetland Restoration,

Fig. 4 The Jauá Lake wetlands in Brazil (top) were degraded by spills that contaminated the groundwater, leading to a need for restoration that included groundwater withdrawal (bottom). (From Streever 1999; with permission of Kluwer Academic Publishers)



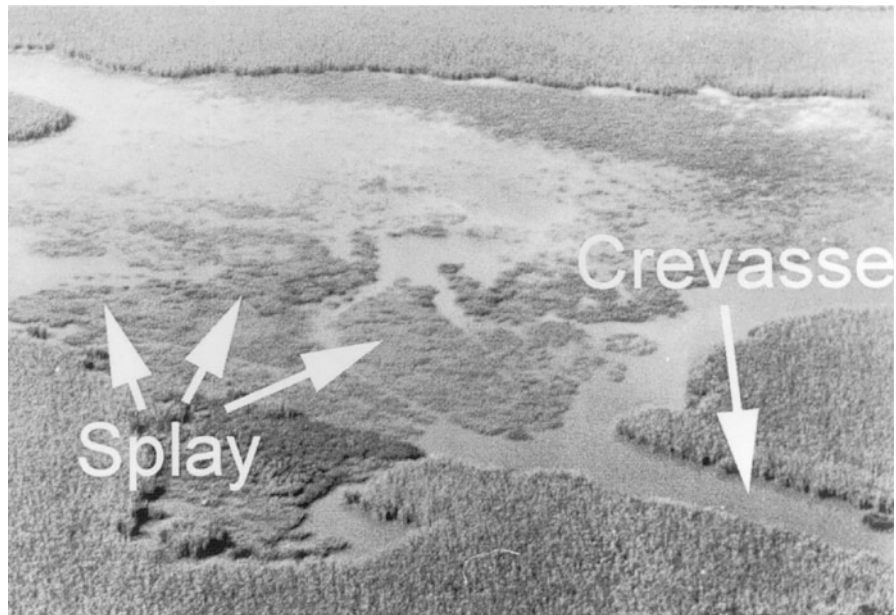
development. In some cases, contaminants enter groundwater. For example, the Jauá Lake wetlands, a system of freshwater wetlands sitting between an extensive stretch of sand dunes and the Atlantic Ocean in Bahia, Brazil, were contaminated when domestic and industrial waste infiltrated into groundwater, and restoration required removal of contaminated groundwater (da Silva et al. 1999) (Fig. 4).

For large-scale restoration, costs can prohibit extensive on-the-ground activities. This reality has prompted development of methods that rely on natural processes to restore wetlands.

For example, in Louisiana, USA, natural levees have been intentionally breached to facilitate deposition of Mississippi River sediments on shallow bay bottoms, leading to development of vegetated wetlands at costs as low as US\$54 per ha by mimicking the natural delta-building process of crevasse-splay formation (Turner and Boyer 1997) (Fig. 5).

Wetland Restoration,

Fig. 5 Crevasse splay restoration project near the mouth of the Mississippi River



Key Research Issues

As the number of restoration projects has increased over the past few decades, the amount of research on restoration has also increased, but the majority of this research has been recorded in reports with limited distribution. Most research consists of site-specific attempts to track wetland development, but some research has attempted to assess restoration efforts by looking at collections (or “populations”) of restored wetlands. Most research, whether it is site-specific or more broad, addresses two questions. First, how do restored wetlands compare to natural wetlands? And second, do restored wetlands follow a developmental performance curve, or trajectory, leading to increased similarity with natural wetlands over time? In addition to addressing these questions, restoration research has contributed to improved methods of plant establishment, development of engineering alternatives for shore protection, a better understanding of relationships among physical and biological components of wetlands, and other topics. Despite the amount of research that has been completed, useful broad generalizations about wetland restoration remain elusive.

Studies comparing natural and restored wetlands often suffer from pseudoreplication (in the sense of Hurlbert (1984)), in which conclusions about all natural and restored wetlands are generalized from data based on replicate samples collected from a single natural reference wetland and a single restored wetland. However, over time data have accumulated from hundreds of studies comparing pairs of natural and restored wetlands. Cumulative results from these studies and from studies of populations of wetlands clearly indicate that restored wetlands often do not have all of the same structural and functional attributes found in nearby natural

wetlands, even many years after restoration efforts have been completed. That is, some attributes of restored wetlands, such as percentage cover by vegetation and abundance of invertebrates, may be indistinguishable from those of natural reference wetlands, but other attributes of restored wetlands, such as soil organic content and rates of nutrient cycling, may be clearly different from those of natural reference wetlands. Furthermore, some differences may persist indefinitely. Zedler (1999) noted that similarity between restored and natural wetlands depends on the degree of degradation at the beginning of the restoration project as well as the degree of effort invested in restoration.

When variables such as abundance of invertebrates or organic content in soils are plotted against time, the resulting curve is sometimes called a “performance curve” or “trajectory.” However, because the term “trajectory” implies a specific type of curve (i.e., the curve described by a projectile moving through space, or a curve that passes through a set of points at a constant angle), the phrase “performance curve” may be more appropriate for describing plots that illustrate wetland development over time. Performance curves can be plotted in at least two ways: as measured values plotted against time, or as relativized values plotted against time. Relativized values are measured values that have been standardized against values found at a natural reference site or sites, usually by dividing the measured value from the restoration site by the measured value of the natural reference site or sites. Relativized values help account for intra- and inter-annual variability in measured values. Because there are few long-term data sets for individual restored wetlands, some researchers have endeavored to analyze data from multiple restored wetlands of different ages, making it possible to plot performance curves across time but adding the confounding

factor of site-to-site variability. Simenstad and Thom (1996) described five types of performance curves, including curves that (1) converge with the mean found for reference wetlands, (2) approach and surpass the mean found for reference wetlands, (3) develop toward the mean found for reference wetlands but then stabilize at a level lower than that of reference wetlands, (4) follow a sigmoidal trend that reaches an asymptote at the mean found for reference wetlands, and (5) decline from a level higher than that of reference wetlands to a level below the mean of reference wetlands. In keeping with results of studies comparing natural and restored wetlands, currently available research clearly shows that different wetland attributes can have different performance curves even within an individual wetland. For most restored wetlands that have been closely examined, performance curves for at least some attributes have not reached levels found in natural reference wetlands even after many years of development.

The Future of Wetland Restoration

Research illustrating the shortcomings of restoration approaches may eventually lead to a reduced willingness to trade natural wetlands for restored wetlands, as is frequently allowed under legislation such as Section 404 of the Clean Water Act in the United States. However, wetland restoration has a short history, and current trends indicate that interest in restoring wetlands will continue to increase dramatically around the world. Research and experience will continue to contribute to improvements in restoration approaches, and innovative techniques are likely to reduce costs per area. The scale of some restoration efforts is also likely to increase, as exemplified by efforts to restore parts of the Florida Everglades and Louisiana's coastal wetlands in the United States. Techniques for wetland restoration used in wealthy nations are increasingly exported and adapted for use in the developing world. Although it is unlikely that restoration will ever lead to complete replacement of wetlands destroyed over the past two centuries, it is becoming increasingly likely that restoration efforts will complement preservation efforts and 1 day stem the ongoing loss of wetlands around the globe.

Cross-References

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- [Managed Retreat](#)
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- [Wetlands](#)

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Wetlands

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Wetlands are neither terrestrial nor aquatic systems and at the coast form part of a continuous gradient between upland and ocean. The most obvious characteristic of all wetlands is continuous, seasonal, or periodic standing water or saturated soil. However, the increasing understanding of wetlands as important ecosystems and the parallel recognition of the threats posed to them by urban development, drainage for agriculture, and other human pressures has led to the need for clearer definitions and classifications. Conservation and management measures focused at wetlands often involve restriction or regulation of activities at the site, and so the need for “delineation” or defining the boundaries of individual wetlands has become of concern to landowners, managers, and political leaders.

While much of the ongoing debate over “what is a wetland?” encompasses a range of environments from peat bogs to vernal pools to cypress swamps, the focus here will be on wetlands found in coastal areas – where the “wet” aspect of the wetland is specifically associated with tidal action, but not necessarily saline tidal waters. These would be classified as part of the marine, estuarine, and riverine-tidal systems under the classification methodology of Cowardin et al. (1979) or the marine and coastal section of the wetland classification approach adopted by the Ramsar Convention Bureau (Scott and Jones 1995). There are many excellent texts which describe the range of processes and characteristics of wetlands around the world including Mitsch and Gosselink (2000), Finlayson and van der Valk (1995), and Mitsch (1994).

Here we will discuss the definitions that have been used to delineate wetlands and how these may be used relative to coastal wetlands, assess the functions and values of various types of coastal wetlands, and evaluate some of the current threats to coastal wetlands and the management measures that are needed to appropriately respond.

Definitions

Although most definitions encompass three fundamental characteristics of wetlands – the presence of water at or near the surface, soil conditions different from those in upland areas, and vegetation dominated by hydrophytes – the diversity of wetland types means that the combination of these three elements into a meaningful definition has been challenging. The purpose of such a definition should be to readily distinguish wetlands from other ecosystems. In some cases the definitions used have emphasized characteristics of the wetland that serve particular purposes or interests. An example is an early definition, the first in the United States, by the US Fish and Wildlife Service (Shaw and Fredine 1956):

The term “wetlands” . . . refers to lowlands covered with shallow and sometimes temporary or intermittent waters. They are referred to by such names as marshes, swamps, bogs, wet meadows, potholes, sloughs, and river-overflow lands.

This definition goes on to encompass shallow lakes and ponds but not deepwater areas and permanent streams. Also excluded are areas that are flooded on such a temporary basis that the area does not support “moist-soil” vegetation. This on the basis that they were of little value to the wildlife species of concern at the time.

Under the Convention on Wetlands of International Importance Especially as Waterfowl Habitat, a global environmental treaty signed in Ramsar, Iran in 1971 and commonly known as the Ramsar Convention, “wetlands” were defined as:

. . . areas of marsh, fen, peatland or water, whether natural or artificial, permanent or temporary, with water that is static or flowing, fresh, brackish or salt, including areas of marine water the depth of which at low tide does not exceed six metres.

The Ramsar Convention also provides that wetlands “may incorporate riparian and coastal zones adjacent to the wetlands, and islands or bodies of marine water deeper than six metres at low tide lying within the wetlands.” This very broad definition encompassed many areas, such as rice fields, that although technically wetlands, were unlikely to be considered of conservation value under Ramsar. While this definition still officially defines wetlands under the treaty, the Contracting Parties have adopted a classification system based loosely on that developed in the United States in the mid-1970s. The Cowardin Classification (Cowardin et al. 1979) divided wetlands into systems, subsystems, classes, and subclasses in

combination with a series of modifiers that describe water regime, water chemistry (e.g., salinity), and soil. The definition of wetlands used by Cowardin et al. (1979) includes descriptions of vegetation, hydrology, and soil:

Wetlands are lands transitional between terrestrial and aquatic systems where the water table is usually at or near the surface or the land is covered by shallow water . . . Wetlands must have one or more of the following three attributes: (1) at least periodically, the land supports predominantly hydrophytes, (2) the substrate is predominantly undrained, hydric soil, and (3) the substrate is non-soil and is saturated with water or covered by shallow water at some time during the growing season each year.

In an attempt to convey the ecosystem concept of wetlands, and to ensure that the essential role of hydrology as a driving force behind the physical, chemical, and biological features of the system is retained, the US National Research Council Committee on Characterization of Wetlands (NRC 1995) developed a broad reference definition of a wetland:

A wetland is an ecosystem that depends on constant or recurrent, shallow inundation or saturation at or near the surface of the substrate. The minimum essential characteristics of a wetland are recurrent, sustained inundation or saturation at or near the surface and the presence of physical, chemical and biological features reflective of recurrent, sustained inundation or saturation. Common diagnostic features of wetlands are hydric soils and hydrophytic vegetation. These features will be present except where specific physicochemical, biotic or anthropogenic factors have removed them or prevented their development.

At the coast, the flood and ebb of the tide will usually be the controlling factors in determining soil inundation or saturation. Importantly, while many coastal wetlands support halophytic vegetation, the presence of salt is not an essential feature of a coastal wetland. They also include areas further inland and the landward limit can be defined by the limit of tidal fluctuation. According to Field et al. (1991) there are over 4.5 million ha of coastal wetlands in the conterminous United States of which only 16% are salt marshes. The remainder are forested and shrub-scrub wetlands and fresh marsh areas. There may be some uncertainty in determining the boundary between tidal and nontidal wetlands and Mitsch and Gosselink (2000) note that tidal freshwater marshes can be described as intermediate in the continuum from coastal salt marshes to freshwater marshes. Similarly, freshwater woody wetlands or swamps, can occur in fresh nontidal waters and in areas influenced by the tide. This tidal–inland boundary has received less attention than that between wetlands and their adjacent uplands, likely because there is no jurisdictional difference between wetlands on either side of it. However, in some areas lands flooded by the tide are in public ownership by statute and it is likely, as this “tidal” boundary changes with sea-level rise, that criteria for the discernment of tidal versus nontidal wetlands in the coastal zone may be required.

Functions and Values of Coastal Wetlands

Coastal wetlands are valuable ecosystems. By providing refuge and forage opportunities for fishes and macrocrustaceans, swamps, marshes, and mangroves around the world are the basis of many communities’ economic livelihoods. Shallow ponds and seed producing vegetation provide overwintering habitat for millions of migratory waterfowl, and the structure of trees and forests is vital to songbirds and waders alike. The role of wetlands in uptaking nutrients and reducing loading to the coastal ocean is widely recognized, as is their value for protecting local communities from flooding—either by damping storm surges from the ocean or providing storage for riverine flood-waters. The economic value of tidal marshes and mangroves has been estimated at \$9990/ha/yr. in a recent assessment of ecosystem services (Costanza et al. 1997).

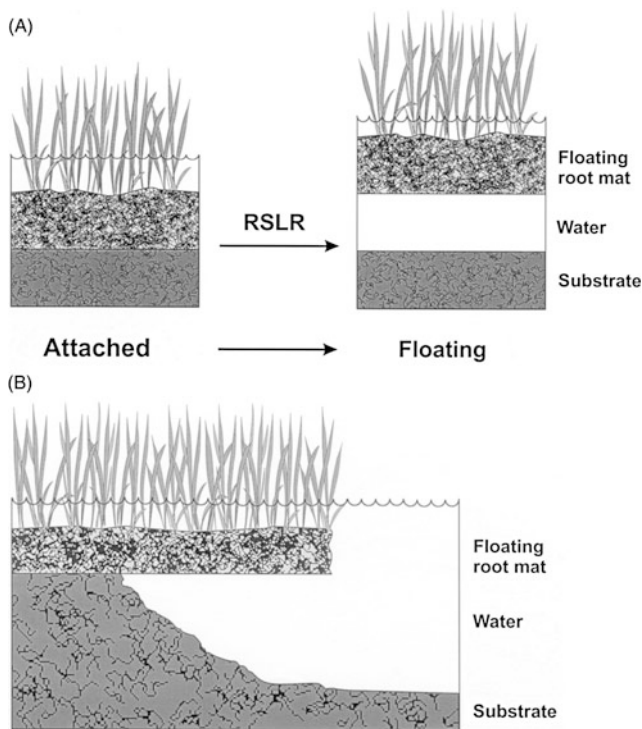
According to Mitsch and Gosselink (2000) coastal wetlands include tidal salt marshes, tidal freshwater marshes, and mangrove swamps. Tidal swamps will also be considered here as they are extensive at the heads of estuaries and in coastal plain environments where tides influence water levels at great distances from the coastline.

Salt Marshes

While saline marshes around much of the United States are dominated by *Spartina alterniflora*, regional variations include *Salicornia virginica* and *Spartina foliosa* in California and southern Oregon, to more diverse assemblages including *Distichlis spicata*, *Triglochin* spp., and *Plantago maritima* in the Pacific Northwest. In the west, these native halophytes are being supplanted at low tidal elevations by invasive *S. alterniflora* which can grow vigorously on open mudflats unavailable to the native species. Salinity typically varies between 15 and 33 ppt depending on coastal physiography and the level of freshwater inputs to the coastal ocean. Coastal salt marshes are highly productive systems. Tides, nutrients, and regular flushing offset the stresses associated with salinity, fluctuating temperatures, and alternate wetting and drying of the substrate. The detritus derived from this primary productivity supports higher trophic levels both via direct consumption and through microbial degradation which enhances the food value of the detritus. By providing shelter and refuge from predation, salt marsh margins are important nursery habitats for many fisheries species. For more detail see ► [Salt Marsh](#).

Mangroves

Mangroves, comprising approximately 60 species of trees and shrubs, are the predominant form of vegetation in the intertidal zone of tropical estuaries, lagoons, and sheltered coastlines. They are largely confined to latitudes between 30° N and 30° S of the equator, with a few notable exceptions that may be explained by the occurrence of warm ocean currents or by the



Wetlands, Fig. 1 Alternative views on the formation of “flotant” marshes in Louisiana. (a) The process described by O’Neil (1949) involving relative sea-level rise (RSLR). (b) The process described by Russell (1942) involving the growth of marsh into open water from an existing stand

presence of relict populations of more poleward past distributions (Duke 1992). According to one recent estimate, the total area of mangroves in the world is 181,000 km² (Spalding et al. 1997). While mangrove trees are notable for their characteristic adaptations to anoxic soil environments, such as prop roots and pneumatophores (see ► [Vegetated Coasts](#)), their ecological function is very similar to coastal salt marshes. For more detail see ► [Mangroves, Geomorphology](#) and ► [Mangroves, Remote Sensing](#).

Tidal Freshwater Marshes

Inland from salt marshes but still close enough to the coast to experience tidal fluctuations in water level are wetlands dominated by a variety of grasses and annual and perennial broadleaved aquatic plants. This wetland type is found most commonly in the mid- and south-Atlantic coasts and in Texas and Louisiana. No known plant species appears exclusively in tidal freshwater areas. Most marshes are dominated by a combination of annuals and perennials most of which also occur in inland freshwater wetlands. Plant diversity is much higher than in salt marshes and on much of the Atlantic coast of the United States there are substantial seasonal changes in plant coverage (Odum et al. 1984). In the late winter and early spring there is a period of almost bare mud with some

remaining stubble from previous years growth, followed by coverage with broad leaved plants in the late spring. By late summer the marshes are dominated by grasses and herbaceous plants. The soil conditions and hydrology of these marshes are very dependent on local conditions; although Odum et al. (1984) note that they are most likely fine mineral soils or peats.

In the Mississippi River delta plain a particular type of tidal freshwater marsh has developed in which, rather than being rooted in sediment, the plants live in an organic mat composed of live roots and decomposing litter. The level of this mat rises and falls with the tide. These marshes are termed “flotant.” While Mitsch and Gosselink (2000) noted similar floating marshes in the Danube Delta and in lakes around the world, in coastal Louisiana O’Neil (1949) reported 100,000 ha of freshwater and brackish floating wetlands. Flotant is described by Sasser et al. (1994) as “wetlands of emergent vegetation with a mat of live roots and associated dead and decomposing organic material and mineral sediments, that moves vertically as ambient water levels rise and fall.” Sasser et al. (1994) suggests that the floating marshes of the Mississippi delta plain have most likely developed in fresh marshes removed from riverine sediment inputs. There are several possible explanations for the formation of floating marshes. O’Neil (1949) suggests that floating marshes form when natural “attached” organic marshes are subjected to subsidence or sealevel rise and the buoyant organic mat is subjected to increasing upward tension. It eventually breaks free from its mineral substrate and floats (Fig. 1a). Alternatively, the floating mat could grow and expand out into open water areas from existing marshes (Fig. 1b).

Tidal freshwater marshes support a variety of waterfowl and animals from alligators to nutria. Their high primary production is only partly consumed directly with most being available to consumers through the detrital foodchain where benthic invertebrates play an important role. Mitsch and Gosselink (2000, p. 279) note that “of all wetland habitats, coastal freshwater marshes may support the largest and most diverse populations of birds.” These include wading birds, dabbling ducks, birds of prey, and shorebirds, and the structural diversity of the wetland vegetation interspersed with shallow ponds is an important factor in their use of these wetlands.

Freshwater Forested Wetlands

Coastal wetland forests frequently have an extremely diverse flora of tree, shrub, and herbs. They can be roughly divided into deepwater swamps, in the southeastern United States dominated by bald cypress (*Taxodium distichum*) and tupelo gum (*Nyssa aquatica*), with a red maple (*Acer rubrum*) and buttonbush (*Cephalanthus occidentalis*) understory; and seasonally flooded bottomland hardwood forests dominated by several oak species (*Quercus* spp.),

green ash (*Fraxinus pennsylvanica* var. *lanceolata*), and other hardwood species.

The character of coastal wetland forests is very sensitive to water level conditions, influenced in many areas by the tidal and gradual changes in sea level; as well as seasonal changes in riverine water levels at the heads of estuaries. Increases in freshwater discharge during floods are unlikely to result in any deleterious effects to freshwater forested wetlands as long as seasonal low-water periods allow for substrate drainage and forest regeneration. Bald cypress regenerates well in swamps where the substrate is moist and competing species are unable to cope with flooding. Seedlings must experience dry periods long enough to allow growth and survival of future flooding. However, prolonged flooding, associated with sea-level rise or a major change in riverine discharge regime, can result in changes. DeLaune et al. (1987) recognized gradients in seedling survival for both *T. distichum* and *Quercus lyrata* along an elevation gradient in a Louisiana coastal swamp. Increased flooding will thus reduce the area of appropriate elevation for seedling survival, and the findings of Rheinhardt and Hershner (1992) suggest that increased saturation of substrate will alter species composition in freshwater swamps. They found *Fraxinus* spp. and *Nyssa sylvatica* in wetter areas compared to *A. rubrum* and *Liquidambar styraciflua* within swamps of the Virginia coastal plain.

Sensitivities, Threats, and Management Approaches

Coastal wetlands are undoubtedly some of the most valuable ecosystems in the world and also some of the most threatened. Climate change and variability compound existing stresses from human activities such as dredging and/or filling for development, navigation or mineral extraction, altered salinity and water quality resulting from watershed management and non-point source contamination, and the direct pressures of increasing numbers of people living and recreating in close proximity.

The survival of coastal wetlands under conditions of altered climate really depends primarily on the ability of the plants to survive. The biogeomorphic processes which control the wetland landscape may produce gradual changes, for example, from marsh to mangrove swamp, as long as the environmental thresholds which control plant survival are not crossed. The challenge in predicting coastal wetland response, either the nature of the gradual change, or when and where the thresholds are exceeded, depends upon interactions among the responses to various climate forcing. The very low gradients of coastal wetlands make them particularly susceptible to changes in water level and any adjustments in

the hydrodynamic forcing from both ocean and watershed. While submergence associated with sea-level rise is frequently considered a serious threat to coastal systems, their inherent adaptation to coastal hydrology makes them perhaps less sensitive than terrestrial ecosystems subjected to frequent flooding by the encroaching sea. An examination of the response of various coastal wetland environments to climate change forcing illustrates the resilient nature of these systems and the complex biogeomorphic interactions that sustain them.

Increased freshwater is generally beneficial to coastal wetlands and a decrease may result in salinity stress for some communities. However, a gradual change does not necessarily result in plant death and wetland loss, and salt and brackish marshes may replace freshwater marshes and swamps. For example, the variation in salinities found by Visser et al. (1998) in marshes dominated by *Spartina patens* was so wide (average salinity 10.4–5.8 ppt), and overlapped with the zone dominated by *S. alterniflora* (average 17.5–5.9 ppt) that it seems likely that an increase of more than 1–2 ppt in salinity would be required to result in a significant change in habitat types in the more saline marshes. Decreases in salinity will likely increase the productivity of salt marsh plants as osmotic stresses are decreased. However, the salinity data of Visser et al. (1998) suggest that decreases in salinity of more than 5 ppt might be required for *S. alterniflora* to be replaced by more brackish marsh species.

Most coastal wetlands already cope with gradual sea-level rise. During the last 100 years, globally averaged sea level has risen at about 1–2 mm/yr. In most instances, these changes have not produced dramatic changes in coastal wetlands and accretionary processes have maintained wetland elevation and hydroperiod. Increased sediment delivery to coastal systems, resulting from future increased frequency of high intensity rainfall events, enhances sustainability by providing for vertical accumulation of substrate. Catastrophic sediment delivery could increase elevation sufficiently to convert coastal wetlands to areas of upland vegetation. However, unless the net effect is greater than the cumulative effects of sea-level rise, the effect should be seen as a type of dynamic equilibrium where episodic inputs counter long-term but low level rise in sea level. Indeed it appears that many coastal wetlands in areas of high subsidence—and thus high relative sea-level rise—maintain their elevation through the episodic input of sediments associated with moderate-low frequency storms such as frontal passages and hurricanes (Reed 1989; Cahoon et al. 1995).

Coastal storms can bring more than beneficial sediments. The resilience of coastal wetlands to storm effects likely varies with salinity. Fresher systems are more susceptible to saline incursions during storms. For example, in South

Carolina the storm surge caused by Hurricane Hugo flooded some low-lying coastal forests greater than 3 m deep with saltwater, and the surge carried saline water (20 ppt) along waterways up to several hundred kilometers inland (Blood et al. 1991). Elevated soil salinity levels lingered for months after Hugo, and had adverse affects on tree survival and growth, as well as on nutrient cycling processes. The effect of tropical cyclone activity on forested wetlands can be structural as well as hydrologic. Low salinity and freshwater forested wetlands differ in their susceptibility to tropical storm wind impacts, with damage to forest types generally greatest to pines, least to swamp, with hardwoods intermediate (Conner 1998). In the Atchafalaya River basin in south Louisiana, Hurricane Andrew caused little damage to cypress-tupelo communities but caused significant damage and mortality to bottomland hardwood forests (Doyle et al. 1995). However, in some areas in the vicinity of Cape Cod, MA, USA, Atlantic white cedar wetlands were more severely affected by blowdowns from Hurricane Bob than adjacent upland forests, a fact attributed to their relatively shallow root systems (Valiela et al. 1998). The long-term effect of storms on coastal wetlands depends upon the recovery time and the return interval of the storms. Increased frequency, if not intensity, of storm impact may therefore result in irreversible damage to forested systems or fragile organic substrates. For minerogenic salt marsh soils, however, storms can provide inputs of sediment essential to vertical sustainability.

It is important to recognize that these various responses may interact in ways that are difficult to predict, and that climatic and other stress factors and their interactions may result in a host of indirect effects on coastal wetlands, such as changes in patterns of herbivory or the rise in importance of alien species.

Management measures adopted to remediate loss of coastal wetlands must emphasize future sustainability of existing systems rather than re-creation of lost, impaired, or threatened habitats. The fisheries productivity, storm protection, and avian habitat provided by wetlands is dependent upon their landscape configuration, their internal structure, and their linkages with other coastal and marine resources. Such architecture is challenging to recreate and likely impossible to recreate given the other changes occurring at our coasts. Thus, management and planning strategies that provide for sustainable vertical accumulation, biodiversity, and linkages, both internally and with watershed and coastal ocean, are essential to wetland survival in the face of climate variability and long-term change. Climate change will continue to occur and it seems unlikely that development pressures will diminish in the near future (Nicholls and Small 2002). Appropriate planning and management to facilitate change and adjustment of the coastal wetland landscape, rather than total loss of this valuable habitat, must be

implemented to ensure the wetlands have a fighting chance (see ► [Coastal Management Practices](#)).

Summary and Conclusions

Coastal wetlands occur at the transition from terrestrial to aquatic ecosystems and on the topographic gradient between upland and sea. Their definition on the ground has been a topic of much discussion in recent decades but importantly, this discussion has clearly identified fundamental wetland characteristics that vary in nature from place to place but which can be used to successfully distinguish wetlands from adjacent lands or waters. At the coast, important gradients in salinity and tidal inundation control the specific ecology and geomorphology of wetlands. When combined with climatic factors these drivers result in environments as different as low salt marshes to cypress-tupelo swamps. Coastal wetlands are undoubtedly among the most productive and valuable ecosystems on earth.

The vital role of vegetation in the form and function of coastal wetlands makes them particularly susceptible to pressures associated with urbanization and industrial development at the coast (see ► [Vegetated Coasts](#)). However, coastal wetlands are able to withstand disturbances and stresses associated with climate variability and in most cases future changes in climate will result in transitions and adjustments between and within coastal wetlands rather than the catastrophic loss sometimes projected. Most mangrove and freshwater forested wetland plant species are millions of years old, and therefore have survived through numerous and very large-scale changes in climate and sea level, including several ice ages and changes in sea level on the order of 100 or more meters. Survival under such circumstances generally depended on the ability to migrate, rather than on the persistence of individual wetlands. Individual coastal wetlands today can either accumulate vertically to maintain their position in the coastal zone or migrate as in the past. Both can occur – where conditions are favorable.

Cross-References

- [Changing Sea Levels](#)
- [Coastal Management Practices](#)
- [Conservation of Coastal Sites](#)
- [Deltaic Ecology](#)
- [Deltas](#)
- [Mangroves, Ecology](#)
- [Salt Marsh](#)
- [Sea-Level and Climate Change](#)
- [Storm Surge](#)
- [Vegetated Coasts](#)

► [Wetland Restoration](#)

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Appendix 1: Conversion Tables

Metric to English units – equivalents of length

1 micron (μ) = 0.001 millimeter (mm) = 0.00004 inch (in.)
 1 mm = 0.1 centimeter (cm) = 0.03937 in.
 1000 mm = 100 cm = 1 meter (m) = 39.37 in. = 3.2808 feet (ft)
 1 m = 0.001 kilometer (km) = 1.0936 yard (yd)
 1000 m = 1 km = 0.62137 mile (mi)
 12 in. = 1 ft = 0.3048 m
 1 cm = 0.39370 in. = 0.032808 ft
 1 km = 10^5 cm = 0.62137 mi
 1 fathom = 6 ft = 1.8288 m
 1 nautical mile = 1.85325 km
 1 in. = 2.54001 cm
 1 ft = 30.480 cm
 1 statute mile = 1.60935 km = 5280 ft
 1 astronomical unit = 1.496×10^8 km = 92,957,000 mi
 1 light year = 9.460×10^{12} km = 5.878×10^{12} mi
 1 parsec = 3.085×10^{13} km = 1.917×10^{13} mi

Square measures

1 square foot = 0.00002295684 acre = 929.0 cm²
 1 acre = 43,560 ft² = 0.0015625 mi²
 1 yd² = 0.836127 m²
 1 hectare = 2.471054 acre
 1 mi² (statute) = 640 acres = 2.5900 km²
 1 cm² = 0.1550 in.² = 0.0010764 ft²
 1 km² = 10^{10} cm² = 0.3861 mi²
 1 mm² = 0.00155 in.² 1 in.² = 6.452 cm²
 1 m² = 10.764 ft² 1 ft² = 0.09290 m²
 1 km² = 0.3861 mi² 1 mi² = 2.5900 km²

Cubic Measures

1 gal (UK) = 4.5461 liters = 1.200956 gal (US)
 1 liter = 0.219969 gal (UK) = 0.264173 gal (US)
 1 gal (US) = 3.7854 liters = 0.832670 gal (UK)
 1 cc = 0.0610 cu. in. = 0.000035314 cu. ft
 1 cu in. = 16.387 cc
 1 cu ft = 28317 cc
 1 mm³ = 0.000061 in.³ 1 in.³ = 16.387 cm³ (cc)
 1 cm³ (cc) = 0.0610 in.³ 1 ft³ = 0.028317 m³
 1 m³ = 35.315 ft³ 1 mi³ = 4.1681 km³
 1 km³ = 0.239911 mi³

Statute miles to nautical miles to kilometers

Statute	Nautical	Kilometers
¼	0.22	0.40
½	0.43	0.80
¾	0.65	1.21
1	0.87	1.61
2	1.74	3.22
3	2.61	4.84
4	3.48	6.45
5	4.35	8.05
6	5.22	9.65
7	6.08	11.27
8	6.96	12.90
9	7.82	14.48
10	8.68	16.10
20	17.36	32.20
30	26.05	48.30
40	34.74	64.35
50	43.42	80.45
60	52.10	96.55
70	61.00	113.00
80	69.60	129.00
90	78.16	145.00
100	87.00	161.00

Fathoms to feet to meters

Fathoms	Feet	Meters
$\frac{1}{4}$	1.5	0.5
$\frac{1}{2}$	3.0	0.9
$\frac{3}{4}$	4.5	1.4
1	6.0	1.8
$1\frac{1}{4}$	7.5	2.3
$1\frac{1}{2}$	9.0	2.7
$1\frac{3}{4}$	10.5	3.2
2	12.0	3.7
$2\frac{1}{4}$	13.5	4.1
$2\frac{1}{2}$	15.0	4.6
$2\frac{3}{4}$	16.5	5.0
3	18.0	5.5
$3\frac{1}{4}$	19.5	5.9
$3\frac{1}{2}$	21.0	6.4
$3\frac{3}{4}$	22.5	6.9
4	24.0	7.3
$4\frac{1}{4}$	25.5	7.8
$4\frac{1}{2}$	27.0	8.2
$4\frac{3}{4}$	28.5	8.7
5	30.0	9.1
$5\frac{1}{4}$	31.5	9.6
$5\frac{1}{2}$	33.0	10.1
$5\frac{3}{4}$	34.5	10.5
6	36.0	11.0
$6\frac{1}{4}$	37.5	11.4
$6\frac{1}{2}$	39.0	11.9
$6\frac{3}{4}$	40.5	12.3
7	42.0	12.8
8	48.0	14.6
9	54.0	16.5
10	60.0	18.3
11	66.0	20.1
12	72.0	21.9
13	78.0	23.8
14	84.0	25.6
15	90.0	27.4
16	96.0	29.3
17	102.0	31.1
18	108.0	32.9
19	114.0	34.7
20	120.0	36.6
30	180.0	54.9
40	240.0	73.2
50	300.0	91.4
60	360.0	109.7
70	420.0	128.0
80	480.0	146.3
90	540.0	164.6
100	600.0	182.9

Appendix 2: Glossary of Coastal Geomorphology

Abrasion – The wearing away of a rock surface by friction as a result of the impact of wind, wave, current, or ice action, particularly when these agents are armed with rock fragments (notably sand and gravel): sometimes termed *corrasion*. Abrasion of a rocky shore can form ramps or platforms, or excavate furrows or potholes, while abrasion of a cliff can result in the cutting of caves, clefts, and crevices. Submarine abrasion takes place on the nearshore seafloor, diminishing offshore as water depth increases; it becomes imperceptible at depths greater than half the wave length (Bradley 1958). Rock fragments (such as quartz grains) can also be worn and pitted by abrasion when they are impacted against each other, or against a rock surface: this is known as mechanical abrasion, and is distinguished from etching produced by chemical solution (Corrosion, *q.v.*) (Margolis 1968).

Abrasion notch – An elongated cliff-base hollow (typically 1–2 m high and up to 3 m recessed) cut out by abrasion, usually where Breaking Waves (*q.v.*) are armed with rock fragments.

Abrasion platform – A smooth, seaward-sloping surface formed by abrasion, extending across a rocky shore and often continuing below low tide level as a broad, very gently sloping surface (plain of marine erosion) formed by long-continued abrasion (Johnson 1916). The intertidal section is typically 50–100 m wide, increasing with tide range.

Abrasion ramp – A smooth, seaward-sloping segment formed by abrasion on a rocky shore, usually a few meters wide, close to the cliff base.

Accretion – A gradual or intermittent natural process of deposition of sediment by wind, wave or current action, or by rivers, glaciers, solifluction or mass movement, resulting in the natural raising or extension of a land area.

Aeolian calcarenite, Aeolianite (Eolianite) – See Dune Calcarenite.

Aggradation – The raising of a land surface by deposition (vertical accretion) of sediment, as on a beach, dune, mudflat, marsh, coastal plain, or delta.

Algal mat – A carpet of blue-green algae (cyanophytes) that stabilize intertidal sediments and precipitate carbonates on the sheltered shores of hypersaline areas, found in the Arabian Gulf and the Red Sea, Shark Bay in Western Australia, the Bahamas, and Texas coast lagoons. See Stromatolite.

Algal rampart, Algal reef – A structure typically 10–20 m, occasionally up to 100 m wide, formed in shore and nearshore waters by the growth of calcareous algae (notably *Lithothamnion*, *Lithophyllum*, or *Porolithon* spp.).

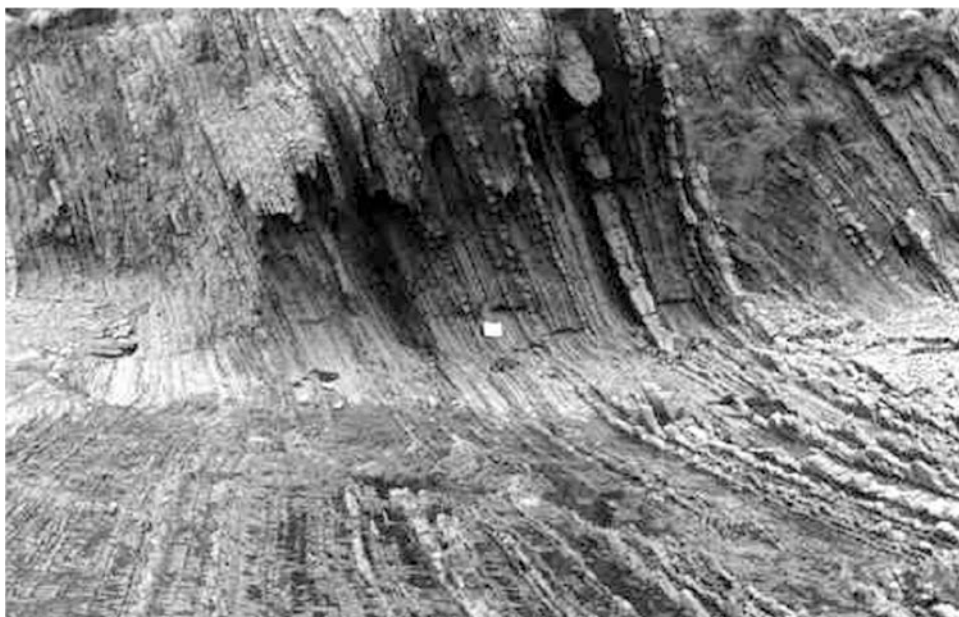
Algal ridge – A structure built by calcareous algae on the surface of a reef or shore platform.

Algal rim – A ridge built on the more exposed (seaward) margins of coral reefs or shore platforms by encrusting calcareous algae, such as *Lithothamnion* or *Porolithon*; they are usually no more than 20 cm high. Networks of ridges constructed by algae may enclose shallow pools on coral reef flats and shore platforms.

Artificial beach – A beach emplaced by human action, as where sand brought from the land, or alongshore or offshore sources is dumped on the shore, strictly where there was no natural beach, but the term is often used where a natural beach depleted by erosion is restored or renourished (see Beach Nourishment).

Atoll – A ring-shaped coral reef structure (ranging from <1 km to >100 km in diameter), partly or wholly enclosing a lagoon (typically 30–100 m deep and several kilometers wide); the outer (seaward) slopes plunge steeply to the deep ocean floor (oceanic atoll) or to the continental shelf (shelf atoll). The term comes from the Maldives, where “atolu” are government districts, each being a circular reef enclosing a lagoon. Atolls with a central (noncoral) island, such as Bora Bora, are termed “almost atolls”; those with ring-shaped reefs enclosing remnants of an earlier atoll, as in Houtman Abrolhos off the coast of Western Australia, “compound atolls.” Some compound atolls take the form of chains of small atolls

Abrasion notch at Cape Liptrap, Victoria, Australia.



Intertidal abrasion platform cut in Chalk near Birling Cap, Sussex, England.



(diameter about a kilometer), known as Faros, as in the Maldiv Islands, Indian Ocean. Horseshoe reefs form where isolated reef platforms (patch reefs) have marginal spurs that have grown to leeward.

Attrition – The wearing down of rock fragments by friction when they are mobilized and thrown against a rock surface, or ground against other rock fragments by wind action, waves, or currents.

Avulsion – The separation of an area of land when a river or estuary channel suddenly changes its course, usually during floods, or when a meander is breached. The term

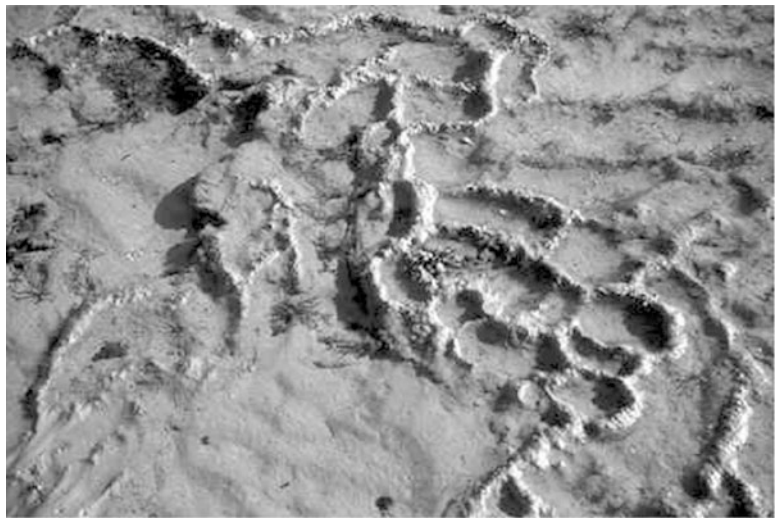
has been used obscurely, mainly by lawyers, to describe rapid erosion of the shore by waves. Another interpretation may be the transference of land along a coast as a result of rapid erosion on one sector and nearby related accretion, but this is not strictly avulsion, the placing of an area of pre-existing land on the opposite side of an earlier channel as the result of river migration.

Backshore – The coastal fringe lying above (i.e., landward of) the normal high tide line, but occasionally inundated by exceptionally high tides or storm surges. Also known as the supralittoral or supratidal zone.

Abrasion ramp at cliff base, North Island, New Zealand.



Algal ridges on shore platform at Port Hedland, Western Australia.



Backwash – The seaward (return) flow of water after the swash (uprush) produced when a wave breaks on the shore.

Bank – A localized broad elevation of the seafloor, smooth in profile, usually submerged even at the lowest tides, and composed of unconsolidated sediment (usually sand, sometimes glacial drift), subject to movement by waves or current action. Typically a few meters wide, but the term is used for larger features, such as the Dogger Bank in the North Sea or the Grand Bank of Newfoundland. See also Bar (shoal). A banner bank trails in the lee of an island or headland.

Bar – An elongated bank, ridge or mound of sediment (sand or gravel) a few meters wide, deposited and shaped by waves and currents, and submerged at least at high tide (i.e., can be partly or wholly exposed at low tide) (Shepard 1952). Bars commonly form off beaches, but also occur off

river mouths and lagoon entrances. A break-point bar is a concentration of sediment formed by breaking waves where material that is being carried shoreward meets that withdrawn from the beach by backwash. Longshore bars are formed parallel to a beach, and at least partly exposed at low tide, while oblique bars run at an angle to the beach, and transverse bars at right angles to it. Bars may also occur in cusped, lunate, looped, crescentic, reticulate, or chevron patterns. Multiple parallel bars and troughs form on gently sloping sandy shores where spilling waves break and re-form: they are exposed at low tide (see Ridge and Runnel). Trailing bars (banner banks) occur in the lee of headlands or islands, and become flying bars when they are disconnected (cf. trailing and flying spits, which differ in that they extend above high tide level). It should be noted that before Shepard defined bars, the terms such as

Amirante Atoll, Seychelles,
Indian Ocean.



Sand bars on the southeast coast of
Port Phillip Bay, Australia.



offshore bar, bay bar, and looped bar (Johnson 1919) were often used to describe features that would now be termed barriers.

Barchan – A dune form, an isolated, mobile crescentic mound of bare sand, typically up to 30 m high, with a steep advancing leeward slope and a gentler concave windward slope, flanked by lateral horns curving downwind and spaced at up to 350 m. Commonly found in deserts,

barchans also form on extensive areas of bare drifting coastal dune sand, and on unvegetated backshores.

Barrier (Coastal barrier) – An elongated ridge of deposited sediment (sand, pebbles, occasionally boulders) that has been built up by wave action above high tide level along the coast or across an embayment (bay barrier): differing from a bar, which is submerged at least at high tide (Shepard 1952). Typically a barrier is backed by a lagoon or

swamp that separates it from the mainland or from earlier barriers, but this is not essential: some barriers abut older land surfaces. A barrier may be from a few meters to more than a kilometer wide and up to 100 km long, attached to the mainland at one end (see Barrier Spit) or both ends, or interrupted by tidal inlets (see Barrier Island).

Short (1999, p. 307): “A barrier is defined as a shore-parallel sub-aerial and sub-aqueous accumulation of detrital sediment (sand/boulders) formed by waves, tides and aeolian processes.”

Barrier beach – A single, narrow (usually <200 m) elongated ridge built parallel to the coast, without surmounting dunes.

Barrier island – A barrier segment bordered by transverse gaps (tidal inlets, lagoon entrances, river outlets) which may be migratory and subject to closure; it usually bears beach ridges, dunes and associated swamps and minor lagoons. Barrier islands are typically 0.5–5.0 km wide, 1–100 km long and 6–100 m high.

Barrier lagoon – A lagoon extending roughly parallel to the coastline, behind a barrier or reef. See Coastal Lagoon.

Barrier reef – An elongated, narrow coral reef built up offshore and parallel to the coast, from which it is separated by a broad lagoon (which may be several kilometers wide), typically with extensive areas too deep for coral growth.

Barrier spit – A barrier attached at one end to the mainland, with or without recurves, and backed by a bay, lagoon or marshland (swamp).

Bay – A general term for a wide (typically >1 km) coastal re-entrant between two headlands, its seaward boundary generally wider than the extent of landward penetration. A small bay is termed a cove, a large bay a gulf. The term bay is also used in the United States for coastal waters largely or entirely cut off from the sea by spits and barrier islands (e.g., Netarts Bay on the Oregon coast) which elsewhere are called coastal lagoons, and for the ovoid depressions of uncertain origin that contain lakes or swamps on the coastal plain of South Carolina.

Beach – An accumulation on the shore of generally loose, unconsolidated sediment, ranging in size from very fine sand up to pebbles, cobbles, and occasionally boulders; often also containing shelly material. Conventionally, shores with silt or clay sediment are not regarded as beaches (but in some languages the terms beach and shore overlap, so that silt shores and clay shores may be translated as silt beaches and clay beaches). A distinction can sometimes be made between an upper beach, often swash-built, around high tide level, and a gentler lower beach exposed as the tide falls.

Short (1999, p. 3) defined a beach as “a wave-deposited accumulation of sediment lying between modal wave base and the upper swash limit where wave base is the maximum depth at which waves can transport beach material shoreward.” However, since wave base is conventionally where the water depth is half the wave length, this limit can be in water at least 50 m deep and perhaps several kilometers offshore. Most accounts of beaches admit their extension below lowest tide level and for some distance (usually not stated) beyond the breaker zone.

Beach berm – An ephemeral flat or landward-sloping step or terrace built on a beach face by swash action. On shingle beaches there may be several such berms, resulting from successive episodes of swash action at successively lower tidal levels. A berm built at the top of a beach may persist to become a Beach Ridge (*q.v.*).

Beach budget – The quantified gains and losses of sediment from a defined beach sector.

Beach compartment – A beach occupying a sector of coast bounded by rocky reefs, promontories or artificial structures such as breakwaters. See Coastal Sediment Compartment.

Beach cusps – Regular successions of half-saucer (crescentic) depressions opening seaward between cusped points in the swash zone on the upper beach face, the points being of coarser material (shingle or coarse sand) than the intervening hollows. Also known as Swash Cusps.

Beach drift (Beach drifting) – The zigzag movement of beach material along the shore in the swash zone when

Coastal sand barrier at Marlo, southeastern Australia.



Barrier reef with storm-tossed boulders.



Barrier spit, Lake Onoke, North Island, New Zealand.



Swash-built sand berm on the Ninety Mile Beach, southeastern Australia.



Shingle beach cusps at Ringstead, Dorset, England.



waves arrive at an angle to the shoreline, producing an oblique swash, followed by a Backwash (*q.v.*) that descends orthogonally from the beach face. The same waves generate a longshore current which carries sediment along the coast in the same direction in the nearshore zone. See Longshore Drifting.

Beach, drift-aligned (drift-dominated) – A beach with an orientation determined by the drifting of sediment alongshore in response to waves arriving at an angle to the coastline: they are aligned at an angle to the dominant direction of wave approach, with alignments parallel to the line of maximum longshore sediment flow, generated by obliquely incident (typically 40–50°) waves. Contrast Swash-Dominated Beach (*q.v.*), aligned parallel to the dominant waves.

Beach face – The seaward slope of a beach between the low tide line and the upper limit of wave swash.

Beach gravel – An American term describing beach sediment coarser than sand, ranging up to cobble size, and often (but not always) well-rounded: the British term shingle is similar.

Beach lobe – Roughly triangular or lobate protrusions from a beach that form and may migrate alongshore downdrift in response to oblique wave action.

Beach nourishment – The natural or artificial supply of sand or gravel to a beach. Also termed Beach Restoration, Beach Fill, and Beach Renourishment (*qq.v.*).

Beach, profile of equilibrium – A beach profile (concave, steepening upward past the high tide line and declining seaward to below the low tide line) that represents the attainment of an equilibrium with incident waves, whereby sediment gains and losses are balanced (Johnson 1919). A cyclic equilibrium is attained where a succession of beach profiles corresponds with changing wave regimes through a Cut-and-fill (*q.v.*) sequence, and a dynamic equilibrium where the beach profile is maintained even

though the beach may be prograding or retreating as the result of erosion.

Beach ridge – An elongated low ridge of beach material (sand, gravel, or shells) piled up above high tide level by swash action. Many beach ridges are surmounted by wind-deposited sand, which may develop into a Foredune (*q.v.*), but a beach ridge is strictly a feature built and shaped by wave action. Sandy beach ridges are typically 5–50 m wide, measured from bordering troughs (Swales, *q.v.*), shingle beaches usually narrower.

Beach ridge plain – A series of beach ridges formed successively, parallel, or roughly parallel to the coastline, and separated by elongated hollows or swales, each ridge marking a former position of the prograding coastline.

Beach rock – A sandstone layer within a beach, formed by interstitial precipitation of carbonates (calcite or aragonite) to cement the beach sand in the zone of fluctuating groundwater levels, usually where the sediments are strongly calcareous, as on shelly, calcarenite, or coralline beaches (Stoddart and Cann 1965). Where the material cemented consists of rounded pebbles the term beach conglomerate is used, while cementation of angular gravel yields beach breccia.

Beach scarp – A steep or vertical cliff cut in the beach face by large waves during a storm or tsunami.

Beach state – The state of a beach may be reflective (with a high proportion of wave energy reflected from beach face), dissipative (wave energy diminished through breaker and surf zones), or intermediate between these two.

Beach, swash-aligned (swash-dominated) – A beach formed where incoming waves are refracted into curved patterns that anticipate, and on arriving fit, its curved outline (Davies 1980), so that the beach is parallel to incoming wave crests (particularly refracted ocean swell).

Beach system – Beaches and the processes at work on them, with adjustments in morphology in response to energy inputs from waves, currents, tides, and winds.

Sandy beach lobe at Lang Lang,
Westernport Bay, Victoria,
Australia.



Storm-piled shingle beach ridge at
Redcliff Point, Dorset, England.



Shingle beach ridge plain at Hoed,
Denmark.



Beach rock exposed on a sandy shore at Gili Bidara, Lombok, Indonesia.



Scarp cut by storm waves, Ninety Mile Beach, southeastern Australia.



Bench – A flat or gently sloping rock ledge, terrace or platform, typically 5–50 m wide, but sometimes much wider, backed and fronted by steeper slopes. A structural bench coincides with the upper surfaces of a hard rock outcrop, formed where weaker overlying material has been removed by erosion, an erosional bench has been planed across tilted or folded rocks by erosion. A bench between high and low tide levels is called a Shore Platform (*q.v.*).

Berm – See Beach Berm.

Beveled cliff – See Slope-over-wall Coast.

Bioconstruction – Formed by organisms (e.g., algae, corals, vegetation) that trap sediment, including the products of their own decay (e.g., shells, peat).

Biodeposition – A little used term for sedimentation generated by organisms, such as coral reefs and algal rims, shelly deposits or peat.

Bioerosion – Removal of rock material by the physical and chemical processes associated with the activities and metabolism of plants and animals that inhabit the shore, especially on limestone coasts (Healy 1968).

Bioherm – A mound, bank, or (biogenic) reef built by sedentary organisms, such as corals, algae, foraminifera, molluscs or gastropods, or a reef built in the nearshore zone by oysters, mussels, gastropods or worms (serpulid and sabellariid reefs). Vermetid reefs are built by long, coiled gastropods.

Structural benches on Bay Cliff, Wonboyn, New South Wales, Australia.



Blowhole – A hole or fissure in the roof of a cave on a rocky coast, usually steeply sloping or vertical, through which geyser-like fountains of water and spray are forced by the intermittent release of compressed air trapped in a cave by incoming waves.

Blowout – A small saucer-shaped hollow or trough excavated in vegetated dunes by wind action, with an advancing nose of sand spilling downwind. See Parabolic Dune.

Blue hole – A submerged sinkhole in a coral reef, generally formed by solution when the reef was subaerially exposed during a low-sea-level-stage.

Bluff – A bold, steep, sometimes rounded coastal slope on which soil and vegetation conceal, or largely conceal, the underlying rock formations, in contrast with a cliff in which these formations are exposed. Bluffs may be termed abandoned, degraded, or fossil cliffs. The term seems to have originated in North America for a headland that was rounded rather than cliffed. There is confusion when the term bluff has been used as a place name for a feature that is actually a cliff. Bluffs may form where cliffs stop retreating (as when basal marine erosion is halted by emergence, the Accretion (*q.v.*) of a protective beach, or artificial protective structures such as seawalls or rock ramparts): they may then become degraded by subaerial processes to gentler slopes on which a soil forms and vegetation establishes. Bluffs are generally relatively stable, in comparison with receding cliffs, but the steep forested slopes on the Pacific coast of Oregon and Washington have receded as the result of occasional slumping.

Bombora – Australian term for a bank or reef that causes local raising and steepening of waves moving through shallow water, producing wave fronts that are good for surfing.

Boulder – A rock particle with a diameter exceeding 256 mm (or about 25 cm). See Granulometry.



Beveled cliff on glacial drift over vertical Chalk, Flamborough Head, Yorkshire, England.

Breakaway – A fracture at the top of a cliff behind an area where a rock mass has subsided or slid seaward, leaving a small scarp.

Breakers, Breaking waves – When waves break they produce (1) surging breakers, which are low and gentle



Reef formed by calcareous tubeworms (*Galeolaria*) at Beaumaris, Victoria, Australia.



Blowhole at Quobba, west coast, Western Australia.

waves until they sweep up a relatively steep beach, (2) plunging breakers, with fronts that curve over and crash (producing little swash but a strong backwash), (3) collapsing breakers that subside as they move toward the shore, and (4) spilling breakers, which are short and high, and produce foaming surf as the swash runs up a beach of gentle gradient (Galvin 1972). Alternatively, there are constructive breakers (which wash sediment up on to the beach) and destructive breakers (which cause beach erosion), depending on whether the swash and backwash achieve net shoreward or seaward movement of beach material.

Break-point bar – A bar built in the zone where waves break, deposition occurring where shoreward drifting beneath incoming waves meets seaward drifting by backwash, leaving a parallel trough to landward.

Breakwater – An artificial structure built into the sea, often curved, and designed to impede wave action so as to shelter a harbor or protect a stretch of coastline. The terms jetty and pier are sometimes used as synonyms.

Bund – A term used mainly in India and southeast Asia for an artificial embankment, usually of earth or gravel, built along the coastline or the banks of a river or estuary.

Cala, Calanque – A deep steep-sided marine inlet on a limestone coast: calas are found in the Balearic Islands and calanques on the coast of Provence.

Calcified seaweed, grit – Coarse angular sand or gravel produced by *Lithothamnion calcaireum*, which grows

extensively on some seafloor areas, as in the western part of the English Channel.

Calcilutite – A fine-grained (silt-sized) calcareous formation with an admixture of clay, often coherent enough to stand as a vertical cliff. The Port Campbell Limestone formation (Miocene) in Victoria, Australia, contains calcilutite deposits that stand in spectacular vertical cliffs.

Calcirudite – A calcareous conglomerate or breccia, consisting of broken or worn fragments of coral, shells, or limestone cemented by precipitated carbonates, often occurring in layers in Dune Calcarenite (*q.v.*).

Calcrete (Caliche) – A hard rock calcareous formation cemented by precipitated carbonates; a limestone or calcareous duricrust usually formed on or immediately below the land surface in an arid or semiarid environment by precipitation of carbonates derived from groundwater that moved upward to an evaporation level. Calcrete can occur as a horizon within dune stratigraphy (usually formed on an old land surface subsequently buried by younger dune sand) or as a caprock (carbonates having been delivered in rising groundwater and precipitated near the drying surface). Known as kunkar in Australia. See Dune Calcarenite.

Can – A fan of beach material washed through a low permeable coastal barrier by Storm Waves (Storm Surges) (*qq.v.*) to form a delta-like projection into a lagoon or on to backing marshland. There are good examples behind the shingle barrier of Chesil Beach, Dorset.

Bluffs on the southeast coast of Phillip Island, Victoria, Australia.



Breakaway on cliff top near Lyme Regis, Dorset, England.

Cape – A large, often rounded coastal protrusion, located where the coastline intersects a range of mountains, hills, or a plateau, usually where a drainage divide reaches the coast. However, some capes are low-lying (e.g., Cape Canaveral (Kennedy) and others on the American Atlantic coast).

Cay – A small low-lying depositional island of coralline sand or gravel (shingle cay) built up just above high tide level by wave action on a reef flat, usually toward the lee side. In the Caribbean a cay is termed a key.

Chenier – A long, low-lying narrow strip of sand, often shelly and typically up to 3 m high and 40–400 m wide, deposited in the form of wave-built beach ridge on a swampy (peat and clay), deltaic, or alluvial coastal plain by wave action. Many cheniers contain shelly sand and gravel. In Louisiana such ridges often have vegetation dominated by oak trees (*chêne*, hence *chênière*), emphasizing the contrast with the adjacent peat or clay terrain (Russell and Howe 1935). Cheniers on the wide coastal plain east of Darwin, Australia, have pandanus palms: had they been originally studied here they may have been called pandaniers. Most cheniers have been deposited at the limit of storm surge swash, and may be termed transgressive, but some may have formed along the coast during a brief phase of raised sea level, followed by progradation that leaves them stranded inland: these may be termed regressive.

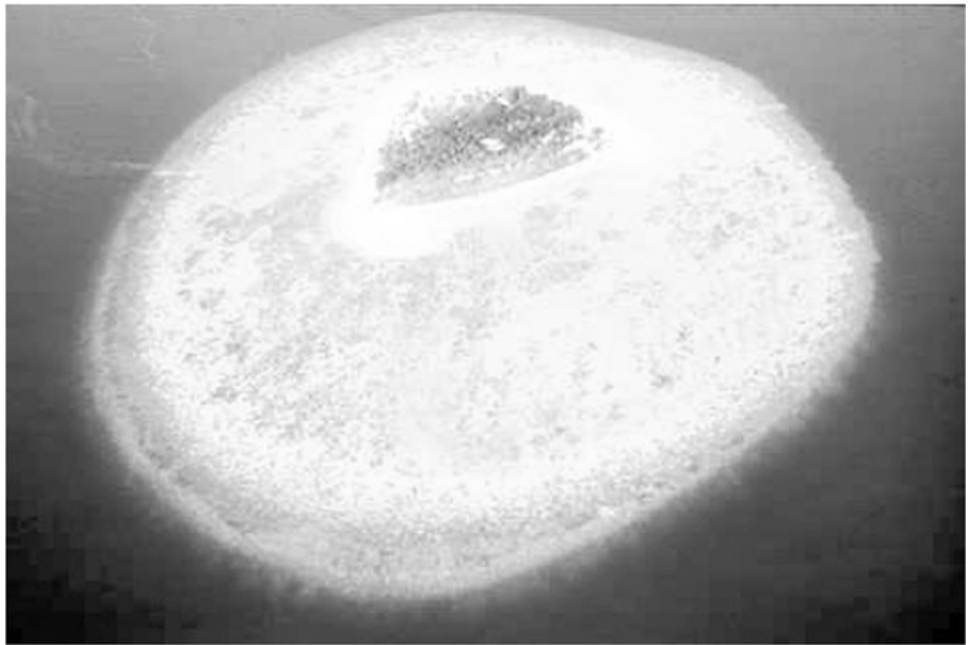
A chenier plain is a coastal plain with several cheniers scattered across it, usually parallel or subparallel to the coastline. Cheniers may pass laterally into beach ridges where the fine-grained substrate becomes sandy or gravelly.

Chine – A deep and narrow ravine cut into soft rock on a cliffed coast by a stream descending steeply to the shore, often over waterfalls. A term used in the Isle of Wight and on the Hampshire coast in southern England.

Calanque at Wied-il-Chazri, Cozo, Malta.



Coral cay in the Maldiv Islands, Indian Ocean.



Clastic – Consisting of broken and transported fragments of pre-existing rock. The term Coarse clastic beach has sometimes been applied to a beach of gravel or Shingle (*q.v.*).

Clay – A sediment consisting of particles with a diameter smaller than 1/256 (about 0.004) mm.

Cliff – A steep (usually $>40^\circ$, often vertical and sometimes overhanging) coastal slope cut into (and thus exposing) rock formations, produced by basal marine erosion (undercutting), but occasionally by faulting or earlier

fluvial or glacial erosion. Cliffs cut in unconsolidated formations are sometimes known as Earth Cliffs (*q.v.*), and there are also Ice Cliffs (*q.v.*) at the seaward terminations of glaciers and ice sheets. Cliffs rising to 100–500 m above sea level are termed high cliffs, while those exceeding 500 m (as in Peru and Western Ireland) are termed megacliffs (Guilcher 1966): the cliff at Enniberg on the north coast of Vidoy (Faerö Islands) is 725 m high. Coastal cliffs are generally receding as the result of marine

Sandy chenier within mangroves,
Karembé, New Caledonia.



Whale Chine, Isle of Wight,
England.



erosion at their base, accompanied by subaerial erosion of the cliff face.

Cliff fall (Rock fall) – The collapse of the face of a rocky cliff into an apron of debris. On some cliffs columns of rock bordered by vertical joints may topple on to the shore.

Clifflet – See Microcliff.

Cliff-top dunes – Usually found where sand blown from a beach moved up and over a cliff, but the link between the beach and the dune has been removed by erosion, exposing the cliff; occasionally the dunes have arrived from inland (Jennings 1967).

Coast – A zone of varying width where the land meets the sea, and where the lithosphere, hydrosphere, and atmosphere meet and interact. Some use the term as a synonym for shore, but generally it is taken to include the land margin, extending inland to the limit of penetration of marine processes or to the first major change in landform features, such as a rising slope, and the nearshore zone, out at least to the line where waves break. A wider definition includes the whole zone between the highest and lowest coastlines related to sea-level changes during the Quaternary (*q.v.*).



Cliff fall and scar on Chalk coast, North Foreland, Kent, England.

In American literature the coast is sometimes elaborated to the coastal zone.

Coastal barrier – See Barrier.

Coastal dimensions – Coastal morphology may be classified as follows:

First-order features – about 1000 km long, 100 km wide, and 10 km high (e.g., continental coasts, related to global tectonics).

Second-order features are about 100 km long, 10 km wide, and 1 km high (e.g., deltas, fiords).

Third-order features about 10 km long, 1 km wide, and 100 m high (e.g., coastal barriers).

Fourth-order features are about 1 km long, 100 m wide, and 10 m high (e.g., foredunes).

Fifth-order features are about 100 m long, 10 m wide, and 1 m high (e.g., beach berms, shore platforms, sand bars).

Sixth-order features are about 10 m long, 1 m wide, and 10 cm high (e.g., beach cusps).

Seventh-order features are about 1 m long, 10 cm wide, and 1 cm high (e.g., current ripples).

In each case the dimension given should be regarded as being within a range of from 50% to five times the figure given (i.e., Sixth-order features 5–50 m long, 0.5–5 m wide, and 5–50 cm high).

Coastal dunes – Hills, banks, or ridges of aeolian (wind-blown) sand backing present or former coastlines.

Coastal erratics – Large shore rocks that came from outcrops elsewhere, delivered to their present position on the coast or on the shore, usually by glaciers or icebergs.

Coastal gorge – See Geo.

Coastal lagoon (Barrier lagoon) – A shallow, often brackish (estuarine) water body formed where a coastal inlet, river mouth or embayment has been enclosed (or almost enclosed) behind a depositional barrier or barrier spit. Known as an *étang* in France, a *half* in Germany, an *estero* in Portugal, a *liman* on the Black Sea coast and a *pond* in New England. A distinction is made between coastal lagoons enclosed by depositional barriers and lagoons backed by coral atolls, or standing between a barrier reef and the mainland coast.

Coastal land reclamation – The making of new land by enclosing or filling shore and nearshore areas. An alternative term *land claim* has been suggested, because this is not strictly reclaiming, but *land claim* risks confusion with territorial claims by political groups.

Coastal landslide – The movement of a mass of rock, earth or debris down a coastal slope. Also known as a *landslip* (Lyell 1833). Where the coastal rocks are fine-grained, the downslope movement may be in the form of a mudflow lubricated by exuded groundwater or surface runoff after heavy rain or melting snow or ice.

Coastal plain – Low-lying coastal terrain, usually of depositional origin, but occasionally consisting of planed-off rock outcrops (see *Strandflat*), between the coastline and rising ground in the hinterland.

Coastal sediment compartment – A sector of coast within which sediment is largely or completely confined, delimited by rocky reefs, promontories, tidal inlets, river mouths, or artificial structures such as breakwaters (Davies 1974). See *Beach Compartment*.

Coastal tors – Outcrops of harder or more massive rock protruding as buttresses (coastal tors) from a coastal slope.

Coastal waters – A general term for the sea area adjacent to the coast, comprising the nearshore and offshore zones. The seaward limit is usually indefinite, and an arbitrary distance (such as 3 nautical miles) has been used in Law of the Sea schedules.

Coastline – The edge of the land at the limit of normal high spring tides; the subaerial land margin, often marked by the seaward boundary of terrestrial vegetation. On cliffed coasts it is taken as the cliff foot at high spring tide level. Use of the term *Shoreline* (*q.v.*) as a synonym for *Coast* or *Coastline* is vague and misleading, and should be avoided – shorelines move to and fro as the tides rise and fall, whereas coastlines are related to the high-spring-tide shoreline, and thus submerged only in exceptional circumstances (e.g., Storm Surges, *q.v.*).

Cobble – A rock particle with a diameter between 64 and 256 mm. See *Granulometry*.

Giant's Rock, a coastal erratic on the shore near Porthleven, Cornwall, England.



Coastal landslide at Blackgang, Isle of Wight, England.



Coastal tor (buttress) at Trewalvas Head, Cornwall, England.



Colk – A relatively deep circular or oval depression in the seafloor (or on the floor of an estuary or lagoon) excavated and kept clear of sedimentation by locally strong current action (including the siphoning that can occur beneath ice). Also known as a tidal colk or scour hole. Originally a pothole on a river bed.

Compound spit – A recurved spit with several recurves on the inner shore marking former terminations.

Concentric (Contraction) ridges – A series of small parallel beach ridges formed successively on the shore of a contracting shallow lake or lagoon by intermittent deposition along the margins of a prograding salt marsh, as described from Lake Reeve, Victoria, Australia, by Jenkin (1966).

Contraposed coast – A discordant coast developed on previously concealed rock formations where marine erosion has removed bordering and overlying weaker deposits (Clapp 1913). Examples of contraposed coasts include sectors where Archaean gneiss has been exhumed from a Pleistocene dune calcarenite cover on the west coast of Eyre Peninsula, South Australia, and where Palaeozoic rocks have been exposed by the removal of a fringe of glacial drift on the coasts of the Llyn Peninsula in North Wales. In southwest England the slope-over-wall profiles formed where a mantle of Pleistocene periglacial Head has been undercut by marine erosion to expose Devonian rocks could also be classified as contraposed.

Coral garden – An open structure built by branching (e.g., staghorn) corals.

Coral rampart – A depositional ridge of coral fragments built near the windward margin of a coral reef by storm waves or surges.

Coral reefs – A structure built in the sea by corals, which form the reef framework, together with algae and other organisms (such as algae, molluscs, crinoids, bryozoans) and precipitated carbonates to make rock formations sufficiently resistant to withstand normal wave action.

Corrosion – Dissolving of rock or minerals by chemical action in water (rain, sea, or spray).

Currents – Movements of water generated by one or more of the following: density currents occur where water of higher specific gravity (colder or more saline) moves to displace water of lower specific gravity; discharge (fluvial jet) currents occur where a river flows into the sea; ebb and flow (flood) currents are generated by falling and rising tides; wave-generated currents such as the longshore currents that develop when waves arrive at an angle to the shoreline; Rip Currents (*q.v.*) flow back into the sea through breaking waves at intervals along the shore; wind-generated currents flow in the direction of the wind; and ocean currents are slow mass movements of water in response to variations in water temperature and salinity, atmospheric pressure and wind stress.

Cusate bar – A triangular depositional bank of sand or shingle, submerged, at least at high tide, extending out from the coast with straight or concave shores that meet in a seaward point, and enclosing a depression occupied by a lagoon at low tide.

Cusate barrier – Similar to a cusate bar, except that it has been built above high tide level.

Cusate foreland – A triangular depositional area of sand or shingle with straight or concave shores extending out to a seaward point, with multiple beach ridges marking stages in progradation on one or both flanks. Cusate forelands may migrate by erosion of one flank and the drifting of beach material to accrete on the other. Known in Britain as a ness (Dungeness and Duddon Ness).

Cusate spit – A triangular spit that projects from the coast with straight or concave shores extending out to a seaward point. Often formed in the lee of an islet, stack, reef or shoal, or on the shores of a bay or lagoon as a result of convergent wave refraction. Formerly known as a cusate bar, but this term is not applicable to a feature built above high tide level (see Bar).

Cut-and-fill – The cyclic sequence of changes on a beach profile resulting from erosion by storm waves (cut) and subsequent restoration by constructive waves (fill).

Delta – A depositional landform produced by sedimentation at and around the mouth of a river. Deltas usually protrude from the coast, and are typically triangular, named from the Greek letter “delta” applied to the Nile delta by Herodotus in ancient times, but they may alternatively be Digitate (finger-like, as in the Mississippi delta), Cusate (Tiber delta), Arcuate or rounded (Niger delta), or Lobate (Rhône delta), in shape.

Dissipative beach – A beach where wave energy is greatly diminished as spilling breakers move in through shallow nearshore water.

Distributary – One of a number of channels formed in a delta region where the river branches downstream.

Drift-aligned beach – See Beach, Drift-Aligned.

Drowned valley-mouth – See Ria.

Dune – A mound or ridge of unconsolidated wind-blown sediment, usually sand but occasionally silt or clay (where the source area is a dry mudflat: crescent-shaped silt or clay dunes on the lee shore of a lake or lagoon are known as lunettes).

Dune calcarenite – A generally consolidated aeolian sandstone lithified by the cementation or partial cementation of dune sand by secondary internal precipitation of carbonates from groundwater. The proportions of carbonate vary, but are typically at least 50%, the dune sand being calcareous or partly noncalcareous (quartzose sandstones with less than 50% carbonate are termed quartz-arenites). Dune calcarenite is usually of Pleistocene age, and may be overlain by unconsolidated Holocene dune topography. Dune calcarenite is also known as Aeolian Calcarenite (*q.v.*), Aeolianite, Calcareous aeolianite, Dune limestone or Dune sandstone; American spelling: eolianite.

Dune swale – A hollow within dune topography, especially between parallel dune ridges. Wet swales (often with marsh vegetation) are termed *slacks* and deeper hollows may be seasonally or permanently occupied by *dune lakes*.

Earth cliff – A term used by May (1977) for a cliff cut in soft rock formations (sand, clay, and chalk) and subject to rapid recession and instability leading to recurrent mass movements.

Edge waves – Standing oscillations that develop at right angles to the coastline as the result of resonance between waves approaching the shore and waves reflected from it.

Coral garden, Great Barrier Reef, Australia.



Dungeness: a cusped foreland in southeastern England.



Emerged beach – A beach that stands above the level at which it originally formed, on or behind the present shore. Known in the earlier literature as a Raised Beach (*q.v.*), when it was thought that upward land movement was necessary to produce such a feature, but it can also be formed as a result of a lowering of sea level, or some combination of land and sea-level change that has left the sea at a relatively lower level.

Emerged shore platform – A shore platform that stands above the level at which it originally formed (often with an emerged beach), on a coast where the land has been

uplifted, sea level has fallen, or some combination of land and sea-level change that has left the sea at a relatively lower level.

Emergence – A rise in the level of the land relative to the sea, achieved by actual uplift of the land, a lowering of sea level, or some combination of land and sea-level change that leaves the sea at a relatively lower level. Also known as a negative change (or fall) in base level. An emerged coast (or feature) is one that stands at a higher level relative to the sea than when it originally formed; an emerging coast is one actually rising relative to sea level.

Cuspate spit, Moreton Island, Queensland, Australia.



Dune calcarenite cliff, Jubilee Point, Victoria, Australia.



Epeirogenic movement – Upward or downward tectonic movements of large (continental) land masses.

Escarpment cliff (bluff) – A steep coastal slope cut across rock formations that are horizontal, or dipping landward.

Estuary – The seaward end of a river, opening toward the sea, typically through a funnel-shaped inlet, and usually subject to tidal movements and incursions of salt water from the sea.

Estuary threshold – A bank of inwashed sand or gravel at the mouth of an estuary.

Eustasy (Eustatic movements) – Worldwide movements of sea level resulting from changes in the volume of water in the ocean basins. Such changes have occurred as a result of the waxing and waning of the Earth's glaciers, ice sheets, and snowfields (glacio-eustatic movements), but similar changes have taken place as a result of modification of the shape and capacity of the ocean basins by deposition of sediment (sedimento-eustatic movements), submarine vulcanicity (volcanic-eustatic movements) or tectonic deformation (tectono-eustatic movements). There are also

Emerged beach, Falmouth Bay, Cornwall.



Emerged shore platform (raised in 1855 earthquake), Wellington, New Zealand.

steric changes due to expansion or contraction of the oceans with rising or falling temperatures.

Fault coast – A steep or cliffed coast produced by faulting, where the seaward slope coincides with the plane of the fault, along which the land has been raised. Some coasts were initiated as fault coasts, but have been cut back by marine erosion and now stand landward of the fault.

Fault-line coast – A steep or cliffed coast following a fault line, where the seaward slope has been formed by differential erosion of rock formations juxtaposed by prior faulting.

Ferricrete – A sedimentary rock formation indurated by the precipitation of iron oxides or other iron compounds, usually derived from percolating groundwater.

Fetch – The distance of open water across which the wind generates waves approaching a coastline from a particular direction.

Fiard (Fjard) – An inlet formed by marine submergence of a river valley (or wide, shallow valley excavated by ice movement) incised in low-lying glaciated rocky terrain. Known in Scotland as a firth.

Fiord (Fjord) – A long, deep steep-sided marine inlet formed by submergence of part of a valley (U-shaped glacial trough) previously shaped by a glacier and incised into coastal uplands. Known in Scotland as a sea loch.

Fitting boulders – Interlocking accumulations of blocks and boulders on the shore, resulting from the jostling of the rocks by wave agitation and their consequent abrasion until they have become mutually worn into a complex three-dimensional jigsaw pattern.

Flandrian marine transgression – The sea-level rise that began about 18,000 years ago, in the Late Pleistocene, and continued into the Holocene. This transgression was originally defined in Flanders, northeast France and

Escarpment cliff, Ballard Down, Dorset, England.



A fiard near Hjortholm, Denmark.



southern Belgium (Dubois 1924). See Late Quaternary Marine Transgression (*q.v.*).

Foredune – A ridge of wind-blown sand at the back of a beach, parallel to the coastline, and retained by vegetation. Some foredunes originate as a result of wind deposition of sand on a beach ridge (*q.v.*), but others may begin to form along an upper beach strandline that persists long enough for plants to germinate and initiate sand trapping. The terms primary dune and frontal dune are synonyms for foredune in Australia.

Foreshore – The shore zone, between high and low tide lines. In some countries there is confusion because the term has been used for the fringe of the land, which is really the backshore.

Fringing reef – A structure built adjacent to the coast by coral and associated organisms. A distinction may be made

between a shore reef, built out as a prograded terrace from the coastline, and an attached reef, which originated a short distance offshore and became linked to the coastline by subsequent deposition. Some fringing reefs decline gently landward, passing beneath a shallow (ca. 1 m) lagoon or moat, sometimes called a boat channel.

Frost shattering – The disintegration of rock surfaces, notably on cliffs and shore outcrops, as a result of repeated expansion and contraction of the rock by freezing and thawing, particularly where contained moisture forms ice crystals as temperature ranges through 0–4°C. Also termed thermal abrasion.

Gat – A channel cut or maintained across a bar by waves or (more often) currents.

Geo – A Scottish term for a long, deep, and narrow steep-sided inlet (coastal gorge or chasm) cut into rocky cliffs

Fiord at Milford Sound, South Island, New Zealand.



Foredune at Seaspray, Ninety Mile Beach, Victoria, Australia.



by marine erosion, usually along structural planes of division (joints, faults, bedding planes, cleavage planes). Known as a zawn or gut in Cornwall. A distinction may be made between geos, excavated entirely by marine erosion, and gorges that are the mouths of deeply incised steep-sided valleys invaded by the sea during the later stages of the Flandrian Marine Transgression (*q.v.*).

Geological column – The sequence of rock formations arranged by age (my = million years):

Geological structure – See Structure

Giant waves – Exceptionally high waves generated by major disturbances (earthquakes, landslides, volcanic eruptions) on the coast or seafloor. See Tsunami.

Grain Size Categories – The Wentworth scale of particle diameters. The ϕ scale is based on the negative logarithm (to base 2) of the particle diameter in millimetres [$\phi = \log_2 d$], so that coarser particles have negative values.

An alternative is a decimal scale centered on 2 mm, in which sand ranges from 0.2 to 2.0 mm, but this does not correspond well with generally perceived grain size categories. The sand range excludes sediment that would be generally classified as fine to very fine sand, and the coarser (>2.0 mm) and finer (<0.2 mm) divisions do not match widely accepted categories of pebbles and cobbles or of silt and clay.

Fringing reef on the southeast coast of Bali, Indonesia.



Quaternary period	Holocene (Recent)	10,000 years
	Pleistocene Epoch	2.3 my
Tertiary period	Pliocene Epoch	5 my
	Miocene Epoch	23 my
	Oligocene Epoch	36 my
	Eocene Epoch	53 my
	Palaeocene Epoch	65 my
Mesozoic Era	Cretaceous Period	144 my
	Jurassic Period	213 my
	Triassic Period	248 my
Palaeozoic Era	Permian Period	290 my
	Carboniferous Period	360 my
	Devonian Period	405 my
	Silurian Period	436 my
	Ordovician Period	510 my
Pre-Cambrian Era	Cambrian Period	560 my

Wentworth scale category	Particle diameter (mm)	Ø scale
Boulders	>256	below -8Ø
Cobbles	64–256	-6Ø to -8Ø
Pebbles	4–64	-2Ø to -6Ø
Granules	2–4	-1Ø to -2Ø
Very coarse sand	1–2	0Ø to -1Ø
Coarse sand	1/2–1	1Ø to 0Ø
Medium sand	1/4–1/2	2Ø to 1Ø
Fine sand	1/8–1/4	3Ø to 2Ø
Very fine sand	1/16–1/8	4Ø to 3Ø
Silt	1/256–1/16	8Ø to 4Ø
Clay	<1/256	above 8Ø

Granule – A rock particle with a diameter between 2 and 4 mm. See Granulometry.

Granulometry – The measurement and classification of the size of sediment grains.

Groyne – A wall built out at right angles from the coastline, intended to intercept drifting beach material. American spelling: groin.

Gulch – A deep and narrow channel cut by abrasion into a cliff or shore platform or across a rocky shore, often extending below low tide level.

Hairpin dune – An elongated, narrow blowout or parabolic dune with parallel trailing arms, well developed in northeastern Tasmania.

Hanging valley – A valley truncated by cliff recession, so that the stream pours out as a coastal waterfall.

Headland – A high protrusion of the land into the sea, usually cliffed. Generally smaller than a Cape or Promontory (*qq.v.*).

Holocene epoch – The last of the geological epochs, which began about 10,000 years ago.

Holocene marine transgression – See Late Quaternary Marine Transgression or Flandrian Marine Transgression.

Honeycomb weathering (Alveolar weathering) – A process producing an intricate pattern of small cell-like cavities on a rock surface, often penetrating a harder (or indurated) crust.

Humate – A consolidated and indurated sandrock formed within a sandy formation (Dunes, Beach Ridges, *qq.v.*) by interstitial precipitation of iron oxides (derived from the thin coating of quartz sand grains by ferruginous material, responsible for their initial yellow or brown color, mobilized by percolating groundwater) and precipitation of organic matter (washed down from surface soil and decaying vegetation) within the zone where the water table rises and falls (seasonally or irregularly). Also known as coffee rock or hardpan.

Honeycomb weathering on Cretaceous sandstone, Otways coast, Victoria, Australia.



Cliffs cut in humate, Chatham Islands, New Zealand.



Hutberge – A high, steep-sided hill rising abruptly from a Strandflat (*q.v.*), formed as a residual island in a shallow sea subject to intensive freeze-thaw processes accompanying tidal alternations. Similar in form, but not origin, to Old Hat Islands (*q.v.*).

Hydro-isostasy – Vertical movements of a coast and continental shelf in response to loading and unloading of water as sea levels rise and fall.

Ice coast/Ice cliffs – Steep cliffs up to 50 m high, formed where glaciers or ice sheets (ice shelves) reach the sea, and from which floes and icebergs become separated

during summer melting. Where the ice is afloat the cliff is called an ice front; where it is grounded, an ice wall.

Ice rafting – The delivery of material (rocks, beach sediments, marsh fragments, driftwood) to the shore on or within layers of ice (often in icebergs detached from a glacier front), to be deposited when the ice melts in summer. See also Coastal Erratics.

Induration – The hardening of a rock surface by precipitation of material (carbonates, iron compounds, silica) from groundwater exudations, forming a resistant crust (case-hardening).

Intermediate beach – A beach state transitional between a dissipative and a Reflective Beach (*q.v.*).

Isostasy – An equilibrium between an area of the Earth's crust floating and the underlying plastic mantle, whereby areas loaded with sediment, volcanic deposits or ice subside, and areas unloaded (e.g., when an ice cover melts) rise (isostatic rebound).

Jetty – A solid structure built out more or less at right angles to the coastline or on either side of a river mouth or lagoon entrance. The terms breakwater and pier are sometimes used as synonyms.

Karst coast – A coast with landforms shaped by solution processes, notably on limestone, where the dissolving of the rock leads to the formation of surface depressions, sinkholes, caves, and underground drainage.

Klint – A term used for a cliff (generally in limestone) in the Baltic region. In some countries, notably Estonia, it describes an active cliff, but more generally it indicates an inland bluff that was an active cliff when sea level was higher.

Landlocked – An area of water (usually a bay or lagoon) surrounded or nearly enclosed by land.

Late Quaternary marine transgression – The worldwide rise of sea level that began about 18,000 years ago, when sea stood between 100 and 140 m below its present level, and the continental shelves were subaerially exposed, as the result of global warming and the release of water from melting glaciers, ice sheets, and snow fields. It came to an end about 6,000 years ago, but the history of sea-level change has been complicated on many coasts by accompanying and continuing uplift or depression of coastal land margins. See Flandrian Marine Transgression. Sometimes called the Holocene marine transgression, but the term Late Quaternary Marine Transgression (*q.v.*) is more accurate as the sea-level rise began late in Pleistocene times and culminated within the past 10,000 years (Holocene).

Lateral grading – A gradual change in the caliber of beach sediment along the shore. Grading is a condition; sorting and attrition are processes that may lead to a beach becoming graded.

Longshore current – The flow of water along the shore or nearshore as the result of oblique waves, often augmented by wind-driven and tidal currents.

Longshore (Littoral) drifting – The movement of beach sediment along the shore (and Nearshore, *q.v.*) by waves arriving at an angle to the coastline (Beach Drifting, *q.v.*) and by currents generated by such waves (nearshore drifting). Also known as Shore Drift (*q.v.*).

Lobate (Looped) bar, barrier – A depositional feature curving out from the coastline and back again, in such a way as to enclose a lagoon or swamp. A lobate bar is submerged at least at high tide, and a lobate barrier has been built up above high tide level.

Log-spiral beach – See Zetaform Beach.

Low islands (Low wooded islands) – Small low-lying depositional islands of coralline sediment, comprising a leeward sand Cay (*q.v.*), mainly of sand, a shingle rampart on the windward side, and an intervening depression occupied by a lagoon with mangroves.

Machair – A low flat or hummocky plain of calcareous sand, generally formed on the landward side of a coastal dune, as on the coasts of Scotland and Ireland.

Macrotidal – Where mean spring tide range is 4–6 m.

Mangrove – A type or community of halophytic trees and shrubs that can grow on shores sheltered from strong wave action (bays, inlets, delta, estuary, and lagoon shores) in the intertidal zone, subject to regular or frequent submergence of their root systems by brackish or seawater. Mangroves are found mainly on tropical and subtropical coasts, but extend locally into temperate latitudes.

Megaripple – A large form or Ripple (*q.v.*) or sand wave with an amplitude of 0.1–1.0 m and length between 1 and 100 m. See Ridge and runnel.

Megatidal – Where mean spring tide range exceeds 6 m.

Mesa – A flat-topped, steep-sided residual hill. The term has also been used to describe small upstanding flat-topped rocks on a shore platform.

Mesotidal – Where mean spring tide range is between 2 and 4 m.

Microatoll – A circular organic reef structure (diameter 1–6 m) consisting of a raised rim built by coral (usually porites), algae and other organisms, surrounding a shallow depression on the shore or on a coral reef.

Microcliff – A small cliff, generally less than a meter high, found on the seaward margins of salt marsh and mangrove terraces, and sometimes in intertidal mudflats. Also known as a Cliflet (*q.v.*).

Microtidal – Where mean spring tide range is less than 2 m.

Mud – A sticky fine-grained sediment (silt, clay, sometimes with organic matter).

Mudflat – A relatively level unvegetated area of fine sediment on the shore, especially in sheltered inlets, estuaries, or tidal lagoons. Intertidal mudflats are commonly exposed seaward or salt marshes of mangroves at low tide, and supratidal mudflats (Tanns in West Africa) occur landward of salt marshes or mangroves on arid or semiarid coasts.

Mudrock – A massive or blocky rock composed of indurated fine-grained sediment (silt and clay), also termed siltstone or claystone, differing from a shale in being nonfissile.

Multicausality – Where similar coastal landforms (e.g., coastal barriers, cusped forelands, laterally graded beaches) may be produced in different ways (Schwartz 1971).

Muricate weathering – the intricate pitting of coastal rock surfaces in the spray zone, especially on limestones, calcareous sandstones, and dune calcarenites, as a result of physical processes (wetting and drying), chemical processes (corrosion by sea spray, i.e., aerated sea water; growth of salt crystals), and biological processes (corrosion by algae and other organisms; scraping and browsing by shore fauna).

Mushroom rock – A table-like stack that has been marginally undercut by abrasion, solution, or bioerosion so that it is surrounded by a notch and visor, and stands above a narrower pedestal.

Low Isles, low wooded islands off Port Douglas, Queensland, Australia.



Megaripples in the Bay of Fundy, Canada.



Natural arch – A tunnel extending through a headland, island or stack, beneath a connecting bridge, usually formed where a cave has been enlarged by abrasion or solution. Also known as a Sea Arch (*q.v.*).

Neap tide – A diminished tide range when the sun and the moon are at right angles in relation to the earth, so that their gravitational effects are not combined.

Nearshore – The shallow water zone between the low tide line and the line where waves begin to break; a zone that migrates to and fro as the tide rises and falls.

Nearshore water circulation – An association of processes that operates as wave move through the

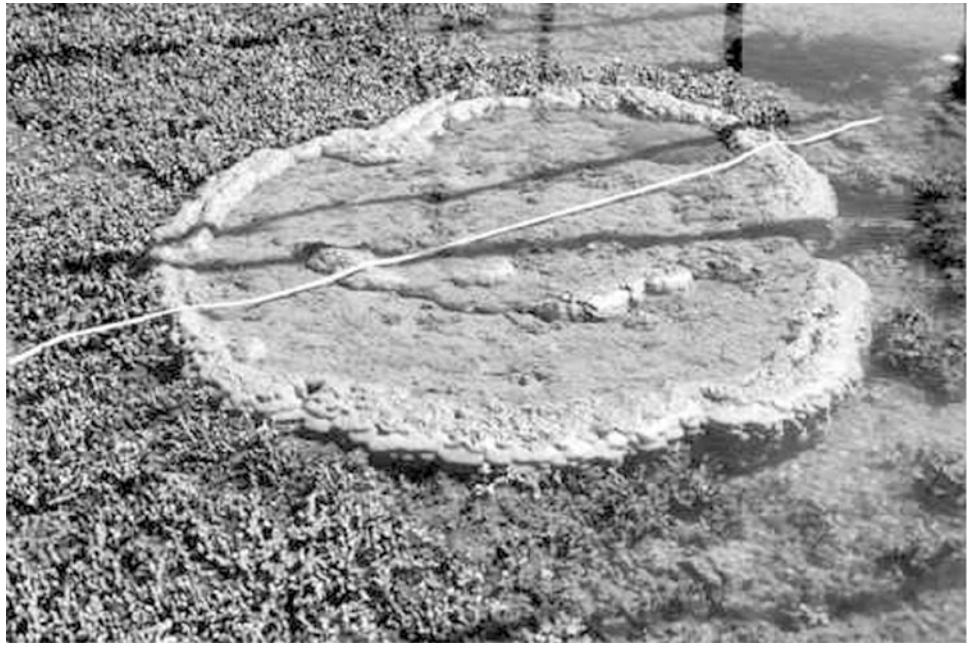
nearshore zone to break on the Shore or Beach, generating Swash and Backwash (*qq.v.*).

Neotectonic – Movements that have produced existing landforms (e.g., fault scarps) and associated structures, generally within Quaternary (*q.v.*) Times.

Notch (Nip) – A narrow hollow excavated along the base of a cliff near high tide level by abrasion, solution or bioerosion, often with an overhanging rock visor (protruding ledge of rock). See Abrasion Notch, Solution Notch.

Old Hat island – A stack or islet surrounded by a flat or gently sloping shore platform, formed where much of a soft

Microatoll, White Lady Bay,
Magnetic Island, Queensland,
Australia.



Mushroom Rock on Pleistocene
dune calcarenite, Sorrento,
Victoria, Australia.



or weathered rock formation has been removed by erosion above a harder horizontal intertidal rock layer, which persists as a structural bench. Originally described from North Island, New Zealand.

Open coast – A coast unprotected by islands, promontories or reefs, and so exposed to the full force of wave action.

Overwash (Washover) – The washing of sediment over the crest of a beach or coastal barrier by exceptionally strong wave swash to form a depositional fan on the landward side. See Washover Fan.

Paired spits – Spits on either side of a coastal inlet, river mouth or lagoon entrance, or protruding toward each other between two islands or between the mainland and an offshore island. They have developed either by convergent longshore drifting or the breaching of a former coastal barrier.

Pan – A shallow steep-sided and flat-floored natural depression on a shore platform or in a salt marsh (see Salt Pan). A soil pan is a crust or subsurface horizon of compacted or indurated sediment (see Humate).

Parabolic dune – A dune with an advancing convex nose of spilling sand and trailing (roughly parallel) arms of

Natural arch, La Marine Porte,
Etretat, France.



Old Hat island, Taylor Island,
Paihia, New Zealand.



partly vegetated sand on either side of an elongated low corridor formed or maintained by deflation. Typically its axial length is more than three times its mean width.

Parallel dunes – A succession of foredunes developed on a prograding coast (usually surmounting Beach Ridges, *q.v.*).

Pebble – A rock particle with a diameter between 4 and 64 mm. See Granulometry.

Periglacial (Paraglacial) – Processes and environments found beyond the limits of glaciation, typified by freezing and thawing, and frequent accumulation and melting of snow.

Pier – An open structure on multiple supports, usually designed to permit ships to berth: beneath it waves, currents and drifting sediment pass almost unimpeded. The term Jetty (*q.v.*) is sometimes used as a synonym. Occasionally these terms are used to describe solid stone structures.

Pit, Pitting – superficial indentations (typically up to 5 mm in diameter), etched on a rock surface by Weathering or Abrasion (*qq.v.*). See Muricate Weathering.

Plunging cliff – A steep or vertical cliff that descends into deep water inshore without any intervening Shore Platform, Rocky Shore, or Beach (*qq.v.*).

Paired spits, Shoal Inlet, Victoria, Australia.



Parabolic dune near Lakes Entrance, Victoria, Australia.



Point – A small protrusion of the land into the sea, usually sharp, tapering and low-lying, but sometimes a high headland (Hartland Point, Devonshire).

Progradation – The building seaward of a coastline by deposition of sediment, as on a beach or a dune, or where a marsh or mangrove shoreline advances.

Promontory – A coastal protrusion or headland, high and bordered by cliffs or bluffs, usually smaller than a Cape (*q.v.*).

Quaternary – The geological period which began about 2.3 million years ago. It comprises the Pleistocene epoch

(2.3 million years ago to 10,000 years ago) and the Holocene or Recent epoch (the last 10,000 years).

Raised beach – A beach that has been uplifted by tectonic movements to stand above the level at which it originally formed, on or behind the present shore. See Emerged Beach.

Rampart – A slight rise or wall-like ridge up to 2 m high toward the seaward edge of a shore platform or reef, or toward the crest of a cliff. Some ramparts are residual rock features that have escaped weathering and erosion; others are constructional, formed by the growth of organisms

Parallel dunes near Robe, South Australia.



Plunging cliff near Milford Sound, South Island, New Zealand.

(see Algal Rim) or depositional, as on the coaming found on some cliff crests. They may consist of boulders, cobbles, pebbles or sand, generally angular and often cemented. See Coral Rampart.

Recurved spit – A spit that ends in a landward hook or recurve (e.g., Sandy Hook, New Jersey).

Reef – A bank, ridge or mound with a rocky structure, either an eroded rock formation or built by organisms such as corals and algae, usually irregular in outline (but often flat-topped), and not moved by waves or current action.

Reflective beach – A beach where wave energy is partly reflected seaward as plunging breakers move in through relatively deep nearshore water.

Relict features – Features that developed under different environmental conditions (climate, vegetation, sea level) in the past, and have persisted in the present coastal landscape, having not yet been destroyed by modern processes.

Reliction – The exposure of land as a result of seafloor emergence during a slow or gradual withdrawal of the relative fall in sea level, as around the Caspian Sea between 1930 and 1977.

Ria – A long, narrow, often branching inlet formed by marine submergence of parts of a river valley that had previously been incised to a lower sea level: a drowned valley-mouth. Cotton (1956) separated *ria sensu lato*, thus defined, from *ria sensu stricto*, where the inlet runs parallel to geological outcrops that run at right angles to the coastline, as in the Rias of Galicia, northwest Spain, which Richthofen (1886) quoted in the original definition. Some of the Galician rias are wide and deep, and the valleys that have been drowned may have been shaped partly by tectonic subsidence or by the recession of bordering scarps.

Ridge and runnel – Several subdued bars and troughs running parallel or nearly parallel to the coastline

Recurved spit near Mackay,
Queensland, Australia.



and exposed at low tide on a sandy shore. Also known as low and ball.

Rip channel – A channel cut by the seaward flow of a Rip Current (*q.v.*) across the nearshore zone, usually through nearshore bars.

Rip current – A strong narrow current (up to two knots) flowing seaward through breakers at right angles or an oblique angle to the coastline.

Ripple marks – Undulations (ridges generally less than 10 cm high, and troughs a few centimeters long between successive crests, often exposed at low tide on the foreshore), formed on a sandy foreshore and nearshore area by waves and/or currents. Some are parallel, others form intersecting, sometimes rhomboidal patterns. They may be symmetrical in cross-section, but are usually asymmetrical with steeper slopes in the direction of wave or current flow and elongated at right angles to this direction.

Rocky shore – An irregular, rugged rocky area between high and low tide where shore platforms have failed to develop, or have been intricately dissected.

Round hole – An enlarged blowhole on coastal slopes above cliffs that have been penetrated by caves, notably in southwest England.

Sabkha – See Sebkha.

Salt marsh – A flat or gently sloping vegetated wetland in the upper intertidal zone on sheltered parts of the coast (estuaries, inlets, lagoon shores). Often in the form of a depositional terrace, periodically submerged, with halophytic grasses, herbs, and shrubs; dissected by tidal creeks, and may contain enclosed Salt Pans (*q.v.*).

Salt pan – An enclosed bare depression within a salt marsh, apt to dry out, leaving an algal or saline crust. The term is also used for shallow artificial basins in which seawater is trapped and concentrated by evaporation to brine, which crystallizes into salt for harvesting.



Ria (Aber Benoit) on the Brittany coast, France.

Ridge and runnel on the shore at Saltburn, Yorkshire.



Salt weathering – Disintegration or decomposition of a rock surface by stress caused by the growth of salt crystallizing from sea spray, resulting in the formation of pits, cavities, or shallow basins.

Sand – A sediment consisting of rock particles with a diameter between 0.125 and 2.0 mm. Subdivisions are very coarse sand (1–2 mm), coarse sand (1/2 or 0.5–1 mm), medium sand (1/4 or 0.25–1/2 or 0.5 mm), fine sand (1/8 or 0.125–1/4 or 0.25 mm) and very fine sand (1/16 or 0.0625–1/8 or 0.125 mm). See Granulometry.

Sandflat – A relatively level unvegetated sandy intertidal area.

Seagrass – A marine grass that grows in the intertidal and shallow subtidal zones.

Sea level – The level at which the sea stands against the coast, conventionally taken as mean sea level, the arithmetic mean of the calm sea surface (excluding waves and oscillations related to winds and atmospheric pressure variations) measured at hourly intervals over at least 18.6 years.

Sebkha (Sbkha) – A low-lying very gently sloping saline area above normal high tide level on an arid coast, subject to occasional flooding by the sea or rain water, and prolonged phases of evaporation and dessication, resulting in hypersaline conditions. Some sebkhas have the branching form of rias. Some carry Salt Marshes (*q.v.*).

Sediment budget – The relative proportions of inputs, outputs and storage in a coastal sediment system (e.g., a Beach, *q.v.*).

Segmentation – The division of an elongated lagoon or bay into a chain of smaller, oval, or circular lagoons by the growth and coalescence of cusped spits and forelands and the erosion of intervening embayments, usually with narrow connecting channels (Price 1947; Zenkovich 1959).

Seiche – Alternating, diminishing fluctuations of water level after a rapid change in atmospheric pressure or where strong wind action has built up water



Rip currents (arrowed) on Woolamai Beach, Phillip Island, Victoria, Australia.



Round Hole at The Lion's Den, Lizard Point, Cornwall, England.

level downwind and lowered it upwind in a landlocked bay, inlet, or coastal lagoon.

Shale – A fine-grained stratified or laminated sedimentary rock which breaks readily into thin layers. Contrast Mudrock (*q.v.*).

Sharm – A long narrow marine inlet on an arid coast, usually at the mouth of a wadi (a generally dry valley in a coastal upland).

Shingle – A British term for Beach Gravel (*q.v.*), a coarse, loose deposit which may range from granules through pebbles to cobbles and small boulders, generally well-rounded particles that vary in shape from roughly spherical through ovoid to flat or platy Shingle beaches are sometimes termed coarse Clastic Beaches, Clastic (*q.v.*) meaning consisting of rock fragments.

Shore – The zone between the water's edge (Shoreline, *q.v.*) at high and low tide. Sometimes referred to as tidelands.

Shore drift – See Longshore drift, which is more accurate because shore drift could include movements landward or seaward across the inter-tidal zone.

Shoreline – The water's edge, moving to and fro as the tides rise and fall, so that there is a low tide shoreline, a mid-tide shoreline, and a high tide shoreline. The term has been used as a synonym for Coastline (*q.v.*), but it is useful to maintain a distinction between the two terms, taking Coastline as equivalent to the high tide shoreline. Where the tide range is large and the shore profile gently sloping there is much variation in the position of the various shorelines.

In the United States the term Shoreline (*q.v.*) is defined legally as mean high water (MHW), as shown on nautical charts produced by the National Oceanic and Atmospheric Administration (NOAA), while shorelines at other levels are called lines (e.g., the mean lower low water line) which is a private-property seaward boundary in some eastern

Sebkha on the coast of King Sound, Western Australia.



states. It should be noted that the American shoreline, thus defined, is submerged at high spring tides, and is not the margin of normally dry land.

Shore platform – A flat or gently sloping smooth or relatively smooth rock surface formed in the zone between high and low tide levels. The term is preferred to wave-cut platform, except on some soft rock outcrops where the platform has been literally cut by wave action alone. Some shore platforms are sub-horizontal, and may be *high tide shore platforms*, submerged only at the highest tides, or *low tide shore platforms*, exposed only at low tide. Others are seaward sloping intertidal platforms, extending down to below low tide level. Some are structural (coinciding with the upper surface of a seaward sloping hard rock formation), others erosional (cut across outcropping structures).

The term Beach Platform (*q.v.*) is not acceptable as an alternative to Shore Platform (*q.v.*) because of the risk of confusion with depositional terraces (e.g., Beach Berm, *q.v.*).

Shore pothole – A circular or oval depression scoured in a rocky shore platform by abrasion where sand or pebbles are circulated by wave motion (Wentworth 1944). Similar features may occur on a limestone coast where soil pipes formed beneath an ancient soil (palaeosol) are exhumed and excavated by wave scour.

Significant wave height – The average height of the highest one-third of waves measured over a 20-min observation period. As the number of waves within 20 min varies (240 5-s waves, but only 80 15-s waves) it may be preferable to measure the highest 33 of a set of 99 successive waves.

Silt – A sediment consisting of particles with a diameter between 1/256 (about 0.004) and 1/16 (0.0625) mm.

Skerry – A low, rugged rocky reef or scatter of reefs, generally intertidal but sometimes extending above high tide level, off a hard rock coast, particularly in Scandinavia and Scotland. Usually there are many skerries; the Norwegian term *skjergaard* has been wrongly translated as “skerry-guard” when in fact it means the area of calm water in the lee of skerries.

Slope-over-wall profile – A steep coast on which an upper, sloping facet descends to a steeper, often vertical, basal cliff. Such profiles are found on steep coasts where the rock formations dip seaward, on soft formations where a subaerial slope is recurrently undercut by marine erosion, and where a weak formation (such as glacial drift) forms a slope above a cliff in more resistant rock. On hard rock coasts that were periglaciated in Pleistocene times, as in southwest England, the upper slope is mantled by frost-shattered solifluction talus deposits (termed head) and the basal rocky cliff has been exposed by later marine erosion. Some slope-over-wall coasts have beveled slopes of uniform seaward gradient (Beveled Cliffs, *q.v.*), but many are convex (hogsback coasts) and some concave. Also known as two-storied cliff.

Solution notch – An elongated cliff-base hollow cut out near high tide level by solution (sometimes also bioerosion), often with an overhanging rock visor (protruding ledge of rock).



Shore potholes formed by abrasion, Pearce's Beach, Rye, Victoria, Australia.

Solution pan – A flat-floored basin developed on a limestone shore platform by solution, bordered by a microcliff or a constructional rim of algae.

Solution pipe – A cylindrical depression formed on a limestone surface by the dissolving of carbonates, often beneath soil and around a root system. There may be no surface depression, but soil has often subsided into a deepening hollow, the margins of which may become case-hardened by secondary precipitation of the leached carbonates. On Dune Calcarene (*q.v.*) coasts solution pipes (with encircling calcrete) may be exposed by wind erosion (as structures sometimes confused with petrified forests) or as shore platforms are cut, and the soil washed out to leave a protruding vase-shaped structure.

Sound – A narrow marine channel or strait between an island and the mainland, or between two islands, or between two land areas. A Strait (*q.v.*).

Spit – A finger-like ridge or embankment of beach material built up above high tide level and diverging from the land at one end (proximal) to terminate (distal end) usually in one or more recurves or hooks curving landward. See Recurved Spit.

Slope-over-wall profile
(periglacial slope), Dodman
Point, Cornwall, England.



Solution notch on a limestone
coast, Baron, Java, Indonesia.



Spring tide – An augmented tide range when the sun and the moon are in alignment with the earth, so that their gravitational effects are combined.

Squeaking sand – Beach or dune sand that emits sounds (squeaks, squawks, shrieks, sings, barks, booms, roars, or whistles) when walked upon, as the result of the mutual impact of sand grains. Sometimes called musical sand.

Stack – An isolated upstanding steep-sided rock pillar, column or pinnacle rising from the shore, a shore platform, or the seafloor close to a cliffed coast.

Stillstand – A condition of stability when the relative levels of land and sea remain the same for a prolonged period (several centuries), either because there has been no change in land or sea level, or because these levels have risen or fallen the same amount.

Storm surge – A temporary abnormal rise of sea level on a coast, as when an exceptionally high tide (often with sea level raised by low atmospheric pressure) is accompanied

by strong wave action generated by an onshore gale (cyclone, hurricane, typhoon).

Storm wave – A short, steep high wave (wave period generally <10 s) generated by strong winds.

Strait – A narrow passage of water between two land areas. A Sound (*q.v.*).

Strand – See Shore or Beach.

Strandflat – An extensive shallow submarine and low-lying emerged coastal platform (up to 65 km wide and typically with transverse gradients of up to 10° and many surmounting mounds and hillocks that were formerly stacks and islands), sharply contrasted with a high and rugged hinterland.

Strandline – See Shoreline.

Stromatolite – A low calcareous sedimentary domal structure, formed in shallow water where a mat or assemblage of blue-green algae have trapped sediment and precipitated calcium carbonate. See Algal Mat.

Stack at Sentinel Rock, Port Campbell, Victoria, Australia.



Structure, Structural features – The assemblage of rock formations (their arrangement, disposition, nature, and form) upon which erosional processes are, or have been, at work in shaping a landscape or coastline.

Subdelta lobe – Portion of a delta formed around the mouth of a distributary channel.

Submergence – A fall in the level of the land relative to the sea, achieved by actual subsidence of the land, a raising of sea level, or some combination of land and sea-level change that leaves the sea at a relatively higher level. A submerged coast (or feature) is one that stands at a lower level relative to the sea than when it originally formed; a submerging coast is one actually subsiding relative to sea level.

Swale – An elongated hollow or low-lying area between dune ridges, usually running parallel to the coastline.

Swamp encroachment – The advance of sediment-trapping swamp vegetation (reedswamp, salt marsh, mangroves) into shallow nearshore water, notably in sheltered bays, inlets and coastal lagoons, resulting in progradation of the land.

Swash – The rapid flow (uprush) of a breaking wave up the beach face.

Swash bar – A nearshore bar built parallel to the shore by wave swash, with a steeper landward slope advancing into a shallow lagoon.

Swash (Swash-backwash) cusps – See Beach Cusps.

Swash-dominated beach – A beach shaped primarily by the action of swash generated by breaking waves that arrive parallel to the shoreline, so that the beach has an alignment at right angles to the dominant direction of wave approach. Contrast with Drift-Aligned Beach (*q.v.*).

Swashway – A low sector of a barrier through which storm waves or surges flow. Sand or gravel swept through a swashway is deposited as a Washover Fan (*q.v.*) on the inner shore of the barrier.

Swatchway – A channel cut and maintained across a bar or shoal by wave and current scour. A term used in a novel by Erskine Childers (1903).

Swash zone – The zone regularly covered and uncovered as breaking waves generate swash and backwash.

Sweep zone – The zone (vertical plane) between high (often summer) and low (often winter) beach profiles, surveyed at right angles to the coastline, and indicating the extent of loss and gain resulting from cut and fill sequences.

Swell – Long, low waves (wave period typically 12–16 s) generated by distant storms and transmitted across oceans and seas.

Tafoni – Large cavities excavated in a rocky cliff by weathering, notably where the rock surface has been indurated and the underlying material is softer.

Tectonic – Crustal movements that produce geological structures, notably folding and faulting.

Terrigenous sediment – Sediment derived from the land surface and delivered to the coast by runoff, landslides, rivers, glaciers, or wind action, in contrast with marine sediment supplied to the coast from the seafloor.

Threshold – A shallow area near the seaward end of a drowned valley. In fiords it is often a rocky sill (sometimes partly depositional), but in estuaries and inlets on high wave energy coasts it is commonly a bank of inwashed sand, prograding landward.

Tidal bore – A large turbulent wave that develops a steep advancing water slope (up to 5 m high) that moves rapidly (5–8 m/s) into a long shallow funnel-shaped inlet or estuary during a rising spring tide on a macrotidal or megatidal coast. Known as the mascaret on the Gironde estuary; eagre on the Trent and Humber; pororocas on the Amazon.

Tidal current – A current generated by the rise or fall of the tide.

Tidal delta – An intertidal or subtidal bar or shoal, typically triangular or lobate, formed on the inner (flood delta) and often also the outer (ebb delta) sides of a tidal inlet or gap between barrier islands. Usually of sand, sometimes gravel, with diverging channel systems.

Swamp encroachment (reedswamp) on the shore of Lake Wellington, Victoria, Australia.



Tidal divide – An intertidal isthmus that emerges between an island and the mainland, or between neighboring islands, as the tide falls. There are usually tidal creeks heading away on either side and converging downstream. Sometimes called a tidal watershed, but this is inappropriate as the term watershed can also indicate a catchment area.

Tidal environment – A Macrotidal (*q.v.*) environment where mean spring tide range is between 4 and 6 m; a Mesotidal (*q.v.*) environment where mean spring tide range is between 2 and 4 m, and a Microtidal (*q.v.*) environment where mean spring tide range is less than 2 m. The term Megatidal (*q.v.*) is used where mean spring tide range is more than 6 m.

Tidal flat – A depositional feature, usually with a very gentle intertidal slope, consisting of sand (Sandflat, *q.v.*) or finer sediment (Mudflats, *q.v.*).

Tidal prism – The volume of seawater that flows in and out of an estuary or lagoon as the result of the rise and fall of the tide; a process known as tidal ventilation. It is computed as the difference between high and low tide volume, or as mean tide range times the mid-tide area of the basin.

Tide-dominated morphology – Landforms such as intertidal Mudflats (*q.v.*), salt marsh, and mangrove swamp, shaped primarily by tides and tidal currents, rather than by wave action. Most intertidal mudflats and sandflats, and most salt marshes and mangrove swamps, have to some extent been shaped by occasional wave action (notably producing shoreward drifting of sediment). It is suggested that tide-dominated shore morphology occurs where significant wave height is less than 20 cm at high tide.

Tide range – Mean tide range is the difference between mean high and mean low tides; neap tide range the difference between mean high neap and mean low neap



Tafoni on granite coast, Victor Harbour, South Australia.

Tombolo at Broulee Island, New South Wales, Australia.



Trailing spit, Halifax Island, Queensland, Australia.



tides; spring tide range the difference between mean high spring and mean low spring tides.

Tombolo – A Ridge or Barrier (*qq.v.*) of sand (occasionally Shingle, *q.v.*) built above high tide level in such a way as to link a former island to the mainland, or unite two islands. Sometimes this Italian term (which actually means the sand dune on top of the barrier) is taken to include the island as well as the depositional linking feature. Where the connecting deposit is submerged at high tide the terms tie-bar or tombolino are used. The term usually indicates a constructional feature, but similar forms can develop as a result of erosion of weak or unconsolidated material in the lee of a Headland or Breakwater (*qq.v.*).

Trailing spit – A spit that runs out from the lee shore of a headland or island. Also termed arrow spit or comet tail.

Transgressive barrier – A barrier that is migrating landward as a result of overwash and the movement of dune sand by onshore winds, encroaching on the lagoon or swamp that lies behind it.

Transgressive dune – A dune migrating away from the prevailing wind, some of which may assume barchan form, with lateral arms pushed leeward, as in desert dunes.

Trottoir – A constructional bench formed by the growth of algae at about mid-tide level on cliffed coasts (notably limestones), particularly in the Mediterranean. In some places the algae have colonized a preexisting rock ledge. Trottoirs have also been described from the Mariana Islands in the western Pacific. The mearls of Brittany are similar features, built by *Lithophyllum*.

Tsunami – Long seismic sea waves, generated by a major disturbance within an ocean basin (mainly due to earthquakes, but sometimes explosive volcanic eruptions or submarine landslides). They are of subdued form in deep water, but on entering shallow nearshore areas their height increases greatly, and can exceed 30 m when they break over the coastline.

Undertow – Laminar basal flow of water seaward beneath incoming waves; a continuation of Backwash (*q.v.*) from the shore into the nearshore and offshore areas. However,

Wave quarrying has disintegrated the shore platform at Tessellated Pavement, Tasmania.



seaward flow of water generally takes the form of segregated Rip currents (*q.v.*) rather than laminar basal flow.

Washover fan (Overwash fan) – A fan of beach material washed over low sectors of a coastal barrier by Storm waves (Storm surges) (*qq.v.*) to form a delta-like projection into a lagoon or on to backing marshland. See Overwash. In Texas the term washaround has been used for features produced by overwash on either side of a residual island or mound.

Wave diffraction – Increased curvature in the pattern of waves moving through a natural or artificial entrance to a bay or lagoon because of frictional retardation by bordering Headlands or Breakwaters (*qq.v.*).

Wave energy environments – High wave energy coasts are those exposed to strong ocean swell and large storm waves, with significant wave (Breaker, *q.v.*) heights of more than 1 m. Moderate wave energy coasts are where significant wave height is 0.3–1 m, and low wave energy coasts (where wave energy is limited by headlands, islands or reefs, or where there is extensive shallow water offshore), have significant wave heights of less than 0.3 m. The term zero wave energy is applied where wave action is completely absent.

Wave quarrying – Excavation or displacement of rock masses by wave impact.

Wave reflection – Where waves breaking against a rock formation, steep beach or artificial structure return seaward.

Wave refraction – Changes in the pattern of waves moving toward a coast because of frictional retardation by the shallowing seafloor and bordering headlands; waves in deep water have parallel crests, but as they move into a gulf or bay these become curved, and may anticipate and fit the outline of a beach or a cliffed coastline cut in soft rock formations.

Wave sluicing – The washing away by waves of material disintegrated from the surface of a shore platform by Weathering (*q.v.*) or erosion.

Waves – Undulations produced on the sea surface by disturbance, generally the frictional drag of wind action,



Wave refraction from seawall, Black Rock, Victoria, Australia.

but see Giant Waves (*q.v.*). Wave height is the vertical distance between adjacent crests and troughs, wave length the horizontal distance between successive wave crests, wave period the time taken by successive crests to pass a

Wave refraction in Seven Mile Bay, Tasmania.



fixed point, wave steepness the ratio of wave height to wave length, and wave velocity the speed at which a wave crest moves forward. Wave energy is taken as length multiplied by the square root of the height.

Weathering – The *in situ* disintegration or decomposition of rock surfaces exposed to the atmosphere and the upper part of an outcropping rock formation as the result of physical, chemical, and biological processes. It is usually indicated by changes in color, texture, composition, coherence, or firmness. Excludes erosion and induration. Residual rocks more resistant to erosion may produce strange forms (columns, pinnacles, pillow or cannon-ball structures, pillars and pedestals, sometimes termed hoodoo rocks).

Wetland – A general term describing swamps, bogs, marshes, and shallow (up to 5 m) lagoons and lakes. In coastal environments these include salt marshes, mangrove swamps, reedswamp, rush swamp, and seagrass beds.

Wind resultant – An expression of the long-term effect of winds of varying strength, duration, and direction, usually developed as a vector diagram from which a directional resultant may be determined.

Zetaform beach – An asymmetrical beach between headlands, where the outline is shaped by waves arriving obliquely, and refracted round the headlands. Also known as a Log-spiral (*q.v.*) or half-heart beach, and sometimes as a Headland (*q.v.*) bay beach (but this term does not indicate that the beach outline is asymmetrical).

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